Doctoral Thesis

Structural, geomechanical and petrophysical properties of shear zones in the eastern Aar massif, Switzerland

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Structural, Geomechanical and Petrophysical Properties of Shear Zones in the Eastern Aar Massif, Switzerland

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Summary

Rock zones containing a high fracture density and/or soft, cohesionless material are highly problematic when encountered during tunnel excavation. For example in the eastern Aar massif of central Switzerland, experiences during the construction of the Gotthard highway tunnel showed that shear zones were responsible for overbreaks in the form of chimneys several meters in height. To understand and estimate the impact of the shear zones on rock mass behaviour and to be able to predict where they may be encountered, knowledge concerning their spatial distribution, structural composition, rock mass strength and deformation characteristics is fundamental.

Mapping at surface and in existing tunnels revealed that the shear zones in the eastern Aar massif generally strike ENE-WSW and NW-SE and dip steeply towards south or north.

The detailed examination of surface and underground outcrops and of two newly drilled boreholes has shown that these structures are primarily ductile but also contain a considerably larger number of fractures than the intact host rock. The fractures oriented parallel and oblique to the foliation are mostly filled by a fine-grained, structureless arrangement of mainly muscovite, biotite and quartz. Higher and highest fracture densities are found in the fine-grained and strongly foliated sections of the shear zone. Along with fracture density, the thickness of the fracture infill increases towards areas with highest fracture densities.

Based on fracture densities and mica contents, it is proposed that the shear zones in the eastern Aar massif may be subdivided into 3 different textural subzones. With increasing fracture density and mica content, these are: (1) a strongly foliated zone, (2) a fractured zone and (3) a heavily fractured, cohesionless zone.

It is argued that the fractures were formed through hydrofracture mechanisms. The fractures are clearly superimposed on the ductile deformation structures and seem to have developed after the Alpine generated shear zones.

A series of laboratory triaxial tests, performed on samples taken from granite and gneiss hosted shear zones revealed that with increasing degree of tectonic overprint, sample strength decreases and rock behaviour shows a transition from brittle to ductile deformation. These trends may be explained by increasing fracture densities, increasing foliation intensity, increasing thickness of fine-grained, cohesionless fracture infill, and increasing mica content associated with the increasing degree of tectonic overprint. As
fracture density increases and the influence of discrete, persistent discontinuities on rock mass strength decreases, behaviour of the test samples becomes more and more representative of rock mass behaviour, i.e. that of a densely fractured continuum.

It may therefore be possible to assign each of the proposed shear zone subzones a typical rock mass constitutive behaviour. Under low confining stress, the host rock and the strongly foliated zone behave in a brittle fashion, the fractured zone exhibits brittle-ductile behaviour and the heavily fractured, cohesionless zone deforms in a ductile manner.

It can be assumed that reasons for the problematic nature of these shear zones during tunnel excavation derive from their special structural characteristics and rock mass constitutive behaviour. The design of adequate support measures along such shear zone structures would therefore benefit by considering and accounting for these factors.
Zusammenfassung


Mit Hilfe von Kartierungen im Feld und in bestehenden Tunnels konnte festgestellt werden, dass die Scherzonen im östlichen Aarmassiv hauptsächlich ENE-WSW und NW-SE streichen und steil nach S oder N einfallen.


Es wird vorgeschlagen, dass die Scherzonen aufgrund unterschiedlicher Bruchdichten und Glimmergehalte in 3 verschiedene Subzonen unterteilt werden können. Mit zunehmender Bruchdichte und Glimmergehalt sind dies: (1) eine "strongly foliated zone", (2) eine "fractured zone" und (3) eine "heavily fractured, cohesionless zone".

Die Brüche wurden vermutlich durch "hydrofracturing" erzeugt. Die Brüche überprägen die duktilen Deformationsstrukturen deutlich und scheinen sich nach der Alpinen Anlage der Scherzonen gebildet zu haben.

Eine Serie von im Labor an granitischen und gneisigen Scherzonenproben durchgeführten Triaxialtests hat ergeben, dass die Festigkeit der Proben mit
zunehmendem Grad tektonischer Überprägung abnimmt. Ihr Deformationsverhalten wechselt gleichzeitig von spröd nach duktil. Die Änderungen im Verhalten der Proben scheinen auf die mit dem Grad der tektonischen Überprägung zunehmende Bruchdichte, zunehmende Intensität der Schieferung, zunehmende Mächtigkeit von feinkörnigen, kohäsionslosen Bruchfüllungen und zunehmenden Glimmergehalt zurückzuführen zu sein. Mit Zunahme der Bruchdichte und gleichzeitiger Abnahme des Einflusses einzelner, persistenter Klüfte wird ausserdem das Verhalten der getesteten Proben mehr und mehr repräsentativ für das Gebirgsverhalten, d.h. für das Verhalten eines intensiv zerbrochenen Kontinuums.

Es sollte deshalb möglich sein jeder der vorgeschlagenen Scherzonen-Subzonen ein typisches Gebirgsverhalten zuzuordnen. Unter der Bedingung von niedrigen Umlagerungsdrücken sollte das undeformierte Nebengestein und die "strongly foliated zone" der Scherzone sprödes, die "fractured zone" spröd-duktils und die "heavily fractured, cohesionless zone" duktiles Deformationsverhalten zeigen.

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1. Introduction

1.1. General Objectives

Because of their influence on rock stability and fluid flow, rock zones containing a high fracture density and/or soft, cohesionless material are of practical importance in the applied geology. For example, tunnelling experience in such zones has shown that the fractured and/or soft, cohesionless nature of the rock mass may result in tunnel face instability, excessive overbreak or large deformations of the tunnel profile due to squeezing or swelling rock (e.g. Deere, 1973, Müller 1978, Keller & Schneider 1982, Schubert 1993, Schubert & Riedmüller 1997, 2000). Extreme water inflow sometimes associated with flowing ground frequently aggravates such instabilities. In the eastern Aar massif of central Switzerland, where sections of the new Gotthard Base Tunnel will be constructed (Fig. 1.1), encounters during the construction of the Gotthard highway tunnel have shown that heavily fractured areas within shear zones were responsible for overbreaks in form of chimneys several meters in height (Keller et al. 1987).

To prevent catastrophic failures and to enable a more rigorous planning of the tunnel design including the determination of suitable support requirements and safety regulations within the areas of the shear zones, it is essential for construction works, i.e. the tunnelling activities currently undertaken in the eastern Aar massif, to predict where such shear zones, may be encountered and to estimate their impact on site characterization of rock mass behaviour. The spatial distribution of the shear zones is generally established by mapping on surface and in existing underground openings. Immediate detection of shear zones during the tunnelling process may be attained by geophysical logging in investigative boreholes ahead of the tunnel face (e.g. Militzer et al. 1986, Brückl & Stötzer 1995). The estimation of the impact of the shear zones on site characterization of rock mass behaviour is commonly obtained from the performance of analytical and numerical calculations. Geophysical logging in investigative boreholes ahead of the tunnel face may also give some information about the geotechnical behaviour of the rock mass (e.g. Militzer et al. 1986). To properly
incorporate the effect of the shear zones within existing analytical and numerical design methodologies, knowledge concerning rock mass strength and deformation characteristics is fundamental. Previous studies examining the geomechanical properties of heavily fractured rock masses, especially fault zones, have shown that rock strength and deformation parameters vary depending on the different structural characteristics within the zones, e.g. fracture densities (e.g. Chester & Logan 1986, Habimana et al. 1998). In general, it is assumed that with an increasing degree of natural fragmentation there is a reduction in rock strength and Young’s modulus and an increase in relative ductility of rock behaviour.

In the eastern Aar massif, the presence of shear zones is common knowledge. Previous studies mainly focused on the spatial distribution of the zones and paid attention to the possibility and the mechanisms of recent movements along these zones (e.g. Jäckli 1951, Eckardt 1957, Steck 1968b, Eckardt et al. 1983). Numerous underground excavations constructed in the eastern Aar massif over the past 150 years resulted in additional knowledge about the spatial distribution of the shear zones (e.g. Schindler 1972, Keller et al. 1987, Schneider 1993). These reports also documented the difficulties the shear zones caused during excavation. Investigations into the structural and geotechnical, i.e. hydrogeological, characteristics of shear zones were only performed for zones in the central part of the Aar massif (e.g. Meyer et al. 1989, Bossart & Mazurek 1991). These studies, which focused on the migration of fluids in the Grimsel rock laboratory, provided detailed maps at a μm- to m-scale and information about the nature of the rock material in several smaller shear zones. The findings could be extrapolated to shear zones within an area of about 1 km³ around the rock laboratory. The authors described the shear zones as ductile shear zones which were reactivated by brittle deformation. Given the relatively high density of shear zones in the eastern Aar massif and the occurrence of ground control problems such as the overbreaks experienced in the Gotthard highway tunnel whenever a highly fractured zone was encountered, it is surprising that the structural and geomechanical properties of these shear zones and their variance within the zones have never been studied rigorously. Also the petrophysical properties which are also assumed to depend on the structural composition of the shear zone and which are important for the applicability of geophysical borehole logging methods were never examined.

This thesis is part of a research project which has been initiated to investigate and better understand the structural, geomechanical, hydrogeological and petrophysical characteristics of the shear zones in the eastern Aar- and Gotthard massifs. The present work was directed towards the improvement of predictions with respect to which impacts the shear zones in the eastern Aar massif have on site characterization of geomechanical rock mass behaviour. Apart from this, the thesis should contribute to an improved general characterization of the relationship in between structural, geomechanical and petrophysical properties of shear zones in crystalline rocks. The thesis was based on surface, underground and borehole observations and involved the performance and evaluation of a series of laboratory triaxial tests on core sections and of geophysical borehole logs. The thesis is composed of two papers. The first paper presents results on the spatial distribution and the structural properties of the shear
zones in the eastern Aar massif. In the second paper, the results of the triaxial compression tests focussing on the geomechanical properties of the shear zone rocks are discussed. The results are assessed with respect to the structural composition of the shear zones. The outcome of geophysical borehole logging across shear zones is presented in the Appendix.
Fig. 1.1. Geographical and geological setting of the Aar massif (after Abrecht 1994), location of the study area and location of Gotthard Base Tunnel and Gotthard highway tunnel.
1.2. Geological and Geomechanical Characterization of Shear Zones

1.2.1. Shear Zones in Geology

Faults are defined as planar discontinuities along which a significant displacement (> 0.5 mm and generally much larger than this minimum value) has taken place (Price 1966, Suppe 1985, Ramsay & Huber 1987). When this displacement is not confined to a single surface, a narrow zone of high strain, called shear zone, is formed.

The presence of shear zones indicates that within a given deforming rock mass, the distribution of strain was heterogeneous. Spatial strain gradients, which are characteristic for shear zones, are also an expression of the heterogeneous strain distribution. The amount of strain is generally highest within the center of a shear zone, decreasing outward into the wall rocks adjacent to the zone (e.g. Davis & Reynolds 1996). Another characteristic feature of shear zones is their curvilinear geometry, encompassing and wrapping around more rigid, less deformed rock bodies (shear lenses).

Four general types of shear zones can be subdivided, based on their dominant type of deformation (e.g. Ramsay & Huber 1987, Davis & Reynolds 1996). A brittle shear zone, generally called a fault zone, contains fractures and other features formed by brittle deformation mechanisms (The term fracture is used in this text according to Griggs & Handin (1960), as a discreate break along which a loss of cohesion has taken place.). A ductile shear zone displays structures that have metamorphic aspect and record shearing by ductile flow. Semibrittle shear zones include en échelon veins or joints and stylolites and involved mechanisms such as cataclastic flow and pressure solution. Brittle-ductile shear zones show evidence for both brittle and ductile deformation. They are for example formed where conditions during shearing either are intermediate between brittle and ductile or change from ductile to brittle or from brittle to ductile. Brittle-ductile shear zones are also often produced by the reactivation of ductile shear zones under brittle conditions or vice versa.

1.2.1.1. Fault Zones

Brittle Deformation:

Deformation is called brittle, if it involves only a very small amount of crystal plasticity prior to transgranular failure on a grain scale and is dominated by fracturing (e.g. Logan 1979, Groshong 1988, Passchier & Trouw 1996). Fault zones are therefore thought to be formed at low temperature and moderate confining pressure, i.e. at shallow depths in the crust, generally within the upper 5 km of the earth’s surface. Besides this, brittle deformation is favored by relatively high strain rates and high fluid pressures that occur during seismic events (Sibson 1977, Wise et al. 1984, Davis & Reynolds 1996).
Features of Fault Zones:

In fault zones the displacement is taken up on a network of closely spaced faults and joints. Fault zones are therefore often marked by striation lineation and brecciated rocks, i.e. cataclastic fault rocks. As the high permeability of the fractured material promotes hydrothermal inflow, faulting is also often accompanied by hydrothermal activity (e.g. Hulin 1929, Higgins 1971, Sibson 1977, Wise et al. 1984).

Studies into the structural composition of fault zones often describe a characteristic internal zonation of the fault zones consisting mainly of a central zone surrounded by a damaged zone (Chester & Logan 1986, Wallace & Morris 1986, Chester et al. 1993, Braathen & Gabrielsen 1998). The central zone characteristically contains a dense network of fractures and cataclastic fault rocks. The damaged zone is represented by fractured host rock.

Cataclastic Fault Rocks:

Cataclastic fault rocks are dominantly produced by fragmentation of rocks and mineral grains, leading to rigid-body rotation, frictional grain boundary sliding and dilatancy among grain fragments, grains or groups of grains, i.e. cataclasis (e.g. Higgins 1971, Sibson 1977, Schmid & Handy 1991). As brittle deformation can also involve a very small amount of crystal plasticity (Groshong 1988), cataclastic fault rocks can also show a limited amount of ductile deformation features such as undulose extinction, deformation twins, deformation bands etc.. Hydrothermal activity is the cause for neomineralisation and alteration in the area of fault zones. The presence of fluids also enables pressure solution to be active.

Due to the intense transgranular and intragranular microfracturing, cataclastic fault rocks are often composed of an unsorted mixture of angular and rounded clasts and a finer-grained matrix (e.g. Passchier & Trouw 1996). However, subsequent rigid-body rotation, grain-boundary sliding, the alignment of mica or elongate porphyroclasts and possibly a weak preferred orientation may induce a crude foliation (Passchier & Trouw 1996).

Cataclastic fault rocks can be classified into cohesionless cataclastic fault rocks (fault breccias and fault gouges) and fault rocks that are cohesive (cataclasites) (e.g. Sibson 1977, Schmid & Handy 1991). The adjectives cohesive and cohesionless describe the inferred cohesiveness of the fault rocks during faulting and not the cohesive state of the fault rocks when exposed in the outcrop. Fault breccias and fault gouges can regain cohesion by syn- or post-tectonic cementation associated with infiltrating hydrothermal fluids (Ramsay & Huber 1987, Schmid & Handy 1991). They are then called cemented breccia or cemented gouge. Fault breccia and fault gouge are differentiated by the amount of matrix material contained in the rock and by the rock fabric. In a fault breccia the amount of matrix constitutes less than 30 % of the rock mass. The fabric of a fault breccia is ususally random. In a fault gouge the matrix makes
up more than 30 % of the rock mass. The matrix usually is dark brown and is composed
of clay minerals and/or extremely fine-grained (< 10μm) remnants of parent rock
(feldspar, quartz, calcite) (Tanaka 1992, Cladouhos 1999). Fault gouge may show a
crude foliation.

1.2.1.2. Ductile Shear Zones

Ductile Deformation:

Ductile behaviour means deformation without development of fractures on the
grain scale (Passchier & Trouw 1996). Deformation is therefore called ductile, if it is
mainly achieved by crystal plasticity and may comprise only a minor amount of
fracturing. Ductile deformation mechanisms become dominant when temperatures and
confining pressures are high, i.e. at deep levels in the crust, and strain rate and fluid
pressure are relatively low (Passchier & Trouw 1996, Davis & Reynolds 1996). Most
ductile shear zones form under metamorphic conditions. However, the threshold from
brittle to ductile depends on the rock type: For example rock units composed of halite
and gypsum will deform ductilely under conditions in which quartz and feldspar are
brittle.

Features of Ductile Shear Zones:

Ductile shear zones characteristically show a gradual decrease in strain away from
the shear zone center (Ramsay & Huber 1987, Davis & Reynolds 1996). The rocks
formed in ductile shear zones are called mylonites (e.g. Bell & Etheridge 1973,
Passchier & Trouw 1996).

Mylonites:

A mylonite is a rock that shows evidence for strong ductile deformation
(Passchier & Trouw 1996). Brittle deformation may play a minor role in isolated
inclusions or clasts. It mainly occurs in strong minerals such as feldspar and weak
minerals like mica. Characteristic fabric elements of a mylonite are its small grain size
and strongly developed, extremely regular planar foliation and straight stretching
lineation. The small grain size is mainly due to intracrystalline deformation and
recrystallisation of the grains of the host rock. Apart from this, small grains may also
have formed by the crystallisation of new minerals (e.g. mica at the expenses of
feldspar). The recrystallised or newly crystallized minerals (quartz, calcite and feldspar)
commonly show evidence of lattice preferred orientation. Mylonite also often contains
remnants of resistant minerals of a larger size than the grains in the matrix
(porphyroclasts). Besides these porphyroclasts, low-strain lenses (shear lenses) around
which domains of high strain anastomose occur in most mylonites. The shear lenses can
be found on all scales.

The fabric of mylonite is strongly dependent on the lithotype and original
structure of the rock in which it develops (Passchier & Trouw 1996). Nevertheless, a
general strong fabric exists for all rock types with increasing metamorphic grade (except high temperature metamorphosis like the granulite facies and in migmatites) and strain rate. The gradual decrease in strain away from the shear zone center characteristically for ductile shear zones is therefore expressed by a decrease in intensity of the mylonitic fabric.

After Passchier and Trouw (1996), mylonites can be classified according to the metamorphic grade (e.g., high-grade mylonite) or by the lithotype or mineralogy in which they are developed (e.g., quartzite-mylonite). Another commonly used classification is based on the percentage of matrix as compared to porphyroclasts. Rocks with 10 to 50 % matrix are classified as protomylonites, rocks with 50 to 90 % matrix are mylonites or mesomylonites, and rocks with more than 90 % matrix are referred to as ultramylonites (e.g., Sibson 1977). The term blastomylonite is used for a mylonite with significant static recrystallisation and the term phyllonite is used for a fine-grained, mica-rich mylonite.

1.2.2. Shear Zones in Rock Mechanics

Regarding strength and deformation characteristics, shear zones showing an increased fracture density are zones of decreased rock strength and increased ductility of rock behaviour (e.g., Zoback & Byerlee 1976, Scott 1994). Parameters vary within the shear zone, depending on the structural properties (Chester & Logan 1996, Habimana et al. 1998). Shear zones showing an increased fracture density and the corresponding shear zone material may be represented by a number of rock mass models designed to estimate geomechanical properties on basis of structural properties.

1.2.2.1. Bimrocks and Bimsoils

One rock mass model which could be useful for the geomechanical characterization of shear zone material are the complex mixtures of soil-like matrices and strong blocks of rock lacking spatial, lithologic and mechanical continuity Medley (1999) grouped into the terms "bimrocks" and "bimsoils" (block-in-matrix rocks or soils). For example fault breccias, fault gouge or a system of fault gouges anastomosing around lenses of intact rock may be represented by bimrocks or bimsoils. Bimrocks are exactly defined as a mixture of rocks composed of geotechnically significant blocks within a bonded matrix of finer texture. Unbonded or uncemented soils containing rocks are termed bimsoils. The expression "geotechnically significant blocks" means that the blocks have (1) mechanical contrast with the matrix, (2) the range in block sizes is between 5 and 75 % of the characteristic engineering dimension (the scale of engineering interest) which describes the problem at hand, such as the diameter of a tunnel or triaxial specimen and (3) the block volumetric proportion (the total volume of blocks divided by the total volume of the bimrock mass) is between 25 and 75 %. The definition of bimrocks and bimsoils ignores rock petrology, genesis and other geological connotations associated with rock names and fabrics.
Strength and deformation properties of bimrocks and bimsoils are simply and directly related to their block volumetric proportions (Lindquist 1994, Lindquist & Goodman 1994, Medley 1999): Below 25 % block volumetric proportion, the strength and deformation properties of the bimrock/bimsoil will be that of the matrix. Between about 25 and 75 %, the friction angle and modulus of deformation of the bimrock/bimsoil mass will proportionally increase and the cohesion decreases due to the presence of blocks. Beyond 75 % block proportion, the blocks tend to touch, the rock mass is no longer matrix-supported and there is no further increase in bimrock/bimsoil strength. The overall strength of bimrocks/bimsoils is independent of the strengths of the blocks. The weakest component of a bimrock/bimsoil is the contact between blocks and matrix.

1.2.2.2. Jointed Rock Masses

Other examples for rock mass models useful for the description of shear zones showing an increased fracture density are contained in the table of jointed rock masses Hoek et al. (1998) and Hoek (1999) composed for estimating the Geological Strength Index (GSI). Hoek et al. (1998) and Hoek (1999) differentiated six classes of rock masses depending on their fracture density. These classes are:

- **intact or massive** (intact rock specimens or massive in situ rock masses with very few widely spaced discontinuities)
- **blocky** (very well interlocked, undisturbed rock mass consisting of cubical blocks formed by three orthogonal discontinuity sets)
- **very blocky** (interlocked, partially disturbed rock mass with multifaceted angular blocks formed by four or more discontinuity sets)
- **blocky/disturbed** (folded and/or faulted with angular blocks formed by many intersecting discontinuity sets)
- **disintegrated** (poorly interlocked, heavily broken rock mass with a mixture of angular and rounded rock pieces) (including fault breccia)
- **foliated/laminated/sheared** (Thinly laminated or foliated and tectonically sheared weak rocks. Closely spaced schistosity prevails over other discontinuity sets, resulting in complete lack of blockiness) (including fault gouge).

Each of these classes can be further differentiated, depending on the surface conditions of the fractures, into:

- **very good** (very rough, fresh, unweathered surfaces)
- **good** (rough, slightly weathered, iron stained surfaces)
- **fair** (smooth, moderately weathered or altered surfaces)
- **poor** (slickensided, highly weathered surfaces with compact coatings or fillings of angular fragments)
- **very poor** (slickensided, highly weathered surfaces with soft clay coatings or fillings).
Hoek & Brown (1997) assumed that the strength of a jointed rock mass depends on the properties of the intact rock pieces and also upon the freedom of these pieces to slide and rotate under different stress conditions. This freedom is controlled by the geometrical shape of the intact rock pieces as well as the condition of the surfaces separating these pieces. Angular rock pieces with clean, rough discontinuity surfaces will result in a much stronger rock mass than one which contains rounded particles surrounded by weathered and altered material. According to these reasons, the class "foliated/laminated/sheared, very poor" will show least rock strength.
1.3. Investigation Methods

1.3.1. Field Work

To investigate the structural, geomechanical and petrophysical properties of the shear zones in the eastern Aar massif, several different investigation methods were used. Air photo analysis and a new evaluation of existing field and tunnelling reports was done for a first localization of the shear zones. Field mapping at surface and in existing tunnels (security tunnel of the Gotthard highway tunnel, tunnels Voralpreuss and Furkareuss of the hydroelectric power station Göschenen), which was done in the summers of 1997, 1998 and 1999, had the purpose to confirm the lineaments detected on air photos and the zones described in existing reports as shear zones, to map additional shear zones and to establish their spatial distribution at a scale of 1:25 000. A full record of the shear zones mapped is stored on the CD attached in the Appendix. A lineament was determined as an outcrop of a shear zone, when mylonites could be detected or when a high fracture density combined with cohesionless areas or striation lineation occurred. However, striation lineation within shear zones was never observed. Apart from mapping of the spatial distribution of the shear zones, field work had the purpose to examine in detail, i.e. on a cm- to m-scale, the structural composition of the zones. To get a broad overview about the structural composition of the shear zones, a large number of shear zones developed in several different kinds of crystalline rocks of the eastern Aar massif and shear zones having different orientations were examined.

1.3.2. Drilling Campaign

As structures and rock material at surface are usually heavily weathered and therefore are not necessarily representative for structures and materials at depth and former outcrops of larger shear zones in underground constructions generally are hidden by a concrete or steel coverage, five cored boreholes were drilled from the underground at two different drill sites. Apart from delivering cores for detailed mapping with respect to structures and nature of the rock material, the boreholes had the purpose to provide undisturbed samples for thin section analysis and laboratory testing. Additionally, the boreholes should be sites for geophysical logging, i.e. the examination of the petrophysical properties of the shear zones.

The two different drill sites were located in the safety tunnel of the Gotthard highway tunnel (Fig. 1.2). The Gotthard highway tunnel is partly situated in the eastern Aar massif. The drill sites both had an overburden of 400 to 500 m. One drill site was placed in a shear zone in Variscan Aar granite (boreholes Ia and Ib), the second one was positioned in a shear zone in the Southern Gneiss Zone of the pre-Variscan basement (boreholes Ia, I Ib1 and I Ib2). The shear zones both had an ENE-WSW strike and dipped steeply to the south. Both shear zones had been sites of overbreaks during construction of the safety tunnel.

At each drill site, one borehole was oriented almost normal to the strike and dip of the shear zone (boreholes Ia and I Ib). The other boreholes (boreholes Ib, I Ib1 and I Ib2)
were oriented parallel to the strike and perpendicular to the dip and were drilled into the center of the shear zone (Fig. 1.3). To enable the boreholes to be normal to the dip of the shear zones, the boreholes had to be near horizontal. The boreholes were drilled with a 3 to 5° dip towards south (boreholes Ia and IIa) or east (boreholes Ib, IIb1 and IIb2), because most of the geophysical borehole logging tools which were intended to be used required water-filled boreholes to be able to operate. Maximum length of the boreholes was 15 m. The boreholes oriented normal to strike and dip had the purpose to deliver undisturbed samples for core logging, thin section analysis and laboratory tests and should provide sites for geophysical borehole logging. Only boreholes oriented in this direction can document the development of the structural properties across the shear zone. The boreholes oriented parallel to the strike and perpendicular to the dip of the shear zone were drilled to get more samples from the central part of the shear zone and to test if this material shows any anisotropy.

To minimize sample disturbance of the highly fractured shear zone material, triple tube drilling flushed with a minimal amount of water was performed. To further protect the rock material, the boreholes were drilled with a relatively large diameter of about 116 mm. The resulting cores had a diameter of 90 mm. The PVC lining holding the cored samples was immediately sealed after extraction.

The resulting cores were logged with respect to their structure and nature of the rock material. Results of this logging are stored on the CD attached in the Appendix. The logs also indicate the core sections taken for thin sections analysis and triaxial laboratory tests.

1.3.3. Thin Section Analysis

To examine shear zone structure in great detail, thin section analyses were performed on samples from different core sections.

To control sample disturbance, the cores were kept in the PVC liner while cutting about 10 to 20 cm long sections from them. The samples, still contained in the PVC liner, then were saturated with fluorescenting epoxy resin. Saturation was done under vacuum conditions, so that the resin was pressed into the sample when it was brought to atmospheric pressure. Fluorescenting resin was used, as this resin can be made visible under UV-light and then indicates open fractures and pore spaces. After hardening of the rock material, rock sections oriented normal to foliation and parallel to stretching lineation were cut from the samples. From these rock sections, thin sections were prepared.

Thin section analysis focused on the observation of the content of the main minerals, foliation intensity, intensity of recrystallization, nature of fractures (intra-, transgranular, shear fractures, extension fractures), orientation of fractures with respect to foliation and nature and thickness of the fracture infill. Apart from these features, when using a microscope operating with UV-light, also the density of fractures and pore spaces could be examined.
1.3.4. Geomechanical Laboratory Tests

In order to determine the geomechanical properties of the shear zones and to show the relation of structural and geomechanical properties, a series of laboratory triaxial tests was performed on selected core sections from the boreholes in the Gotthard safety tunnel. The objectives of the tests were to obtain stress-stain curves to be able to observe the rock behaviour with respect to the applied stress. The elastic constants Young's modulus (E) and Poisson's ratio (ν), the peak strength (σ_{max}) and the Mohr-Coulomb residual strength parameters residual cohesion (c_r) and residual internal angle of friction (ϕ_r) were also determined.

It was decided to perform triaxial tests, where σ_1 > σ_2 = σ_3, because of the high friability of some samples. To be able to test these samples, a confining pressure, which holds the sample together at the beginning of the test, was needed. To give the possibility to compare the testing results of different samples with each other, all samples were tested under the same conditions. Tests were performed at the rock mechanics laboratory (LMR) of the EPFL Lausanne.

1.3.4.1. Sample Preparation

Given the highly fragmented nature of some samples, sample disturbance had to be controlled by keeping the samples in the PVC liner during end preparation. Sample ends were cut through the plastic tubing, considering that they are flat and perpendicular to the core axis. After end preparation, samples were removed from the plastic tubing and supported with adhesive tape and a calcite-like paste (HILTI MD 2000) where necessary. With only the exception of samples GN5a-c, which were drilled to smaller diameters of 85 and 55 mm, the original 90 mm diameter of the cored samples was maintained. The diameter of samples GN5a-c had to be reduced to reduce time for sample consolidation. All diameters satisfied the ISRM standards of having a diameter which is about 10 times the maximum grain size or the fissure spacing (ISRM Commission 1983). The height of the specimen, according to ISRM recommendations, was approximately two times the diameter of the samples.

To fit the specimen into the triaxial cell, steel disks/end platens, which had the same diameter as the specimen, were put under and on top of it. The upper end platen was split and had spherical seats incorporated. The whole sample stack was jacketed with a rubber membrane. The jacketed specimen was additionally attached to the end platens by an adhesive tape, which was wrapped around the rubber membrane at the site where end platens and sample met (Fig. 1.4).

1.3.4.2. Test Setting and Load Path

The triaxial tests were performed as consolidated, drained compression tests (CD tests). Consolidated, drained tests were carried out to minimize the effect of excess pore pressures, as samples were tested with their natural water contents to prevent any
negative effects on the rock's structure associated with drying. Tests were performed according to ISRM recommendations (ISRM Commission 1983). The loading system consisted of a servo-controlled, stiff testing machine (Fig. 1.5), making it possible to follow the full stress-strain curve into the post-peak region. The maximum load capacity of the testing frame was 200 t. Both, LEGEP and HOEK type triaxial cells were used. Confining pressures were provided by means of hydraulic oil pumps.

Axial loads were applied at a constant displacement rate of 0.1 mm/min, except for samples GN5a-c where a loading rate of 0.2 mm/min was applied. The confining pressure for the tests was set at about 5 MPa, with the exception of specimen GN3b where a confining pressure of 1 MPa was used. Loading commenced in this fashion up to the peak strength after which a form of controlled multiple failure state loading (see Kováři & Tisa 1975) was employed to increase confinement through several residual stages of deformation and strength. In contrast to Kováři & Tisa's (1975) multiple failure state triaxial tests, increase in confining pressure to maintain different limit equilibrium states were interpreted as not being those for peak strength conditions but instead, residual strength conditions. Accordingly it is assumed that at peak strength a major change to rock fabric exists in the form of a failure plane and any subsequent conditions can only be applied to that of a residual strength state. Confining pressures were increased and held constant two times in succession from 5 to 10 MPa and 10 to 15 or 20 MPa until residual strengths at those intervals was reached (Fig. 1.6).

1.3.4.3. Parameters Recorded and Calculated

Parameters recorded during the tests included the axial load (\(\sigma_1\)), confining pressure (\(\sigma_2 = \sigma_3\)), axial sample displacement (\(\Delta h\)) and the volume change of the confining pressure fluid in the triaxial cell. The record of the movement of the loading piston (\(\Delta h\)) was used to calculate the axial strain:

\[ \Delta h/h = \varepsilon_{\text{axial}} \] (1).

Volumetric strain could be calculated from the volume changes of the confining pressure fluid:

\[ \Delta V/V = \varepsilon_{\text{volumetric}} \] (2).

Radial strain was determined assuming the relationship for axial symmetry of a cylindrical volume:

\[ \varepsilon_{\text{radial}} = 0.5 (\varepsilon_{\text{volumetric}} - \varepsilon_{\text{axial}}) \] (3).

From the measured and calculated data, stress-strain curves for each sample were plotted. Figures showing the stress-strain curves are contained in the Appendix.

Approximations of the Young's modulus (E) were calculated from the axial stress-strain curves as (1) a Tangent modulus (\(E_T\)), which is the slope of the axial stress-
strain curve at 50 % of the peak strength and (2) a Secant modulus ($E_s$), which is the slope of a straight line joining the origin of the axial stress-strain curve to the point on the stress-strain curve at 50 % of the peak strength. Values of Poisson’s ratio ($\nu$) were determined from stress-strain curves by dividing $E_T$ through the slope of the $\sigma_1$-$\varepsilon_1$ curve at 50 % of peak strength. Calculations were restricted to simple Young’s moduli and Poisson’s ratios of an isotropic material due to limitations in testing instrumentation, although a transverse anisotropic model would have been more appropriate for the strongly foliated nature of the shear zone rocks.

Through the testing procedure and load paths applied, a Coulomb residual strength envelope could be obtained for each sample tested. The residual cohesion ($c_r$) and residual internal friction angle ($\phi_r$) could be established from the residual strength envelope through the Coulomb criterion relationship:

$$\tau_r = c_r + \sigma_n \tan \phi_r$$  \hspace{1cm} (4)

where $\tau_r$ = residual shear strength along failure plane, $c_r$ = residual cohesion, $\sigma_n$ = normal stress on the plane of failure, $\phi_r$ = residual angle of internal friction. The parameters were calculated with the help of the program ROCKDATA (Shah & Hoek 1991a). Assuming a linear Coulomb envelope, the program was used to derive values of the residual cohesion and friction angle by means of a linear regression fitting routine. Shah & Hoek (1991b) note that the linear regression fit typically provides an excellent correlation to a uniform set of data points given the Coulomb failure envelope’s linear nature.

The Coulomb criterion implies that a continuous, smooth shear plane develops at peak strength and that the peak strength envelope is linear, which is a simplification at best for most rock materials. However, Brady & Brown (1993) suggest that the Coulomb criterion should provide a relatively good representation of the residual strength conditions. The residual strength, defining the load bearing characteristics of the rock subsequent to the full development of a failure plane, should generally increase with the presence of a rough failure plane or even multiple failure planes. The values determined on basis of the Coulomb criterion therefore should describe the lower limit of the residual strength.

1.3.5. Geophysical Borehole Logging

To characterize the shear zones in the eastern Aar massif by selected geophysical and geomechanical in situ-properties and to analyze the relation of these and the structural properties, caliper, full wave sonic and spectral gamma ray logs were performed in boreholes Ia and IIa. The tools used were slimhole logging probes manufactured by Robertson Geologging, Deganwy (UK).
1.3.5.1. Caliper Logging

Caliper logging provides a continuous measure of the diameter of a borehole as a function of depth. Information about the diameter or changes in the diameter is generally needed for the interpretation of logs, e.g. sonic logs, where the logged values are dependent on the borehole diameter. Apart from this function, a caliper log is able to indicate incompetent or fractured formations, as borehole overbreaks or a widening of the borehole due to erosion by the circulating borehole fluid especially occur within these formations.

Logging was performed with a ROBERTSON borehole-geometry tool. The caliper consisted of two independent pairs of linked arms, which is four arms oriented at right angles. With the two independent pairs of arms, the tool could provide continuous measurements of the borehole diameter in two directions being perpendicular to each other, i.e. in a X and a Y direction. For correlation with other wireline logs, a natural gamma detector was also included in the probe.

To enable the sonde to translate electronic signals into metric diameter values, a manual calibration of the probe using a calibration jig had to be made before each run. Logs were performed by uphole logging with a logging speed of 0.8 to 1.1 meters per minute.

1.3.5.2. Full Waveform Sonic Logging

Full waveform sonic logging gives the possibility to determine interval transit times/velocities of different wave types, e.g. compressional and shear waves \((v_p, v_s)\). It also enables the observation of wave frequencies and amplitudes needed for the calculation of specific attenuation \((Q)\). Velocities and attenuation are a measure of the rocks capacity to transmit elastic waves, which is largely dependent on the elastic properties and density of the material through which the sound waves travelled (Tixier et al. 1959). Variations in the velocity and specific attenuation of the waves may therefore be correlated with changes in the elastic properties and densities of the different rock sections penetrated by the borehole, i.e. with changes in the lithology, the porosity or fracture density. For example compressional and shear wave velocities and amplitudes generally decrease with increasing porosity and/or fracture density. Knowledge of the compressional and shear wave velocities, together with the density of the corresponding rock sections, additionally enables several dynamic elastic properties, e.g. the dynamic Young’s modulus, of the rock sections to be estimated (e.g. Hudson 1993b, Kelly & Mareš 1993). Correlation of the mechanical properties with the attenuation characteristics of compressional and shear waves is a possibility of estimating the degree of fracturing of the rock material (e.g. Cheng et al. 1982).

Sonde Construction and Measurement:

Data were collected with a ROBERTSON full waveform sonic tool consisting of a double array of sonic pulse transmitters and receivers. The transmitter-receiver arrays
were arranged inverted (Fig. 1.7). Receiver spacing was 40 cm. To prevent waves moving through the body of the sonde, polyurethane tube sections were mounted between transmitters and receivers.

Compressional waves were generated in the borehole fluid by bursts of high-frequency sound (approximately 23 kHz) emitted by the transmitters. For the formation and propagation of the compressional waves, it was therefore essential that the borehole was filled with water. In the case of "fast" rocks surrounding the borehole (shear velocity in rocks is greater than the borehole-fluid velocity), the compressional waves were refracted at the borehole wall and were partly transformed into shear, pseudo-Rayleigh and Stoneley waves (Fig. 1.8) (e.g. Astbury & Worthington 1986). Along the fluid/rock interface, all wave types were refracted back into the fluid as a compressional wave. In "slow" rocks, where shear wave velocity is less than the borehole-fluid velocity, refraction for a shear wave does not occur. In such a case, pseudo-Rayleigh waves would also be missing and the arrivals of these wave-types would therefore not be recorded (Astbury & Worthington 1986).

The compressional wave travelling along the fluid/rock interface is normally the first wave arriving at the receiver. This first arrival, i.e. the time required for the compressional wave to traverse the distance between transmitter and receiver, was recorded in the receivers as compressional wave transit time. As the sonic probe consisted of two transmitter-receiver pairs, every receiver noticed two signals of compressional wave transit time (Fig. 1.7). From these signals the sonde automatically calculated a compressional wave interval transit time (average transit time needed by the waves to travel the distance between the two receivers). The values are borehole compensated, which means that the effects of poor centralization, an uneven borehole diameter or of caving, which tend to increase one path length and simultaneously decrease the other, are minimized. Corresponding to the span between the two receivers, the interval transit times calculated represent average values over 40 cms of borehole length. Rock sections taking less than 40 cms of borehole length therefore can be identified on the sonic log only by their influence on the mean interval transit time. As the communication with the surface system operated on the basis of one data request for each centimeter of sonde travel and due to a limited length of the data package that can be sent, the probe information had to be sent in two packages on subsequent data requests and the computer at surface registered one complete measurement only every 2 cms of logging depth. In the log file, compressional wave interval transit times were recorded for each centimeter of sampling, which was due to a repetition of each value.

Images of the complete wave train (amplitudes measured over transit time) the full waveform sonic tool provided in a near and a far receiver every 20 cms represented the wave trains recorded in receiver RX1 and receiver RX2 resulting just from sound emission from transmitter TX1. Like the interval transit times of the compressional waves, the wave trains represent averages of wave behaviour over 40 cms. Again, due to the fact that the communication with the surface system operated on the basis of one data request for each centimeter of sonde travel and due to a limited length of the data package that can be sent, full waveform data was complete only after a certain length of
sonde travel, i.e. 20 cms. The equipment gave the possibility of displaying the wave trains in the amplitude-time mode in which the wave train is shown as a full wave form or in the intensity modulated-time mode in which the wave train is presented as a variable density log. The variable density log provides a top view of the wave train measured over transit time. The waveforms are clipped and the relative amplitude is illustrated by variations in the display colours.

Logs were performed by uphole logging. The probe was raised in the boreholes with a logging speed of 0.5 to 0.8 meters per minute.

Determination of compressional and shear wave velocities from the full wave train:

From the full wave trains, compressional wave velocities \( (v_p) \) of the different rock materials logged are usually established by picking the transit time at the onset of the wave train, i.e. at the first clearly visible arrival (Fig. 1.9). Shear wave velocities \( (v_s) \) assuming "fast" formations, are determined by picking the transit time at the moveout of the shear-pseudo-Rayleigh wave packet. Transit times can be picked for each wave type in the wavetrain of the near (RX1) and the far receiver (RX2) resulting in two different transit time values per depth interval for each wave type. The time difference between these two transit times represents the time needed for the wave to travel over the distance of 40 cms separating the two receivers. Interval transit times can be calculated by dividing the time difference through 40 cms. Also the interval transit times calculated in this way represent borehole compensated values. However, changing diameter spike effects can not be eliminated (see e.g. Labo 1987).

Calculation of Young’s modulus:

On basis of the established compressional and shear wave interval transit times, the dynamic Young’s modulus \( (E_d) \) of the rock material within a depth interval can be calculated assuming linear elastic isotropy of the material:

\[
E_d = \rho \left(3v_p^2 - 4v_s^2\right)/\left((v_p/v_s)^2 - 1\right) \tag{5}
\]

where \( \rho = \) mass density of this depth interval, \( v_p = \) velocity of the compressional wave, \( v_s = \) velocity of the shear wave.

The density of the rock material needed for the calculation was determined by measuring the volume and the mass of different, homogeneous sections of the borehole cores (same core sections as used for triaxial laboratory testing). The different core sections represented differently developed areas of the shear zones. The density determined on these borehole sections was then extrapolated to the remaining parts of the borehole.
1.3.5.3. Spectral Gamma Ray Logging

Spectral gamma ray logging provides a continuous record of the naturally occurring gamma ray radioactivity in the borehole emanating from the decay of the naturally occurring, unstable, radioactive atomic nuclei potassium, uranium and thorium (K, U, Th) and their daughter products. In contrast to simple gamma ray logs, the spectral gamma ray log is able to differentiate quantities of potassium, uranium and thorium. This differentiation is possible, as the radiation from each of the three elements is distinct, with different peak frequencies in the energy spectrum.

Of the three naturally occurring, unstable, radioactive elements, potassium is by far the most abundant (e.g. Rider 1991, Serra et al. 1980). It is common in many rocks of the earth's crust. The original sources of potassium are chiefly the acid and intermediate igneous rocks, in which it is mainly present through feldspars and micas. During the alteration process, feldspars and mica are destroyed and transformed into clay minerals. In evaporites potassium occurs in salt minerals. Original uranium is also generally associated with acid-to-intermediate igneous rocks (e.g. Rider 1991, Serra et al. 1980). During alteration and weathering, uranium is incorporated in soluble salts. The products, which result from the solution of these salts, are unstable and pass easily out of solution. Uranium therefore has a very heterogeneous, sedimentary distribution. Moreover, its solubility in the subsurface, which is a function of its loose attachments, makes it susceptible to leaching and redeposition and its distribution is therefore even more irregular. Thorium also has its origins mainly in the acid and intermediate igneous rocks (e.g. Rider 1991, Serra et al. 1980). During alteration or weathering, thorium has a tendency to concentrate in residual minerals such as bauxite and clay minerals. Amongst the clay minerals, it is abundant in kaolinites. Significant concentrations are also found in heavy minerals such as monazite.

Since clay minerals almost always contain potassium or thorium, gamma ray logging was invented and is still principally used to locate clay-rich layers (increase in gamma radiation) and to calculate shale volume (e.g. Rider 1991). However, there can be other rocks with a high gamma radiation, e.g. intermediate and acid igneous rocks (Rider 1991). The spectral gamma ray log can also be an indirect method for the detection of fractured zones when uranium is present in formation waters circulating especially along fractured areas and then is precipitated and accumulated in the fractures (e.g. Fertl & Rieke 1980, Kamineni et al. 1988, Rider 1991).

Sonde Construction and Measurement:

Logging was performed with a ROBERTSON natural gamma ray spectroscopy sonde. The concentration of the elements potassium, uranium and thorium were calculated by the sonde software from the known ratios of radioactive to non-radioactive isotopes in naturally occurring mixtures (e.g. Serra et al. 1980). To be able to make this transformation, the sonde was calibrated manufacturally with a gamma ray source of known intensity. The sonde automatically applied a compensation for changes in the detector gain characteristics with temperature.
Logs were performed by uphole logging. To collect the gamma rays uniformly over the whole borehole diameter, the sonde was centered. Since the rate of emission of gamma rays is statistical in nature, the rate of counting by the detector is dependent on the basic time constant over which the counting takes place. Fluctuations in the rate of counting will be smaller if the time constant is longer. However, since a borehole tool is constantly moving, too long a time constant will blur bed boundaries and mix several lithologies (Rider 1991). The probe was therefore raised in the boreholes with a recommended logging speed of 0.3 to 0.4 meters per minute.
Fig. 1.2. Location of the Gotthard highway tunnel and the two drill sites. For legend see Fig. 1.1.

Fig. 1.3. Schematic illustration of the orientation of the boreholes at the different drill sites.
Fig. 1.4. Specimen seated and prepared for testing. Note the use of adhesive tape and calcite-like cement.

Fig. 1.5. Servo-controlled, stiff testing machine at the rock mechanics laboratory of the EPFL, Lausanne.
Fig. 1.6. Example of a multi-stage load path applied during the triaxial testing. Example shows load path for sample GR4.
Fig. 1.7. Schematic sketch of the full waveform sonic sonde construction and path of the P-waves producing the measured compressional wave interval transit time.
Fig. 1.8. Propagation paths for compressional (P-), shear (S-), pseudo-Rayleigh, Stoneley and direct fluid waves (modified after Labo 1987).

Fig. 1.9. Full waveform signal showing the arrival of P-, S-, pseudo-Rayleigh and Stoneley waves as a function of time (modified after Labo 1987).
2. Structural Characterization of Shear Zones in the Eastern Aar Massif, Switzerland

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2.1. Introduction

Rock zones containing a high fracture density and/or soft, cohesionless material are of practical importance in tunnelling. Experience in such zones has shown that the fractured and/or soft, cohesionless nature of the rock mass may result in tunnel face instability, excessive overbreak or large deformations of the tunnel profile due to squeezing or swelling (e.g. Deere 1973, Keller & Schneider 1982, Schubert 1993, Schubert & Riedmüller 1997). Important water inflow sometimes associated with flowing ground frequently aggravates such instabilities. In the eastern Aar massif of central Switzerland, where sections of the new Gotthard Base Tunnel will be constructed (Fig. 2.1) (see e.g. Zbinden 1999), experiences during the construction of the Gotthard highway tunnel (Fig. 2.1) showed that heavily fractured areas within shear zones were responsible for overbreaks in the form of chimneys several meters in height (Keller et al. 1987).

To prevent catastrophic failures and to enable a more rigorous planning of the tunnel design including the determination of suitable support requirements and safety regulations within the areas of the shear zones, it is essential for construction works, e.g. the tunnelling activities currently undertaken in the eastern Aar massif, to predict where such shear zones may be expected and to estimate their impact on site characterization of rock mass behaviour. The spatial distribution of the shear zones can be established by mapping on surface and in existing underground openings. The estimation of the impact of the shear zones on site characterization is commonly obtained from analytical and numerical calculations. To properly incorporate the effect of the shear zones within analytical and numerical design methodologies, knowledge concerning the structural characteristics of the shear zones is fundamental as they influence the geotechnical rock mass properties (e.g. Chester & Logan 1986, Habimana et al. 1998).

The presence of shear zones in the eastern Aar massif is common knowledge. The shear zones primarily raised interest due to their prominent geomorphological appearance. Previous studies mainly focused on the distribution of the zones and gave attention to the possibility and the mechanisms of recent movements along these zones (e.g. Jäckli 1951, Eckardt 1957, Steck 1968b, Eckardt et al. 1983). Numerous underground excavations constructed in the eastern Aar massif over the past 150 years resulted in additional knowledge about the spatial distribution of the shear zones (e.g. Schindler 1972, Keller et al. 1987, Schneider 1993). These reports also documented the difficulties the shear zones caused during excavation. Investigation of the structural characteristics of shear zones was only performed in the central part of the Aar massif (e.g. Meyer et al. 1989, Bossart & Mazurek 1990, 1991), where shear zones presumably developed under similar geological conditions as in the eastern part. These studies, which focused on the migration of fluids in the Grimsel rock laboratory, provided detailed maps at a μm- to m-scale and information about the nature of the rock material in several small shear zones. The findings could be extrapolated to shear zones within an area of about 1 km³ around the rock laboratory. The authors described the shear zones as ductile shear zones that were reactivated by brittle deformation.
Given the relative abundance of shear zones in the eastern Aar massif and the occurrence of ground control problems such as the overbreaks experienced in the Gotthard highway tunnel whenever a highly fractured zone was encountered, it is surprising that the structural properties of these shear zones have not been studied rigorously. Besides, it was not verified whether the structural properties established for the area of the Grimsel rock laboratory are applicable to the shear zones of the eastern Aar massif. Also the genesis of the different structural components occurring within the shear zones was not explained convincingly.

This work presents results based on extensive field work at surface and in existing tunnels, the examination of newly drilled borehole cores and a critical analysis of previous reports concerning the structural characteristics of the shear zones in the eastern Aar massif. On basis of these data, a structural model of the zones is proposed. Also suggestions for the genesis of the different structural components occurring within the shear zones are made.

2.2. Geographical and Geological Setting of the Study Area

The Aar massif is situated in the central Swiss Alps (Fig. 2.1). With an area of about 2000 km², it represents the largest crystalline region in Switzerland. It outcrops in the form of a 115 km long and 23 km wide NE striking mountain range. The study area is located in the eastern part of the massif (Fig. 2.1).

The Aar massif consists of a pre-Variscan, polyorogenic and polymetamorphic basement (mainly gneisses, schists and migmatites), which was intruded by late-Variscan magmatic rocks (granites, diorites, syenites, less abundant volcanites and aplitic and lamprophyre dykes) (Labhart 1977, Abrecht 1994) (Fig. 2.1). Carboniferous, Permian and Mesozoic sediments cover the basement and are folded and faulted into it (Labhart 1977, Abrecht 1994). The southern boundary of the massif is marked by a tectonic contact along the Rhine-Rhone-valley (Figs. 2.1 and 2.2). To the east, north and west the massif dips under Mesozoic sediments of the Helvetic domain.

Rocks and structures of the Aar massif are characterized by an overprint of the Tertiary Alpine collision during which the Aar massif was compressed, shifted and thrust towards NW (Steck 1968a, Schmid et al. 1996, Pfiffner & Hitz 1997) (Fig. 2.3). Due to the intense compression, the southern margin of the Aar massif was even backfolded (Steck & Hunziker 1994, Schmid et al. 1996). The pre-Variscan basement, the late-Variscan magmatic rocks and the sedimentary cover of the Aar massif were affected, during this Alpine collision, by heterogeneous ductile deformation and a metamorphic overprint under at most medium greenschist facies/biotite zone conditions (Labhart 1977, Steck 1984, Choukroune & Gapais 1983, Frey et al. 1980). The intensity of ductile deformation and metamorphic overprint generally increases from the NW to the SE of the massif. Deformation was localized on all scales, from mm to km, in ductile and brittle-ductile shear zones (Choukroune & Gapais 1983, Meyer et al. 1989). Alpine shear zones and foliation mainly strike ENE-WSW and dip steeply to the south.
In addition, WNW-ESE and NE-SW striking, steeply dipping shear zones occur (Pflugshaupt 1927, Steck 1968a, Labhart 1977, Choukroune & Gapais 1983). After Choukroune & Gapais (1983), the ENE-WSW striking shear zones represent reverse shear zones, the WNW-ESE and NE-SW striking zones are dextral and sinistral strike-slip zones, respectively.

During uplift of the Aar massif, the area was affected by brittle deformation which, after Steck (1968a), Meyer et al. (1989) and Bossart & Mazurek (1990), generated shear fractures and joint systems locally filled with hydrothermal crystallizations. Due to the weakness of the existing ductile shear zones, most of the brittle fractures concentrated within mylonites.

Uplift of the Alps is still going on (Pavoni 1979, Gubler et al. 1981, Kahle et al. 1997). In the eastern Aar massif uplift rates of up to 0.6 mm/year were determined (Kohl et al. 2000), with the fastest rates generally occurring in the southern part of the massif. Precision levelling measurements across shear zones in the southern Aar massif showing constant differential displacements along some single zones (Eckardt et al. 1983, Fischer 1990) and locally displaced moraines (Jäckli 1951, Eckardt 1957) suggest that quaternary movements take place along isolated shear zones. Movement may be caused by continuing NNW-SSE oriented horizontal compression (Pavoni 1979, Funk & Gubler 1980). Uplift may also be a post-glacial decompression feature (Funk & Gubler 1980).

2.3. Geomorphological Expression of the Shear Zones

The shear zones in the eastern Aar massif are often marked by significant steps in the land surface. The shear zones are frequently represented by natural embankments parallel to the slope axis, which are generally associated with depressions containing small lakes or swamps on the hillside (Fig. 2.4c). The shear zones also often appear as deeply eroded gorges with up to 60 m high walls (Figs. 2.4a and b), for example in the Teiftal.

2.4. Spatial Distribution of the Shear Zones

Air photo analysis and field work revealed that the shear zones appear in all formations of the massif, are mainly some meters wide and can often be followed over several kilometers parallel to their strike (Fig. 2.5). The shear zones can have mean distances of only tens of meters, for example along the southern rim of the Aar massif (Fig. 2.5). Shear zones having the same strike direction and being separated from each other by only a few meters often run parallel over distances of several kilometers (Fig. 2.5). Two main strike directions are recognized (Fig. 2.5).
The majority of the zones strikes ENE-WSW and usually dips steeply towards the south with the dip steepening from the northern to the southern part of the massif (Figs. 2.5 and 2.6). Along the southern rim of the Aar massif shear zones often dip steeply to the north. The ENE-WSW striking shear zones are particularly well developed in the pre-Variscan basement (Fig. 2.7) and are, on a large scale, parallel to the ridge axes, to most of the geological contacts (e.g. lamprophyre dykes, petrographical changes) and to the strike of the dominant Alpine foliation.

A smaller set of shear zones is striking NW-SE (Fig. 2.5). These shear zones dip steeply to the SW or NE and are essentially occurring in the late-Variscan granites (Fig. 2.7).

Shear zones of both strike directions could be followed from late-Variscan granites into the pre-Variscan basement and vice versa.

2.5. Structural Composition

2.5.1. Structure as Seen in Outcrops

Shear zones studied in field and underground outcrops comprised shear zones of both direction sets. However, the majority of the outcrops studied dealt with ENE-WSW striking shear zones.

Continuous outcrops across entire shear zones are rare because shear zones typically provide a weakness along which water seeps, streams flow, and vegetation grows. Outcrops of larger shear zones in underground constructions, which created problems during the construction phase, are generally hidden by a concrete or steel coverage. However, detailed descriptions can be presented for two shear zones believed to be representative for most of the shear zones in the eastern Aar massif. Both shear zones have an ENE-WSW strike and dip steeply to the SE. One shear zone is situated in a granite, the other is developed in gneiss (Fig. 2.8). The outcrops expose the shear zones across their strike and dip.

2.5.1.1. Shear Zone in Granite (Swiss map coordinates (m): 685613/172373)

Figure 2.9 shows a shear zone in the Variscan Aar granite. The shearzone is about 2 m wide and is exposed parallel to its strike over a distance of 600 m and a height of 100 m following topography. Within the exposed area, the shear zone does not branch and is relatively constant in thickness.

The host granite mainly consists of quartz and feldspar with subsidiary muscovite, biotite and chlorite. The rock is coarse-grained and slightly deformed, showing an ENE-WSW striking, cm-spaced foliation wrapping around feldspar porphyroclasts. The foliation is locally folded. Meter- to dm-spaced fractures are mainly subparallel to the foliation.
The mica content and the foliation intensity increase towards the inner part of the shear zone, producing a strongly foliated mylonite or even phyllonite in the shear zone center. The fracture density also increases towards the center. As fractures do not show any visible offset or striation lineation, they could not be established as brittle faults. Fracture density is strongest within two phyllonitic zones oriented parallel to the local foliation. The zones of highest fracture density contain structureless areas of fractured quartz clasts enclosed in a light brown, very fine-grained material.

2.5.1.2. Shear Zone in Gneiss (Swiss map coordinates (m): 693850/169700)

The shear zone developed in the Southern Granite Gneiss of the pre-Variscan basement and located in the southern part of the eastern Aar massif is shown in Figures 2.10 and 2.11. The shear zone has a width of about 10 m and is exposed parallel to its strike over a distance of 250 m and a height of 100 m following topography.

The host rock is an orthogneiss mainly consisting of quartz, feldspar and mica. The orthogneiss has a strong, ENE-WSW striking foliation and shows cm-big feldspar porphyroclasts. Meter- to dm-spaced fractures are mainly parallel to the foliation.

Foliation intensity of the gneiss and the content of mica clearly increase and grain size reduces towards the core of the shear zone. Areas showing an extremely high mica content are phyllonites. The core of the shear zone is represented by an extremely fine-grained and strongly foliated ultramylonite in which few feldspar porphyroclasts with diameters of up to 1 cm are present. The ultramylonite has a greenish colour due to a high biotite content. Quartz veins mainly oriented parallel to the foliation occur everywhere in the shear zone.

Fractures overprint ductile structures. In this outcrop also, the fractures could not be established as faults. Highest fracture density occurs along a transition from phyllonite to ultramylonite. It is assumed that the density of microfractures increases significantly with the density of the visible fractures, because the rock material is extremely friable in the zone of highest fracture density. This zone also contains cm-thick stripes of extremely fine-grained, cohesionless material forming an anastomosing network around lozenge-shaped, coarser-grained or quartz-rich rock domains. The stripes show an internal foliation, which is locally folded or dislocated.

2.5.2. Structure as Seen in Borehole Cores

As structures and rock material at surface are usually heavily weathered and, therefore, are not necessarily representative of structures and materials at depth, two boreholes were cored from the underground at two different sites. Both boreholes were oriented normal to the strike and dip of ENE-WSW striking shear zones and, therefore, were also oriented perpendicular to the penetrative ENE-WSW striking foliation. The boreholes had the purpose to provide undisturbed cores for core logging and thin section analysis.
Both drill sites were located in the safety tunnel of the Gotthard highway tunnel (Fig. 2.8). The drill sites both had an overburden of 400 to 500 m, which should have prevented the rocks from any surface influence. One drill site sampled a shear zone in the Variscan Aar granite (borehole Ia), the second one was positioned in a shear zone in the Southern Gneiss Zone of the pre-Variscan basement (borehole IIa). Both drill sites had been sites of overbreaks during the construction of the safety tunnel.

2.5.2.1. Borehole in Granite (tunnelmeter from northern portal: 2014)

Borehole Ia (Fig. 2.12) starts in a light grey coloured granite mainly consisting of quartz (~ 65 vol.-%), feldspar (~ 20 vol.-%) and biotite (~ 10 vol.-%). Muscovite, chlorite, sphene, opaque minerals/ores, remnants of amphiboles and zircons are accessory minerals. The granite is coarse-grained and isotropic and contains fractures with m- to dm-spacing.

From a borehole length of about 3 m onwards, the core becomes slightly darker due to an increased amount of biotite. Additionally, an ENE-WSW striking, cm-spaced foliation and a stretching lineation defined by the preferred orientation of mica, boudins and feldspar-porphyroclasts are visible. The rock becomes a mylonite. Foliation intensity and the content of biotite and muscovite further slightly increase and grain size simultaneously decreases with increasing borehole length. Foliation intensity and mica content are highest and grain size is smallest in between 4.6 and 5.1 m where the rock is a ultramylonite. Further on along the borehole, the foliation intensity decreases.

Like in the field outcrops, ductile structures are overprinted by fractures. Fracture density increases from a length of about 3.1 m onwards and is significantly increased from a length of 3.6 m onwards. Macroscopically visible fractures are predominantly parallel to the foliation. Microfractures recognized in thin section are mainly oblique to the foliation. The microfractures occur as mostly open, trans- and intragranular fractures (Fig. 2.13). Intragranular fractures especially occur in quartz and feldspar grains. As the macro- and microfractures generally show no significant shear displacement, they seem to have formed by opening. From a borehole length of about 3.65 m onwards, where the content in coarse-grained quartz is markedly smaller, transgranular fractures parallel and oblique to the foliation and filled with μm-thick, extremely friable or cohesionless, light brown, fine-grained material occur (Fig. 2.14). The fine-grained fracture infill is a structureless arrangement of biotite, muscovite, quartz and chlorite (e.g. Fig. 2.15). In Figure 2.16 the generally sharp boundary between mylonite and fracture infill is visible. Filled fractures and open fractures merge. The thickness of the fracture infill and therefore the abundance of the structureless, fine-grained material increases with borehole length until the occurrence of heavily, fragmented, cohesionless material.

An extremely high fracture density, occurring from a borehole length of about 3.7 m onwards, resulted in the disruption of the rock fabric. A structureless arrangement of several cm-big, mainly angular rock fragments of the mylonite separated by an up to 1
A mm thick matrix of fine-grained material was formed (Fig. 2.14). The fine-grained matrix is extremely friable and consists of biotite, muscovite, quartz and chlorite.

Sharp boundaries at a length of about 4.2 m and 4.5 m separate the cohesive material from an about 25 cm thick zone of relatively soft, cohesionless material which also consists of up to several cm-big rock fragments, separated by a cohesionless matrix (Figs. 2.17 and 2.18). The noncohesiveness of the matrix seems to be due to the fact, that grain boundaries are open pores. The matrix is thicker than in the cohesive rock mass and the rock fragments appear to be floating in the matrix, with few grain-to-grain contacts. In some places the rock fragments have been rotated since the foliation of different rock fragments has different orientations. The rock mass is a microbreccia. A grain size analysis of the microbreccia (sieve and hydrometer analyses according to the Swiss Standards SNV 670 810 and SNV 670 816) revealed that the material consists mainly of gravel- (grain sizes bigger than 2 mm) and sand- (grain sizes bigger than 0.063 mm) grain sizes (Fig. 2.19a).

From a length of 4.5 m onwards, a cohesive mylonite occurs again. The mylonite contains fractures parallel and oblique to the foliation, locally filled with mm-thick zones of fine-grained material, and is friable. From 4.6 to 5.0 m, in the ultramylonite, the rock material is extremely friable.

At about 5.1 m, an other small cohesionless zone occurs. It is therefore obvious that the cohesionless zones concentrated along transitions from mylonite to ultramylonite.

Further on in the borehole, foliation intensity decreases and the rock becomes less and less fractured.

2.5.2.2. Borehole in Gneiss (tunnelmeter from northern portal: 3454)

Borehole IIa (Fig. 2.20) starts in a light grey gneiss mainly consisting of quartz (~ 55 vol.-%), plagioclase (~ 20 vol.-%) and mica (biotite and muscovite) (~ 20 vol.-%). Pyrite, carbonate, sphene and opaque minerals/ores different from pyrite are accessory minerals. The biotite has locally been transformed into chlorite (pseudomorphoses of biotite by chlorite). The gneiss is coarse-grained and shows an ENE-WSW striking, cm- to mm-spaced foliation and a stretching lineation defined by the linear alignment of mica, feldspar- and pyrite porphyroclasts. The rock contains fractures with cm- to dm-spacing mainly oriented oblique to the foliation.

From a length of 1.9 m onwards, the gneiss shows a stronger foliation and stretching lineation, is finer-grained (e.g. due to the intensified recrystallisation of quartz) and contains a higher amount of mica (~ 30 vol.-%). The amount of plagioclase simultaneously decreases. The stronger foliation, the fine-grained nature and the high amount of mica are typical of a mylonite. At a length of about 5.0 m, foliation becomes very strong and grains are very fine. The rock contains a high amount of biotite and
muscovite (~ 40 vol.-%) and almost no more plagioclase. The rock is a ultramylonite. From a borehole length of 7.2 m onwards the mylonite occurs again.

The fracture density in the mylonite increases from a borehole length of 2.6 m onwards, with both macroscopically visible fractures and microfractures oriented oblique and parallel to the foliation. Microfractures are transgranular and intragranular and developed by opening (Figs. 2.21, 2.25) and by shearing (Figs. 2.22, 2.27). Along the shear fractures only small shear displacement has taken place. The fractures can be open (Figs. 2.21, 2.22) but are mainly filled by a structureless arrangement of fine-grained muscovite, biotite, quartz, carbonate and remnants of plagioclase and pyrite (Figs. 2.23, 2.24, 2.25, 2.26, 2.27). The biotite of this infill is also locally transformed into chlorite. Occasionally, the fracture infill is cohesionless (Fig. 2.28). Like in the granite material, filled fractures and open fractures merge. The thickness of the fracture infill increases from an average of about 50 μm to a thickness of about 100 μm at a borehole length of 5.0 m. Due to the high fracture density and the presence of cohesionless fracture infill, the mylonite is significantly more friable than in the section before 2.6 m.

At a length of about 5.0 m, where the ultramylonite starts, density of macroscopically visible fractures does not change. However, the density of microfractures and the thickness of the fracture infill further increases with increasing borehole length to an average thickness of about 500 μm at 6.9 m. The ultramylonite is more friable than the mylonite of the core section up to 5.0 m.

The white, soft, almost cohesionless, structureless material occurring between 7.1 and 7.2 m is separated by sharp boundaries from the surrounding ultramylonite and mylonite. In thin section, the material is composed of, on average, 1 mm large, angular and rounded rock fragments floating with few grain-to-grain contacts in a fine-grained, cohesionless matrix (Figs. 2.29, 2.30, 2.31, 2.32). The rock is a microbreccia. According to a grain size analysis (sieve and hydrometer analyses according to the Swiss Standards SNV 670 810 and SNV 670 816), the microbreccia mainly consists of sand- (grain sizes bigger than 0.063 mm) and silt- (grain sizes bigger than 0.002 mm) grain sizes (Fig. 2.19b). In comparison to the cohesionless material in borehole 1a, clay-grain sizes are more abundant (up to 5 %). The rock fragments are often composed of polygonal, annealed quartz grains. Other rock fragments contain finer-grained quartz, carbonate and mica. These fragments are foliated and therefore must be derived from mylonite and ultramylonite. The foliation within different rock fragments has different orientations and therefore shows some rotation. According to thin section-, X-ray diffractometry- and microprobe-analyses, the structureless matrix is mainly composed of fine-grained biotite, muscovite, probably some illite, quartz, remnants of plagioclase, carbonate and pyrite (Figs. 2.33 and 2.34). The matrix therefore consists of the same material as the fracture infill described earlier.

Beyond 7.2 m, to the end of the borehole at 15.1 m, cohesive mylonite with few fractures occurs.
2.6. Discussion

2.6.1. Structural Characterization of the Shear Zones

The study of outcrops at surface and in underground constructions and the examination of cores suggest that the structural characteristics of the shear zones in the eastern Aar massif are similar, independent of their strike and the lithology they affect.

The deformation structures are primarily ductile. A characteristic progressive evolution from the host rock to mylonite and locally ultramylonite is documented by the formation and intensification of a foliation and a stretching lineation and the progressive replacement of the host rock by a finer-grained material containing newly crystallized and recrystallized minerals. Rock compositions simultaneously show an increase in the mica content. Towards the mylonitic and ultramylonitic zones, the network of small-scale shear zones anastomosing around more competent, lozenge-shaped lenses increases in density.

The shear zones also show a higher fracture density than the host rock. The fractures oriented parallel and oblique to the foliation can be open but are mainly filled with a fine-grained, often cohesionless, structureless arrangement of micas and quartz. Highest fracture densities are restricted to strongly foliated and fine-grained parts of the shear zone and are mainly developed along transitions from mylonite to ultramylonite. Along with the fracture density, the thickness of the fracture infill increases rendering the mylonite more and more friable towards the zone of highest fracture density. In the heavily fractured zones, structureless and almost cohesionless microbreccia consisting of mylonite fragments enclosed in a fine-grained, cohesionless matrix occur. The matrix is composed of the same material as the fracture filling and is therefore assumed to represent this infill.

Substantial structural differences among different shear zones, which can generally be described as ductile shear zones containing a fractured zone, may occur due to differences in the bulk rock composition and the original structure of the host rock (e.g. foliated gneiss versus isotropic granite). Observed differences in between shear zones in isotropic granite and in gneiss are listed in Table 2.1.

The structural composition of the shear zones observed in the eastern Aar massif is similar to the composition of the shear zones described in the area of the Grimsel rock laboratory. After Meyer et al. (1989) and Bossart & Mazurek (1990, 1991), the mylonites and ultramylonites also contain low cohesion breccia composed of mylonite fragments enclosed in a fine-grained matrix of mica, quartz, feldspar, chlorite, epidote and carbonate. The breccia also preferentially formed in areas of high ductile strain and high deformation gradients.
2.6.2. Structural Model of the Shear Zones

According to models of joint zones and fault zones (e.g. Chester & Logan 1986, Wallace & Morris 1986, Chester et al. 1993, Braathen & Gabrielsen 1998), it is proposed that the shear zones in the eastern Aar massif can be subdivided into 3 different structural zones characterized by different fracture densities and mica contents. These are, with increasing fracture density and mica content, a strongly foliated zone, a fractured zone and a heavily fractured, cohesionless zone (Fig. 2.35). The mica content is increased towards the heavily fractured, cohesionless zone due to an increasing thickness of fracture infill and because the intensely fractured zones are especially developed along mylonite to ultramylonite transitions. The zones are oriented approximately parallel to the rock foliation and often form a symmetric sequence on both sides of the heavily fractured, cohesionless zone which may anastomose within the fractured zone (e.g. Fig. 2.10). The cohesionless zone may eventually be absent (e.g. Fig. 2.9). Due to different rheological properties, the zones may vary in thickness along both their strike and dip.

The strongly foliated zone may have a thickness of up to several meters. The zone is characterized by mylonites or ultramylonites slightly more fractured than the host rock.

The strongly foliated zone gradually changes into the fractured zone. At surface, the fractured zone is often weathered. The zone may be a few meters thick. Like the strongly foliated zone, the fractured zone is composed of mylonite or ultramylonite, but these rocks are more fractured. The fracture infill is up to 1 mm thick. Extremely strong fragmentation towards the heavily fractured, cohesionless zone may result in the disruption of the rock fabric. Under these circumstances a structureless, cohesive microbreccia is formed consisting of about 1 cm large rock fragments of mylonite separated by a mm-wide fine-grained matrix. Due to the high degree of fragmentation and the thickness of the fracture infill, the rock material in the fractured zone is extremely friable.

The cohesionless zone has a distinct and sharp contact with the adjacent cohesive material. At surface, this zone is strongly weathered and easily eroded. Its thickness varies from a few centimeters to several decimeters. The zone, which is usually lighter in colour than the surrounding rock, is composed of structureless microbrecciae that can be distinguished from microbreccia of the fractured zone by the increased amount of matrix material and their cohesionless or almost cohesionless nature. The material of the cohesionless zone may have some cohesion due to the fine-grained nature of the matrix minerals. However, swelling clay minerals could not be observed.
2.6.3. Fracturing Mechanism and Genesis of Shear Zones in the Eastern Aar Massif

The examination of the genesis and age of the shear zones in the eastern Aar massif goes beyond the scope of this work. At this point only working hypotheses are set.

2.6.3.1. Fracturing Mechanism

The studies show that similar fractures and brecciae occur at surface and at depths of 400 to 500 meters. The fact that filled and open fractures oriented parallel and oblique to the foliation do not destroy each other but merge suggests that all fractures were formed during the same deformation event. Therefore, the majority of open fractures observed in surface outcrops does not seem to represent near surface stress relaxation joints. Similarly, the open fractures and brecciae observed in the borehole cores are not due to drilling processes, the excavation process of the tunnel or to stress redistribution related to tunnelling.

Fracturing can also not be explained by faulting, since the fractures do not display significant shear displacements. A lack of significant shear displacements was also reported for the fractures within the shear zones in the Grimsel area (Bossart & Mazurek 1990). It is therefore not possible that quaternary differential displacements like they were measured by Eckardt et al. (1983) and Fischer (1990) or displacements of moraines like they were observed by Jäckli (1951) and Eckardt (1957) have taken place along the examined shear zones. Locally occurring faults showing striation lineation are independent of the shear zones.

The abundance of fractures developed by opening and the observation of rotated mylonite fragments in the microbrecciae suggest a bulk volume increase. The volume increase may have been produced by an increase in the amount of pore fluids probably resulting from dehydration reactions during the metamorphic overprint or the intrusion of hydrothermal solutions. Hydrothermal overprinted granites are common in the eastern Aar massif (Keller et al. 1987). The increased amount of pore fluids elevated the pore fluid pressure causing fracturing, i.e. hydrofracturing.

2.6.3.2. Age of Shear Zones and Fracturing

Shear zones situated in the pre-Variscan basement and penetrated by granitic dykes must be older than or contemporaneous with the late-Variscan granite. Shear zones developed in the granite and those that can be traced from the granite into the pre-Variscan basement are of Alpine age (Steck 1966, 1968a, Choukroune & Gapais 1983). Shear zones in the basement which cannot be traced into the granite may have been reactivated during the Alpine orogenesis or were newly formed along pre-existing zones of weakness (e.g. lamprophyre dykes, petrographical changes). Milnes & Pfiffner (1977) describe intense penetrative deformation in the Aar massif in upper Oligocene times. Rb-Sr age determinations on phengite, biotite, plagioclase and K-feldspar of
mylonites developed in the Grimsel granodiorite yielded about 26 Ma (Kralik et al. 1992).

The fractures are clearly superimposed on the ductile deformation structures and, therefore, are younger. As the structural characteristics of the shear zones are similar and independent of their strike and the lithology they affect, fractures within late-Variscan granites and in the pre-Variscan basement seem to have been formed by the same event. This event must have taken place later than the Alpine formation of the shear zones in the granite. Kralik et al. (1992) reported a tectonic-hydrothermal pulse occurring about 26 Ma ago which could have caused the hydrofracturing. However, Rb-Sr and K-Ar ages obtained from the fine-grained material of the breccia in the Grimsel area suggest ages of 10 to 7 Ma (Kralik et al. 1992).

2.7. Conclusions

Results from this study show:

(1) The shear zones examined in the eastern Aar massif are primarily ductile but also contain a considerably larger number of fractures than the host rock. The shear zones comprise 3 different subzones with different fracture densities and mica contents: With increasing fracture density and mica content, these zones are a strongly foliated, a fractured and a heavily fractured, cohesionless zone.

(2) The fractures mainly were developed by opening and show an infill of fine-grained biotite, muscovite and quartz. It is argued that the fractures were created by hydrofracturing. The fractures are superimposed on the ductile deformation structures of the Alpine deformation phase and therefore have been formed later.
Fig. 2.1. Geographical and geological setting of the Aar massif (after Abrecht 1994), location of the study area and location of Gotthard Base Tunnel and Gotthard highway tunnel.
Fig. 2.2. Cross section through the Aar massif along line A-B of Fig. 2.1. For legend see Fig. 2.1. (after Labhart 1998)
Fig. 2.3. Kinematic evolution of the Aar massif (simplified after Schmid et al. 1996).
Fig. 2.4. Geomorphological appearance of shear zones in the eastern Aar massif:
a) gorge with steep walls, b) gorges parallel to the slope axis, c) natural embankments parallel to the slope axis.
Fig. 2.5. Example of the spatial distribution of shear zones mapped in the eastern Aar massif. Section shows the southern rim of the Aar massif in the area north of Sedrun. Topography is printed with approval of the Swiss Federal Office of Topography.
orientation of shear zones in the northern and middle part of the study area:

<table>
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histogram of strike direction

orientation of shear zones in the southern part of the study area:

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<td>Axial</td>
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<td>N = 295</td>
<td>N = 295</td>
</tr>
</tbody>
</table>

histogram of strike direction

Fig. 2.6. Orientation of shear zones in the eastern Aar massif. Histograms of strike direction have class intervals of 10°. Equal area plots show lower hemisphere projection.
Fig. 2.7. Comparison of the orientation of shear zones in the eastern Aar massif within formations of different age. Histograms of strike direction have class intervals of 10°.
Fig. 2.8. Location of the outcrops and boreholes described. For legend see Fig. 2.1.
Fig. 2.9. Detailed photos and schematic drawing of the shear zone situated in the area of the Variscan Aar granite body.
Fig. 2.10. Photo of the outcrop and detailed schematic drawing of the shear zone situated in the area of the Southern Granite Gneiss of the pre-Variscan basement.
Fig. 2.11. Detailed photo and schematic drawing of the central part of the shear zone situated in the area of the Southern Granite Gneiss of the pre-Variscan basement.

light grey to light brown mylonite/phyllonite, foliated in cm- to mm-scale, fractured in cm- to dm-scale parallel to foliation

quartz, quartz partly shows mm- to cm-wide foliation and fractures with mm- to cm-spacing parallel to foliation, some quartz has a round shape and shows no structures, amorphous quartz

(area painted in grey)
light brown material, coarse- to fine-grained, coarser-grained material and areas of quartz rich rock are surrounded by anastomosing system of extremely fine-grained material, mm-wide foliation and folds are visible in this zone, minor faults can be realized which dislocated the foliation

light brown to light grey mylonite/phyllonite, foliated in mm-scale, fractured in cm- to dm-scale parallel to foliation

green-grey material, ultramylonite, approximately 1.5 m wide, foliated in mm-scale, fractured with cm-spacing parallel to foliation
Legend for profiles of cores la and IIa in Figs. 2.12 and 2.20:

- **isotropic Aar granite**

- **ductilely overprinted rock,**
  - coarse-grained, foliated in cm- to mm-scale;
  - protomylonite

- **strongly ductilely overprinted material: foliated in cm- to mm-scale,**
  - minerals are fine-grained, high amount of mica;
  - mylonite and ultramylonite,
  - fractures with cm- to mm-spacing

- **strongly ductilely overprinted material: foliated in cm- to mm-scale,**
  - minerals are extremely fine-grained, high amount of mica;
  - mylonite and ultramylonite,
  - material is characterized by fractures,
  - material is friable, soft, cohesive

- "cohesionless, heavily fractured zone",
  - breccia zone, microbreccia,
  - heavily fragmented, cohesionless, structureless material

- **foliation**

- **fractures**
material of sample GR3

3.5 m

heavily fragmented rock material, fractures enclose wider, light-brown, cohesionless area of fine-grained mica

material of sample GR4

4.0 m

sharp boundary in between cohesive and cohesionless material

rock structures (e.g. foliation) are not destroyed, but cross the boundary without showing any rupture

material of sample GR5

4.5 m

sharp boundary in between cohesionless and cohesive material

material, which shows ductile structures; foliation in mm-scale, porphyroclasts, boudins are visible,
up to 1.5 cm thick, light-brown, cohesionless or extremely friable layers of fine-grained mica and + chlorite,
layers are parallel to foliation,
fractures parallel and oblique to foliation, fractures are mainly open, but can also be filled with fine-grained mica and quartz,
material seems to be similar to rock up to 3.70 m, but has more fractures and material is more friable, soft,
material is cohesive

material which shows ductile structures; foliation in mm-scale, very fine-grained material, mylonite/ ultramylonite,
material is cohesive, contains fractures mainly parallel to foliation,

Fig. 2.12. Sketch of the core of borehole Ia located within Variscan Aar granite.
Fig. 2.13. Thin section photo of sample GR3 from borehole 1a (granite). Picture shows transgranular, open fractures intersecting quartz (qz) and feldspar minerals (plag, kfsp) and finer-grained foliation domains consisting mainly of biotite, muscovite and sphene. Transgranular fractures are oriented almost perpendicular to foliation. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.14. Thin section photos of sample GR4 from borehole la (granite). Photos show mm-wide opened fractures having an infill, which is arranged in a structureless manner. Black areas represent rock fragments, light green coloured areas had been pore spaces, which are filled now with a fluorescent resin. Abundance of light green colour therefore shows the cohesionless nature of the fracture infill. Fluorescence of the resin is visible, as thin section was illuminated with UV fluorescent light. For the preparation of the fluorescence microscopy specimens, the specimen was saturated with fluorescenting resin.

Width of view is 0.5 mm.
Fig. 2.15. Thin section photo of sample GR4 from borehole la (granite). Photo shows a fractured plagioclase-grain (plag). The central fracture is oriented oblique to foliation and has a fine-grained infill of quartz, biotite and muscovite. Minerals representing the fracture infill are arranged in a structureless manner. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.16. Thin section photo of sample GR4 from borehole 1a (granite). Picture shows the sharp borderline in between a rock fragment (upper part of the figure) and a fracture infill (lower part of the figure). Rock fragment represents a part of the ductilely deformed host rock and mainly consists of feldspar (plag, kfsp), quartz (qz) and biotite. The biotite was partly transformed into chlorite (pseudomorphoses). Fracture infill consists of fine-grained quartz, biotite, chlorite, muscovite and opaque minerals. Infill minerals are arranged in a structureless manner. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed
Fig. 2.17. Thin section photo of sample GR5 from borehole 1a (granite). Photo shows cm-large rock fragments (consisting mainly of quartz, feldspar, biotite, +/- chlorite and muscovite), which were originally separated by a fine-grained matrix. Matrix was washed out due to preparation of the thin section. Picture shows minerals under crossed polars.
Fig. 2.18. Thin section photo of sample GR5 from borehole 1a (granite). Photo shows mm-large rock fragments originally separated by a fine-grained matrix. Matrix was washed out due to the preparation of the thin section and cavity was filled instead with a fluorescent resin. Fluorescence of the resin is visible, as thin section was illuminated with UV fluorescent light. Width of view is 0.5 mm.
Heavily fractured, cohesionless material from meter 4.2 to 4.5 of core Ia (granite).

Heavily fractured, cohesionless material from meter 7.1 to 7.2 of core IIa (gneiss).

Fig. 2.19. Grain size distribution of the heavily fractured, cohesionless material within the borehole cores in granite and in gneiss.
gneiss, light grey in colour, foliated in mm- to cm-scale, fine-grained, high amount of mica, mylonite, contains many fractures parallel and oblique to foliation, contains several cm-thick quartz veins, material is friable

**material of sample GN2**

material for sample GN3 was taken from the core at a borehole length of about 5.5 m and consists of fractured ultramylonite

material for sample GN4 was taken from the core at a length of about 6.15 m and consists of strongly fractured ultramylonite

undisturbed sample, grey material, foliated in mm-scale, fractured along foliation planes, has cohesion, but is extremely friable

**material of sample GN5**

cohesionless material, white colour, material consists of light grey, foliated rock fragments, fractures separating the rock fragments are filled with fine-grained mica and quartz, microbreccia, cohesive material, grey colour, foliation in mm-scale, fine-grained, mylonite, only few fractures parallel to foliation planes

Fig. 2.20. Sketch of the core of borehole IIa located within the Southern Gneiss Zone.
Fig. 2.21. Thin section photo of sample GN3 from borehole IIa (gneiss). Picture shows a transgranular, open fracture intersecting quartz minerals (qz) and finer-grained foliation domains. Finer-grained foliation domains mainly consist of fine-grained quartz, biotite, muscovite and calcite minerals. Larger biotite minerals are primarily found in the tails of former plagioclase-porphyroclasts. Transgranular fracture is oriented almost perpendicular to foliation. Fracture has no infill. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.22. Thin section photo of sample GN3 from borehole IIa (gneiss). Picture shows a transgranular, open shear fracture intersecting a fine-grained foliation domain. Fine-grained foliation domain mainly consists of biotite, muscovite and quartz. Transgranular fracture is oriented oblique to foliation. Note that fracture shows horsetails at its beginning. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.23. Thin section photo of sample GN2 from borehole Ila (gneiss). Photo shows a fracture having an infill of fine-grained quartz, biotite and muscovite. The minerals are arranged in a structureless manner. Fracture infill was partly washed out due to preparation of the thin section. Fracture infill is bordered by a fine-grained foliation domain consisting mainly of fine-grained quartz, biotite and muscovite and an area of coarser-grained minerals. Fracture is oriented parallel to foliation. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.24. Thin section photo of sample GN2 from borehole IIa (gneiss). Photo shows a fracture having an infill of fine-grained biotite, muscovite and quartz. The minerals are arranged in a structureless manner. Fracture is oriented parallel to foliation. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.25. Thin section photo of sample GN3 from borehole IIa (gneiss). Picture shows a transgranular fracture having an infill of fine-grained quartz, biotite, muscovite and carbonate. The minerals are arranged in a structureless manner. The transgranular fracture is oriented almost perpendicular to foliation. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.26. Enlargement of Fig. 2.25.
Fig. 2.27. Thin section photo of sample GN3 from borehole I1a (gneiss). Photo shows a transgranular shear fracture having an infill of fine-grained quartz, biotite and muscovite. The minerals are arranged in a structureless manner. The transgranular fracture is oriented oblique to the foliation. Fracture is surrounded by quartz, biotite, muscovite and carbonate minerals. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Picture shows minerals under crossed polars.
Fig. 2.28. Thin section photo of sample GN5 from borehole IIa (gneiss). Picture shows mm-wide opened fracture having an infill, which is arranged in a structureless manner. Black areas represent rock fragments, green coloured areas had been pore spaces, which are filled now with a fluorescent resin. Abundance of light green colour therefore shows the cohesionless nature of the fracture infill. Fluorescence of the resin is visible, as thin section was illuminated with UV fluorescent light. Width of view is 0.5 mm.
Fig. 2.29. Thin section photo of sample GN5 from borehole IIa (gneiss). Photo shows rock fragments enclosed in a fine-grained matrix. Arrangement has no structure. Rock fragments mainly consist of coarse-grained, statically recrystallized and fine-grained, dynamically recrystallized quartz. Largest rock fragment in this picture is 800 μm wide. Fine-grained matrix surrounding the rock fragments is mainly composed of biotite and muscovite. Picture shows minerals under crossed polars.
Fig. 2.30. Thin section photo of sample GN5 from borehole IIa (gneiss). Photo shows rock fragments enclosed in a fine-grained matrix. Arrangement has no structure. Rock fragments mainly consist of fine-grained, dynamically recrystallized and coarse-grained, statically recrystallized quartz. Largest rock fragment in this picture is about 500 μm wide and 1mm long. Fine-grained matrix surrounding the rock fragments is mainly composed of biotite and muscovite. Matrix seems to part the rock fragments mainly in two directions. Picture shows minerals under crossed polars.
Fig. 2.31. Thin section photo of sample GN5 from borehole Ha (gneiss). Photo shows rock fragments enclosed in a fine-grained matrix. Arrangement has no structure. Rock fragments mainly consist of coarse-grained, statically recrystallized and fine-grained, dynamically recrystallized quartz. Besides quartz, carbonate, biotite and muscovite can be clearly seen in the fragments. Largest rock fragment in this picture is 700 μm wide and 1.5 mm long. Fine-grained matrix surrounding the rock fragments is mainly composed of biotite and muscovite. Matrix seems to part the rock fragments preferably in two directions. Picture shows minerals under crossed polars.
Fig. 2.32. Thin section photo of sample GN5 from borehole IIa (gneiss). Picture shows mm-large, angular and rounded rock fragments enclosed in a fine-grained matrix. Black areas represent rock fragments, green coloured areas had been pore spaces within the fine-grained matrix, which are filled now with a fluorescent resin. Abundance of green colour therefore shows the cohesionless nature of the matrix. Fluorescence of the resin is visible, as thin section was illuminated with UV fluorescent light. Note that the rock fragments in this specimen resulting from gneiss material are much more fine-grained than the rock fragments in microbreccia resulting from granite material. Width of view is 0.5 mm.
Fig. 2.33. X-ray diffractometry of heavily fractured, cohesionless material from meter 7.1 to 7.2 of core IIa, sample a. X-ray diffractometry plot shows well defined peaks representing muscovite, quartz and albite.
Fig. 2.34. X-ray diffractometry of heavily fractured, cohesionless material from meter 7.1 to 7.2 of core IIa, sample b. X-ray diffractometry plot shows well defined peaks representing muscovite, quartz and albite.
Fig. 2.35. Schematic illustration of the structural composition of shear zones in the eastern Aar massif.
Tab. 2.1. Comparison of structural differences between shear zones in isotropic granite and shear zones in gneiss.

<table>
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<th>shear zones in granite</th>
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<tbody>
<tr>
<td>width of shear zone</td>
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<td>wide</td>
</tr>
<tr>
<td>fracture density</td>
<td>small</td>
<td>high</td>
</tr>
<tr>
<td>grain size of microbreccia</td>
<td>large</td>
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3. Geomechanical Properties of Shear Zones in the Eastern Aar Massif, Switzerland and their Implication on Tunnelling

S. Laws, E. Eberhardt, S. Loew & F. Descoëndres

Submitted to
Rock Mechanics and Rock Engineering
3.1. Introduction

Rock zones containing a high fracture density and/or soft, cohesionless material are highly problematic when encountered during tunnel excavation. Experience in such zones has shown that the fractured and/or soft, cohesionless nature of the rock mass may result in tunnel face instability, excessive overbreak or large deformations of the tunnel profile due to squeezing or swelling (e.g. Deere 1973, Müller 1978, Keller & Schneider 1982, Schubert 1993, Schubert & Riedmüller 1997, 2000). Extreme water inflow sometimes associated with flowing ground frequently aggravates such instabilities. In the eastern Aar massif of central Switzerland, where sections of the new Gotthard Base Tunnel will be constructed (Fig. 3.1) (see e.g. Zbinden 1999), experiences during the construction of the Gotthard highway tunnel (Fig. 3.1) showed that heavily fractured areas within shear zones were responsible for overbreaks in the form of chimneys several meters in height (Keller et al. 1987).

To enable a rigorous planning of the construction method and the tunnel design including the determination of suitable support requirements and safety regulations, it is essential for the construction of a tunnel, e.g. the Gotthard Base Tunnel, to estimate the impact of the shear zones on site characterization and rock mass stability. To properly incorporate the effect of the shear zones within analytical and numerical design methodologies, knowledge concerning the rock mass strength and deformation characteristics is required. Previous studies examining the geomechanical properties of heavily fractured rock masses, especially fault zones, have shown that rock strength and deformation parameters vary depending on the different structural characteristics within the zones, e.g. fracture densities (e.g. Chester & Logan 1986, Habimana et al. 1998). In general, it is assumed that with an increasing degree of fragmentation there is a reduction in rock strength and Young’s modulus, and an increase in plastic yielding.

Shear zones in the eastern Aar massif have been widely studied, but mostly with respect to their spatial distribution (e.g. Jäckli 1951, Eckardt 1957, Steck 1968b, Eckardt et al. 1983, Keller et al. 1987, Schneider 1993). Given the relative abundance of the shear zones and the occurrence of ground control problems such as the overbreaks experienced during the construction of the Gotthard highway tunnel whenever a larger zone was encountered, it is surprising that the geomechanical properties and their variance within the shear zones have never been rigorously studied. Studies examining the geomechanical properties of such heavily fractured shear zones and fault zones and relating these properties to the structural characteristics are generally rare.

This paper presents the results of an extensive investigation into the geomechanical properties of the shear zones in the eastern Aar massif and their dependence on structural characteristics. Methodologies and trends established herein may also be applied to other zones in crystalline rocks showing similar structural properties. The paper first describes the structure of the shear zones and then presents results from a series of triaxial compression tests performed on different shear zone samples. The analysis concentrates on the description of the rock behaviour with respect to the applied stress and includes the determination of elastic constants and Mohr-
Coulomb strength parameters. Variations in rock behaviour and the validity of the parameter values are assessed with respect to the structural composition of the shear zones.

3.2. Shear Zone Structure and Characteristics

3.2.1. General Characterization of Shear Zones

Shear zones are narrow zones of high strain characterized by spatial gradients of finite strain. The amount of strain is generally highest within the center of the shear zone, decreasing outward into the wall rocks adjacent to the zone (e.g. Ramsay & Huber 1987, Davis & Reynolds 1996). Another characteristic feature of shear zones is their curviplanar geometry, encompassing and wrapping around more rigid, less deformed rock bodies (shear lenses).

Four general types of shear zones can be subdivided, based on their dominant type of deformation (e.g. Ramsay & Huber 1987, Davis & Reynolds 1996). A ductile shear zone displays structures that have a metamorphic aspect and record shearing by ductile flow. The deformation processes within ductile shear zones is mainly achieved by crystal plasticity and thus comprises only a minor amount of fracturing. The rocks formed in ductile shear zones are called mylonites (e.g. Passchier & Trouw 1996). A brittle shear zone, generally referred to as a fault zone, contains fractures and other features formed by brittle deformation mechanisms. Displacement is taken up on a network of closely spaced fractures. Due to the high permeability of the fractured material, hydrothermal inflow is promoted and faulting is often accompanied by hydrothermal activity (e.g. Hulin 1929, Higgins 1971, Sibson 1977, Wise et al. 1984). Semibrittle shear zones include en échelon veins or joints and stylolites and involve mechanisms such as cataclastic flow and pressure solution. Brittle-ductile shear zones show evidence for both brittle and ductile deformation.

3.2.2. Characterization of the Shear Zones in the Eastern Aar Massif

3.2.2.1. Geological Setting

The Aar massif is situated in the central Swiss Alps (Fig. 3.1). It outcrops in the form of a 115 km long and 23 km wide NE striking mountain range. The study area is located in the eastern part of the massif (Fig. 3.1).

The Aar massif is built up by magmatic rocks (dominantly granites) enclosed in a crystalline basement (primarily gneisses) (Labhart 1977, Abrecht 1994) (Fig. 3.1). Rocks and structures of the Aar massif are characterized by an overprint of the Tertiary Alpine collision during which the rocks were affected by heterogeneous ductile deformation and a metamorphic overprint under at most medium greenschist facies/biotite zone conditions (Labhart 1977, Steck 1984, Choukroune & Gapais 1983, Frey et al. 1980). The deformation was localized on all scales, from mm to km, in
ductile and brittle-ductile shear zones (Choukroune & Gapais 1983, Meyer et al. 1989). During subsequent uplift of the Aar massif, the area was affected by brittle deformation which generated shear fractures and joint systems, locally filled with hydrothermal crystallizations (Steck 1968a, Meyer et al. 1989, Bossart & Mazurek 1990). Due to the weakness of the existing shear zones, most of the brittle fractures concentrated within mylonites. Uplift of the Aar massif is still going on (e.g. Pavoni 1979, Gubler et al. 1981, Kahle et al. 1997, Kohl et al. 2000). It is assumed that the quaternary movements are bound to isolated shear zones (Jäckli 1951, Eckardt 1957, Eckardt et al. 1983, Fischer 1990).

3.2.2.2. Structural Composition

The majority of shear zones in the eastern Aar massif are about 2 to 15 m in width, strike ENE-WSW and dip steeply to the S or N (Laws et al. 2001). A minor shear zone set strikes NW-SE and dips steeply to the SW or NE.

The deformation structures of the shear zones in the eastern Aar massif are primarily ductile (Laws et al. 2001). A characteristic progressive evolution from the host rock to mylonite (rocks with 50 to 90 % matrix as compared to the porphyroclast content) and locally ultramylonite (rocks with more than 90 % matrix) is documented by the formation and intensification of a foliation and a stretching lineation and the progressive replacement of the host rock by a finer-grained material containing newly crystallized and recrystallized minerals. Rock compositions simultaneously show an increase in the mica content. Towards the mylonitic and ultramylonitic zones, the network of small-scale shear zones anastomosing around more competent, lozenge-shaped lenses increases in frequency.

The shear zones additionally show a higher fracture density than the host rock (Laws et al. 2001). The fractures oriented parallel and oblique to the foliation can be open but are mainly filled with a μm-thick, often cohesionless, fine-grained, structureless arrangement of micas and quartz. Microfractures represent trans- and intragranular fractures (Fig. 3.2). Higher and highest fracture densities are restricted to fine-grained and strongly foliated sections of the shear zone. Along with fracture density, the thickness of the fracture infill generally increases towards areas with highest fracture densities, rendering the mylonite more and more friable. In heavily fractured zones, fracturing destroyed the foliation. Within these zones, structureless and cohesionless microbreccias occur. The microbreccias consist of mylonite fragments separated by a mm-wide, fine-grained matrix. The matrix is composed of the same material as the fracture infilling and is therefore assumed to represent this infill.

3.2.2.3. Structural Model

Considering a number of more than 100 shear zones examined, based on fracture density and mica content, the shear zones in the eastern Aar massif can be divided into 3 different textural subzones (Laws et al. 2001). With increasing fracture density and mica content, these include a strongly foliated zone, a fractured zone and a heavily fractured,
cohesionless zone (Fig. 3.3). In general the strongly foliated, the fractured and the cohesionless zones may be visualized as having interval thicknesses of 50 %, 40 % and 10 % along the total shear zone width, respectively. However, due to rheological variations, the thickness of these zones may vary along both their strike and dip. The zones are oriented approximately parallel to the rock mass foliation and often form a symmetric sequence centered by the heavily fractured, cohesionless zone. The cohesionless zone may also anastomose within the fractured zone, or it is absent.

The strongly foliated zone may have a thickness of up to several meters. The zone is characterized by mylonites or ultramylonites slightly more fractured than the host rock.

The strongly foliated zone gradually progresses into the fractured zone. The fractured zone may be a few meters thick. Like the strongly foliated zone, the fractured zone is composed of mylonite or ultramylonite, but these rocks are intensely fractured. The fractures may contain infills with thicknesses of up to 1 mm. The increased thickness of fracture infill is also responsible for an increased amount of mica in the rock material. Due to the high degree of fragmentation in the fractured zone and the thickness of the fracture infill, the rock material in this zone is extremely friable. Extremely strong fragmentation towards the heavily fractured, cohesionless zone may result in the disruption of the rock fabric. Under these circumstances a structureless microbreccia is formed consisting of about 1 cm large rock fragments of mylonite separated by a mm-wide fine-grained matrix. Microbreccias classified into the fractured zone have some cohesion which may have formed secondary.

The cohesionless zone has a distinct and sharp contact with the surrounding cohesive material. The thickness of the zone varies from a few centimeters to several decimeters. The cohesionless zone consists of structureless microbreccias that can be distinguished from those of the fractured zone by the increased amount of matrix material and by their cohesionless nature. The rock material in this zone may, however, show some signs of cohesion due to the fine-grained nature of the matrix minerals. Swelling clay minerals were not observed.

3.2.2.4. Examples

Figures 3.4 and 3.5 show examples of the distribution of fracture density, thickness of fracture infill, mica content and foliation intensity across shear zones in the eastern Aar massif.

It has to be noted that for the diagram illustrating the distribution of fracture density, only macroscopically visible fractures were taken into account. Indeed, for a representative statement about fracture density and the illustration of friability of the rock material, also microfractures have to be considered. In the shear zones of the eastern Aar massif, a high number of short, unconnected microfractures especially occurs in the fine-grained material like the ultramylonite.
In this respect, although this is only visible in the diagram for the shear zone in granite, if macroscopically visible fractures and microfractures are considered, fracture density within the rock material of both shear zones presented is high in the fractured zone and further increases towards the heavily fractured, cohesionless zone. However, within the shear zone in gneiss fracture density increases less regular than in the shear zone in granite. The rock material of the shear zone in gneiss already shows a high number of macroscopically visible fractures at the transition from the strongly foliated to the fractured zone, then this number decreases and is only increased again within the heavily fractured, cohesionless zone. Microfracture density is low at the transition from the strongly foliated to the fractured zone but then constantly increases towards the heavily fractured, cohesionless zone. Together, macroscopically visible fractures and microfractures result in a high fracture density at the transition from strongly foliated to fractured zone, then the density decreases and close to the transition to the cohesionless zone fracture density is high again.

3.3. Triaxial Testing

To determine the geomechanical properties of the different shear zone subzones described in the preceding section, a series of laboratory triaxial tests was performed on shear zone samples taken from the eastern Aar massif. The tests were performed at the rock mechanics laboratory (LMR) of the EPFL, Lausanne.

3.3.1. Sampling and Sample Preparation

Samples were obtained from two borehole cores. The boreholes were drilled across two different shear zones intersecting the safety tunnel of the Gotthard highway tunnel (Figs. 3.6, 3.7). One borehole was drilled through a shear zone in Aar granite (borehole la), the second was drilled through a shear zone in gneiss (borehole Ha). Both shear zones had been sites of overbreaks during tunnel construction. The drill sites had an overburden of 400 and 500 m, respectively.

The boreholes were oriented normal to the strike and dip of the shear zones and the rock mass foliation. To keep disturbance of the highly fractured shear zone material at a minimum, and due to the heterogeneous nature of the rock material and its expected control on laboratory test results, relatively large diameter cores were drilled using a 116 mm diameter core bit. The resulting cores had a diameter of 90 mm. To further protect the rock material, triple-tube drilling flushed with a minimal amount of water was performed. Samples were sealed and stored in the plastic tubing.

Samples obtained by borehole drilling and then tested are described in Table 3.1 and illustrated in Figures 3.10 and 3.11. For testing purposes, one sample was taken from each shear zone subzone for both the granite and the gneiss hosted shear zones. In addition, because fracture density in the fractured zone strongly increases, several samples were taken from this zone. The position of the samples within the shear zones
from which they were taken is shown in Figures 3.4 and 3.5. All samples had their axis oriented normal to the foliation.

Due to the highly fragmented nature of some samples, sample disturbance had to be controlled by leaving the samples in their PVC liner during end preparation. Sample ends were cut through the plastic tubing, considering that they are flat and perpendicular to the core axis. Samples were removed from the plastic tubing only after end preparation and were then supported with adhesive tape and a calcite-like paste (HILTI MD 2000) where necessary. With the exception of samples GN5a-c, which were drilled to smaller diameters of 85 and 55 mm, the original 90 mm diameter of the cored samples was maintained. All diameters satisfied ISRM standards with respect to having a diameter 10 times greater than the maximum grain size or fissure spacing (ISRM Commission 1983). The height of the specimen, according to ISRM recommendations, was approximately two times the sample diameter.

3.3.2. Test Setting, Load Path and Parameters Calculated

The triaxial tests were performed as consolidated, drained compression tests (CD tests). Consolidated, drained tests were carried out to minimize the effect of excess pore pressures, as samples were tested with their natural water contents to prevent any disturbance to the rock’s structure associated with drying. Tests were performed according to ISRM recommendations (ISRM Commission 1983). The loading system consisted of a servo-controlled, stiff testing machine with a maximum load capacity of 200 t. Both, LEGEP and HOEK type triaxial cells were used. Confining pressures were provided by means of hydraulic oil pumps.

Axial loads were applied at a constant displacement rate of 0.1 mm/min, except for samples GN5a-c which where loaded at a rate of 0.2 mm/min. The confining pressure for the tests was set at 5 MPa, with the exception of specimen GN3b where a confining pressure of 1 MPa was used. Loading proceeded in this fashion up to the peak strength after which a form of controlled multiple failure state loading was employed to increase confinement through several stages of residual deformation and strength (see Kováři & Tisa 1975). In contrast to Kováři & Tisa’s (1975) multiple failure state triaxial tests, increases in confining pressure to maintain different limit equilibrium states were interpreted as not being those for peak strength conditions but instead, residual strength conditions. Accordingly, it was assumed that at peak strength a major change to the rock’s fabric occurs through the formation of a failure plane and that any subsequent conditions can only be applied to that of a residual strength state. Confining pressures were increased and held constant two times in succession from 5 to 10 MPa and 10 to 15 or 20 MPa until residual strengths at those intervals were reached (Fig. 3.8).

Axial strain ($\varepsilon_{\text{axial}}$) was measured using a linear displacement transducer and volumetric strain ($\varepsilon_{v}$) was established through volume changes to the confining pressure fluid in the triaxial cell. From these parameters, the radial strain ($\varepsilon_{\text{radial}}$) was determined assuming axial symmetry of a cylindrical volume:
\[ \varepsilon_{\text{radial}} = 0.5 (\varepsilon_r - \varepsilon_{\text{axial}}). \]

The validity of this assumption was found to be somewhat limited given the inhomogeneous nature of the deformations observed in most of the tested specimens.

Approximations of the Young's modulus (E) were calculated from the axial stress-strain curves as: (1) a Tangent modulus (E_T), i.e. the slope of the axial stress-strain curve at 50 % of the peak strength, and (2) a Secant modulus (E_S), taken as the slope of a straight line joining the origin of the axial stress-strain curve to the point at 50 % of the peak strength. Values of Poisson's ratio (v) were determined by dividing E_T through the slope of the \( \sigma_1 - \varepsilon_{\text{radial}} \) curve at 50 % of peak strength. Calculations of the elastic constants were restricted to assumptions of an isotropy due to limitations in sampled core orientations. Preferably, a transverse anisotropic model would have been more appropriate for the strongly foliated nature of the shear zone rocks. As loading was performed normal to the foliation, values determined for Young's modulus and Poisson's ratio may, in fact, only be considered valid for a similar direction.

Through the testing procedure and load paths applied, a Coulomb residual strength envelope could be obtained for each sample tested, making it possible to establish the residual cohesion (c_r) and residual internal friction angle (\( \phi_r \)). The parameters were calculated with the program ROCKDATA (Shah & Hoek 1991a). Assuming a linear Coulomb envelope, the program was used to derive values of the residual cohesion and friction angle by means of a linear regression fitting routine. The Coulomb criterion implies that a continuous, smooth shear plane develops at peak strength and that the peak strength envelope is linear, which is a simplification at best for most rock materials. However, Brady & Brown (1993) suggest that the Coulomb criterion does provide a relatively good representation of the residual strength conditions. The residual strength, defining the load bearing characteristics of the rock subsequent to the full development of a shear plane, should, if compared to the shear strength examined in direct shear tests, increase with shear plane surface roughness or with the development of multiple shear planes (e.g. Brady & Brown 1993). This is of practical importance in the context of the highly fractured rock types tested in this study. Accordingly, values determined on basis of the Coulomb strength criterion were treated as representing the lower limit of the residual strength of the rocks tested.

### 3.4. Experimental Results for Granite Hosted Shear Zone Samples

Axial stress-strain curves resulting from the triaxial compression tests are provided in Figure 3.9. Results from the analysis are summarized in Table 3.2.

The axial stress-strain curves obtained for the granite-based samples clearly flatten and widen with the samples approaching the cohesionless zone, i.e. with increasing degree of tectonic overprint (increasing mica content and foliation and...
fragmentation intensity) (Fig. 3.9a). The curves thus demonstrate an increase in nonlinearity upon initial loading, representing crack closure, and an increase in nonlinearity leading to yielding and peak stress, implying the occurrence of an increased amount of plastic deformation prior to attaining peak stress. Apart from this, a decrease in elastic stiffness with increasing degree of tectonic overprint is demonstrated. Values of Young's modulus were seen to decrease across each shear zone subzone (Fig. 3.16), beginning with the strongly foliated zone neighbouring the intact host rock. Compared to 45 GPa reported by Schneider (1992) for intact Aar granite, Young's moduli values of 17 GPa were tested for sample GR1 in this study. The ratios of axial to lateral strain, i.e. Poisson's ratio, show no clear trend (Fig. 3.16). However, an increase of the Poisson's ratio with less intact rock material can be suspected. Results for the samples from the cohesionless zone show Poisson's ratio values of approximately 0.5, which is the upper limit for Poisson's ratio indicating that the material did not change its volume (e.g. Prinz, 1997). A relatively high Poisson's ratio would also indicate a high amount of plastic deformation contributing to the overall strain. Given the increasing degree of plastic deformation contributing to the overall strain, it is questionable whether the use of elastic parameters like E and v to describe the deformability characteristics of this rock material is applicable.

With increasing tectonic overprint peak strength values also decrease considerably (Figs. 3.9a, 3.16, Tab. 3.2). Assuming a uniaxial compressive strength for an intact, homogeneous Aar granite to be 176 MPa (Schneider 1992), a 38 % reduction in peak strength values already exists between the host rock and the strongly foliated zone.

The post-peak drop in stress to the residual strength likewise decreases with increasing tectonic overprint (Fig. 3.9a). For example, in the case of GR1, representing the least overprinted material, the peak strength drops about 86 % to the residual strength value. Post-peak drops of 82 %, 74 %, and 13 % were observed for samples GR2, GR3 and GR4, respectively. Sample GR5 shows strain hardening. The decreasing drops in peak stress to residual strength values imply a transition from typically brittle (samples GR1 and GR2) to fully ductile rock behaviour (sample GR5) with increasing tectonic overprint.

The transition from brittle to ductile rock behaviour and thus the increasing amount of plastic deformation contributing to the overall strain can likewise be related to observations of the failure mode of the different samples (Fig. 3.10). Samples GR1 and GR2 failed in a typically brittle manner, i.e. by the development of a single failure plane with a 30° angle towards $\sigma_{\text{axial}}$ (Fig. 3.10, Tab. 3.2). The failure plane of sample GR1, which showed no macroscopically visible fracture, would appear to have developed through the interconnection of stress-induced microfractures and foliation planes. Sample GR2 mainly failed through the propagation of a pre-existing macroscopically visible, filled fracture oriented oblique to the specimen axis. Thin sections of the failed sample show that the development of the failure plane also involved propagation along foliation planes. More highly tectonically overprinted samples showed an increase in the occurrence of plastic deformation processes through sample barrelling prior to failure. In addition, the formation of a distinct, brittle failure
plane also became less prominent. For example, sample GR3 having a stress-strain curve showing a clear transition from brittle to ductile rock behaviour failed by barrelling and shearing along several failure planes enclosing an area of crushed and intact rock material (Fig. 3.10). Shear planes at angles of 40° towards $\sigma_{axial}$ can be discerned (Fig. 3.10, Tab. 3.2). Thin sections show that the multiple failure planes developed by the interconnection of pre-existing microfractures and foliation planes which all were present in the sample in high numbers (Fig. 3.12). Samples GR4 and GR5 showing an extremely high fracture density failed by desaggregation into cm- to mm-large rock-fragments. It can only be assumed for these samples showing ductile behaviour that failure occurred through large deformations in the form of barrelling over the entire volume of the sample. Sample GR4 may have failed along the numerous pre-existing microfractures. Sample GR5 appears to have failed along the macroscopically visible fractures filled with a cohesionless material forming the matrix of the rock.

Mohr-Coulomb analysis of the residual cohesion shows that values systematically increase with increasing tectonic overprint (Figs. 3.15a, 3.16, Tab. 3.2). Values for the residual angle of internal friction also increase with increasing tectonic overprint (Figs. 3.15a, 3.16, Tab. 3.2). However, this increase is not as well defined as the increase in values for the residual cohesion. The increases in residual cohesion and residual angle of internal friction with increasing degree of tectonic overprint can be associated with an increase in residual strength, which is formed by cohesion and frictional resistance along the failure plane. This increase in residual strength suggests that an increase in load carrying ability following failure exists relative to the intact host rock with increasing degree of tectonic overprint. The trend of the residual values seems to be related to the different failure mechanisms of the samples. Observations of the tested samples suggest that the residual values increase with the number of failure planes enclosing an area of crushed and intact rock material. In contrast to the single, smooth failure plane seen in sample GR1, for which virtually no residual cohesion exists, such failure zones consisting of a network of coalescing fractures may act to increase residual cohesion with increasing thickness of the zone. The failure zones also represent a higher frictional resistance to shear than a single, smooth failure plane would.

3.5. Experimental Results for Gneiss Hosted Shear Zone Samples

The axial stress-strain curves for the shear zone samples from the gneiss host rock show a less regular pattern than those taken from the shear zone hosted in granite (Figs. 3.9b, 3.16, Tab. 3.2). However, similar trends were observed in the stress-strain curves, the exception being sample GN2 which proved to have significantly lower strengths than all other samples tested. Also in the stress-strain curves for the shear zone in granite, the pattern of continuously decreasing values of Young's moduli and peak strength, a decrease in the post-peak drop in stress to the residual strength and increasing values of Poisson's ratio demonstrate a transition from brittle to ductile rock behaviour with increasing degree of tectonic overprint. When compared to the uniaxial
The compressive strength of 167 MPa and the Young’s modulus values of up to 38 GPa for intact biotite gneiss with a loading direction perpendicular to foliation (Schneider 1992), decreases of 70% and 80%, respectively, can already be observed at the transition from intact rock to the strongly foliated zone of the shear zone. A generally increasing trend in Poisson’s ratio values is not very clear. Similar to the granite-hosted shear zone samples, it is questionable whether the elastic parameters E and ν are applicable for describing the deformability of rock material exhibiting a high amount of plastic deformation prior to failure. Peak strength drops about 16%, 4%, 25% and 11% to the residual strength values of samples GN1, GN2, GN3b and GN4, respectively. Samples GN5a-c show strain hardening.

The transition from brittle to ductile rock behaviour was also documented in the failure mode of the samples (Fig. 3.11). However, none of the samples from the shear zone in gneiss appears to have failed along a single failure plane. Sample GN1 failed along a failure zone oriented at an angle of about 60° towards σaxial (Fig. 3.11, Tab. 3.2). Observations in thin sections revealed that the failure planes developed through the coalescence of pre-existing microfractures and foliation planes. Samples GN3b and GN4 also failed along failure zones which increase in thickness with increasing degree of tectonic overprint. Failure of sample GN3b occurred through the coalescence of pre-existing, macroscopically visible fractures oriented obliquely to the sample axis. The failure zone of sample GN4 formed through the propagation and coalescence of pre-existing, filled fractures oriented obliquely to the specimen axis and of foliation planes, as it can be seen in thin sections (Figs. 3.13, 3.14). Failure of the sample also involved barrelling of a part of the sample, i.e. the occurrence of a high amount of plastic deformation (Fig. 3.11). Figure 3.13 illustrates examples of failed rock material taken from the barreled sample area. In comparison to the rock material illustrated in Figure 3.14, the material from this area contains a higher amount of fine-grained mica, a higher number of foliation planes and a higher number of fractures. The fractures are mainly short and have extremely rough surfaces. Samples GN2 and GN5a failed by desaggregation into cm-large rock fragments. Desaggregation occurred along pre-existing fractures.

The failure modes again correlate well with the values of residual cohesion and residual angle of internal friction, which seem to increase with increasing degree of tectonic overprint (Figs. 3.15b, 3.16, Tab. 3.2). However, this trend was not necessarily conclusive for all samples tested given the relatively small data set involved.

In comparison to the granite based material, the material from the shear zone in gneiss generally seems to be weaker and shows less brittle rock behaviour. For example, sample GN1, comparable to sample GR1 in terms of structural shear zone characterization (i.e. representing the strongly foliated zone), shows an axial stress-strain curve comparable to that of sample GR2. Peak strength of sample GN1 is even lower than that of sample GR2. Furthermore, the post-peak drop in stress of sample GN1 is more flat than that of sample GR2. Also the failure mode of sample GN1 suggests the involvement of more plastic deformation mechanisms than that for sample GR2.
3.6. Comparison of Experimental Results for Fault Zone Materials

Results of drained triaxial tests performed on faulted rock materials have been similarly presented by Chester & Logan (1986), Habimana et al. (1998) and the AlpTransit Gotthard AG (1999)/Ehrbar & Pfenniger (1999). These laboratory studies primarily differ from one another in their host rock lithology, however similarities do allow for generalized comparisons to be made with fracture density and its influence on test results.

In this respect, Chester & Logan (1986) tested sandstone samples from the Punchbowl fault zone in California and defined samples as being either undisturbed, containing single and/or multiple subsidiary faults or consisting of fault gouge material. The fault gouge represented the material of highest fracture density and clay content. Habimana et al. (1998) tested rock material from a fault zone developed in quartzitic sandstone. The samples were differentiated by the degree of folding, faulting and jointing phenomena (i.e. "tectonisation"). Heavily tectonised samples represented soil-like material. Faulted gneisses, schists and phyllites from the Tavetsch massif (Fig. 3.1) were tested by the AlpTransit Gotthard AG (1999). The tested material was divided into slightly faulted rocks ("kakiritisated" rocks), intensely faulted, fine-grained rocks with low cohesion ("regional Kakirites") and extremely heavily faulted, very fine-grained, clay-rich, structureless and cohesionless rocks ("zonal Kakirites").

A comparison of Young's moduli established in these testing campaigns and scaled with respect to fracture density is presented in Table 3.3. The decrease in values for the Young's modulus seen in Table 3.3., illustrates that there is a general decrease in stiffness with increasing fracture density. Chester & Logan (1986) additionally describe a gradual decrease in strength and an increase in ductility of rock behaviour with increasing fracture density and clay content. A decrease of rock strength and an increase in plastic yielding with increasing degree of tectonisation is also reported by Habimana et al. (1998). Apart from a decrease in values for Young's modulus, the AlpTransit Gotthard AG (1999) and Ehrbar & Pfenniger (1999) show increasing values of Poisson's ratio and a reduction in post-peak drop in stress to the residual strength with increasing fracture density. The results from the studies concur with those established for the eastern Aar massif shear zone rocks.

3.7. Discussion

3.7.1. Strength and Deformability Characteristics of the Shear Zone Samples

In summary, the tests on the shear zone samples demonstrate that an increasing degree of tectonic overprint corresponds with: an increase in stress-strain nonlinearity upon initial loading and in the area of yield and peak strength, a decrease in elastic stiffness, a decrease in peak strength and an increase in residual strength occurring in
the axial stress-strain curves of both the granite hosted shear zone samples and the
gneiss hosted shear zone samples. Failure modes show a transition from failure along a
single plane to failure along a zone. It may be assumed that the failure zone increases in
thickness with increasing degree of tectonic overprint. The trends may be explained by
the increasing densities of pre-existing discontinuities, especially of such that are
inclined at angles from 25 to 40° towards $\sigma_{\text{axial}}$, the increasing thickness of fracture infill
and the increase in the mica content visible in the tested samples with increasing degree
of tectonic overprint (Tab. 3.1, Figs. 3.4, 3.5, 3.10, 3.11). Pre-existing discontinuities
are represented by fractures and foliation planes. However, as the foliation planes were
oriented normal to $\sigma_{\text{axial}}$ during the tests, their influence should have been smaller. The
assumed influence of fracture density on rock behaviour may be confirmed by the fact
that the tested faulted rock samples, mainly differing from each other in their fracture
density, show similar trends in rock behaviour than the shear zone samples. The trends
in rock behaviour of the faulted material were explained by Chester & Logan (1986)
with the difference between the gouge and the sandstone in the number of pre-existing
defformational features and clay content, i.e. fragment-size.

As stress-strain nonlinearity during the initial stages of loading represents crack
closure, it is obvious that an increase in pre-existing discontinuities, i.e. fractures, would
result in an increase in plastic to elastic strain ratios upon initial loading. This increase is
especially visible in the axial stress-strain curves of the samples GR1 to GR3 exhibiting
a continuously increasing density of pre-existing macroscopically visible fractures and
microfractures and clearly showing a continuously increasing nonlinearity upon initial
loading.

After closure, the pre-existing fractures and the foliation planes may be viewed as
planes of weakness and stress concentrators along which new, stress-induced fractures
may initiate, propagate and coalesce relatively easy, especially if they are inclined at
angles greater than 25 to 40° towards $\sigma_{\text{axial}}$. An increasing number of pre-existing
discontinuities should therefore lead, when critically stressed, to an increasing number
of actively propagating fractures widely distributed throughout the sample. This
increasing number of active fractures would coincide with an increasing occurrence of
plastic deformation mechanisms, such as cataclastic flow, along the fractures. When
cohesionless fracture infill is present, the effect of cataclastic flow should even be
intensified with the intensification increasing proportionally with the thickness of the
fracture infill. The increasing content in mica, which often deforms by slip along grain
boundaries (e.g. Passchier & Trouw, 1996), should also contribute to an increase in the
amount of plastic strain recorded. In the axial stress-strain curves, this increase in plastic
deformation is denoted by a decrease in elastic stiffness, an increase in nonlinearity
leading to yielding and peak stress and an increase in residual strength. As such, a full
transition can be traced through the stress-strain curves from typically brittle to fully
ductile rock behaviour. An example of the influence of increasing foliation intensity and
increasing mica content on the amount of plastic deformation is given by sample GN4,
where areas with a higher number of foliation planes, in contrast to sample areas with a
smaller number of foliation planes, show barrelling prior to failure. Sample GR5
provides an example of the influence of increasing thickness of cohesionless fracture
infill on the amount of plastic deformation. In this sample, rock fragments appear to float in a fine-grained, cohesionless mica-matrix representing fracture infill. The mechanical behaviour of the sample should therefore be largely controlled by the response of this matrix. In contrast to sample GR4 for which the fracture density is the same but infills are mainly cohesive, sample GR5 reacts in a fully ductile manner when loaded.

Fracture initiation, propagation and coalescence originating along pre-existing fractures and foliation planes may also explain the observed continuous decrease in strength with increasing degree of tectonic overprint, i.e. fracture density and foliation intensity. The effect of increasing microfracture densities acting to promote fracture initiation, propagation and coalescence, on uniaxial compressive strengths has, for example, been shown previously by Eberhardt et al. (1999) for granite samples.

An increasing number of coalescing discontinuities would also act to promote the observed change from failure along a single failure plane to failure along a failure zone. The associated increase in the amount of plastic deformation contributing to the overall strain manifests itself in form of sample barrelling. As previously explained, the change in failure mode towards failure along a zone is responsible for a reduction in post-peak drop in stress to residual strength and an increase in the residual strength itself.

According to these findings, lower strength and elastic moduli values seen in gneiss-based samples relative to those of corresponding granite-based samples, may be explained by higher fracture densities, foliation intensities and mica contents seen in the gneiss-based shear zone samples (Tab. 3.1). Sample GN2 shows a higher density in long, macroscopically visible fractures than all other gneiss based samples (Fig. 3.11) resulting in a non-linear rock behaviour and a relatively low yield stress. The lower peak strength of sample GN3b relative to GN4 may be explained by the lower confining pressure at which sample GN3b was tested (Tab. 3.2).

3.7.2. Strength and Deformability Characteristics of the Sheared Rock Mass

Laboratory observations suggest that it may be possible to assign each shear zone subzone a typical rock mass constitutive behaviour and a corresponding failure mode. For example, with low confining pressures, the host rock and the strongly foliated zone would behave in a brittle fashion and would tend to fail through the development of a single, discrete shear plane. Accordingly these rock would have a relatively high peak strength and low residual strength. The fractured zone represents a transition from brittle to ductile behaviour and therefore tends to fail along a failure zone as opposed to a single, discrete plane. As a result, the peak strength of these rocks would be generally lower than those of the strongly foliated zone but the rock may carry a higher post-peak load after failure. Material from the heavily fractured, cohesionless zone behaves almost entirely in a fully ductile manner and fails through large deformations in the form of barrelling over the entire volume of the sample. Rock strength generally increases with increasing consolidation (i.e. strain hardening).
However, as the samples only represent parts of the sheared rock mass, laboratory testing may only provide an approximation of the large scale rock mass properties. For example, the strongly foliated zone, having a relatively low fracture density, contains widely-spaced, discrete, persistent fractures/joints locally controlling rock mass behaviour. As the tested samples of this zone did not contain such fractures, they may not be representative of the rock mass of this zone. Towards the cohesionless zone, as fracture density increases, the influence of persistent, discrete fractures locally controlling rock mass behaviour decreases. The heavily fractured area of the fractured zone and the cohesionless zone finally contain such a high fracture density that rock pieces between fractures are small compared to the overall size of the shear zone subzone. These areas of the shear zone can therefore be regarded as isotropic. In samples GR4, GN4, GR5 and GN5a-c the size of individual rock pieces is also small with respect to the overall size of the sample and the samples contain a certain amount of inhomogeneities characteristically for the shear zone subzones, especially fractures. Samples GR5 and GN5a-c even almost contain the cohesionless zone in its complete thickness. Samples GR4, GN4, GR5 and GN5a-c therefore may be regarded as isotropic and representative for the rock mass of the corresponding shear zone areas.

### 3.7.3. Hoek-Brown Estimation of the Strength of the Sheared Rock Mass

To derive more information with respect to the large scale rock mass properties and their variance across the shear zones the samples originated from, the Hoek-Brown method (Hoek & Brown 1997, Hoek et al. 1998, Hoek 1999a) was used for an estimation of the overall strength of the fractured shear zone rock masses. The Hoek-Brown method appeared to be most suitable for this purpose, as the method focuses on the role of pre-existing discontinuities, joint surface roughness and fracture infill as the controlling influence on rock mass strength. As described previously, these factors also have a controlling influence on the laboratory determined peak strengths. However, the method does not account for the direction of loading with respect to the foliation, i.e. for the anisotropy created by the foliation of the sheared rock mass. The values determined in the laboratory are valid only for a loading direction perpendicular to the foliation.

Input values used for the estimation and subsequent results are given in Table 3.4. The uniaxial compressive strength (σ_c) of the intact rock pieces in the rock mass and the value of the Hoek-Brown constant (m_i) for these intact rock pieces were based on those obtained for the strongly foliated zone. In this respect, σ_c was calculated from the peak strengths established for samples GR1 and GN1 by a transformation of these values into equivalent values for a 50 mm core (Hoek & Brown 1997). M_i was estimated from the appropriate tables provided by Hoek & Brown (1997) and Hoek (1999a). The different shear zone subzones were characterized using the Geological Strength Index (GSI) (Hoek et al. 1998, Hoek 1999a). For the fractured zone several different rock strength values were estimated, as fracture density in this zone strongly increases. Given that the majority of fracture surfaces observed were not smooth (Fig. 3.2b) and infills were relatively thin and compact, surface conditions of the fractures were generally regarded as good.
The resulting estimated strengths of the rock mass reflect the variations of the peak strengths determined in the laboratory, showing decreases with increasing fracture density. However, the estimated rock mass strength of the strongly foliated zone of the shear zone in granite was found to be less than half the laboratory value. This discrepancy between rock mass values and laboratory determined values decreases towards the cohesionless zone, where strengths are approximately the same. Rock strengths of the shear zone in gneiss show a similar trend. It should be noted that the difference of 50% between laboratory determined strength and those for the rock mass are in accordance with general assumptions used in the mining and tunnelling industries. The simplified analysis confirms the assumption that as test samples approach the cohesionless zone, they become more and more representative of rock mass behaviour (i.e. as fracture density increases and the influence of discrete, persistent discontinuities on rock mass strength decreases).

3.7.4. Implications of the Shear Zones on Tunnel Stability

Once consideration has been given to the micromechanical and rock mass scale strength and deformability characteristics of the shear zones in the eastern Aar massif, practical implications can be drawn with respect to these shear zones and their effect on tunnel stability.

The high density of foliation planes and fractures parallel and oblique to the foliation, particularly within the fractured and the cohesionless parts of the shear zone, implies that in low stress environments in the area of the shear zones an increased number of ground control problems in the form of unravelling of interlocking blocks and wedges (i.e. structure controlled instability) may be expected. The size of the blocks and wedges should generally decrease as foliation intensity and fracture density increases. In heavily fractured rock masses, failure would occur as a progressive process starting with the unravelling of smaller wedges exposed at the excavation surface and gradually work its way back into the rock mass (Hoek et al. 1995). Such failures which may result in high chimneys were, as mentioned earlier, in fact observed during the construction of the Gotthard highway tunnel in areas where the fractured and the cohesionless shear zones subzones were intersected (e.g. Keller et al. 1987).

In high stress environments, the low strength and high deformability characteristics of the shear zones examined would suggest that the sheared rock mass would predominantly fail by sliding along discontinuities and crushing of intact rock pieces (i.e. stress controlled instability). In the more heavily fractured central parts of the shear zone showing ductile rock behaviour, this type of failure may result in floor heave and sidewall closure (i.e. squeezing). High pore water pressures which may occur in these central parts, may increase the magnitude of the resulting displacements. Like in the case of the structure controlled instabilities, failure would occur as a progressive process (Hoek et al. 1995).

After, for example, Deere (1973) reporting about engineering problems associated with "foliation shear zone" structures, seriousness of engineering problems increases
with the length over which the zones are encountered. The "foliation shear zones" are described as having similar structural and geomechanical properties as heavily fractured areas within the shear zones of the eastern Aar massif. In this respect it can be explained why the shear zones in the eastern Aar massif, having fractured and cohesionless zones being relatively small in thickness, never caused any serious stability problems during previously undertaken tunnelling projects. The shear zones were also almost always intersected nearly perpendicular to the tunnel axis.

3.8. Conclusions

Results from this study show:

(1) Rock strength and deformability characteristics of the rock material within shear zones in the eastern Aar massif are related to their structural characteristics. The geomechanical properties are mainly influenced by fracture density, foliation intensity, thickness of fine-grained fracture infill and mica content of the material.

(2) As fracture density increases and the influence of discrete, persistent discontinuities on rock mass strength decreases, behaviour of the test samples becomes more and more representative of rock mass behaviour. Samples with rock pieces between fractures being small compared to the overall size of the sample may be regarded as isotropic and representative for the rock mass.

(3) With increasing fracture density and mica content, the shear zones in the eastern Aar massif can be divided into a strongly foliated zone, a fractured zone and a heavily fractured, cohesionless zone. In this respect, it is possible to assign each of these shear zone subzones a typical rock mass constitutive behaviour and a corresponding failure mode. The strongly foliated zone, similar to the neighbouring host rock, behaves in a brittle fashion, failing along a single, discrete shear plane and having a relatively high peak strength and low residual strength. The fractured zone represents a transition from brittle to ductile behaviour and therefore tends to fail along a failure zone as opposed to a plane. As a result, the peak strength is generally lower but the residual strength is higher than that of the strongly foliated zone. Material from the heavily fractured, cohesionless zone behaves in a fully ductile manner and fails through large deformations in the form of barrelling over the whole volume of the sample. Rock strength increases with increasing consolidation (strain hardening).

(4) Reasons for the problematic nature of the shear zones during tunnel excavation may be given by their special structural and geomechanical characteristics. The design of adequate support measures in the area of such shear zones may therefore require the consideration of these factors and the factors should be included in numerical modeling calculations.
Fig. 3.1. Geographical and geological setting of the Aar massif (after Abrecht 1994), location of the study area and location of Gotthard Base Tunnel and Gotthard highway tunnel.
Fig. 3.2. a) Transgranular, open fractures intersecting quartz (qz) and feldspar minerals (plag, kfsp) and finer-grained foliation domains. Sample results from a shear zone hosted in granite (sample GR3, borehole Ia). b) Transgranular fracture showing an infill of fine-grained quartz, biotite, muscovite and carbonate arranged in a structureless manner. Sample results from a shear zone hosted in gneiss (sample GN3, borehole IIa). Thin sections show a cut perpendicular to foliation and parallel to stretching lineation. Crossed polars.
Fig. 3.3. Schematic illustration of the structural composition of shear zones in the eastern Aar massif.
Fig. 3.4. Fracture density, thickness of fracture infill, mica content and foliation intensity measurements along a transect crossing a shear zone in granite (borehole 1a). Figure also shows the sampling locations GR1 - GR5.
Fig. 3.5. Fracture density, thickness of fracture infill, mica content and foliation intensity measurements along a transect crossing a shear zone in gneiss (borehole IIa). Figure also shows the sampling locations GN1 - GN5.
Fig. 3.6. Location of the two drill sites. For legend see Fig. 3.1.

Fig. 3.7. Photo of triple-tube drilling normal to the strike and dip of the shear zone. Borehole was drilled only slightly inclined from horizontal. Width of view is 2.60 m.
Fig. 3.8. Example of multi-stage load path applied during triaxial testing. Example shows load path for sample GR4.
Fig. 3.9. Stress-strain curves for tested samples from a) the shear zone in granite, b) the shear zone in gneiss.
Failure along a single, irregular, rough failure plane. As intact sample showed no signs of macroscopically visible fractures, failure plane must have developed through the coalescence of stress-induced microfractures and foliation planes.

Failure along a single, irregular failure plane which is coated with a fine-grained material. In the vicinity of the failure plane the rock is crumbled and weak. Failure plane primarily developed by the propagation of a macroscopically visible, filled fracture oriented oblique to the specimen axis. The development of the failure plane also involved propagation along foliation planes.

Failure process followed several stages involving barrelling and the progressive development of several failure planes enclosing an area of crushed and intact rock material. Failure appears to have occurred through the interconnection of pre-existing, filled microfractures and foliation planes.

Failure by desaggregation of the cohesive microbreccia into various, cm- to mm-large rock fragments, no failure plane.

Failure by desaggregation of the cohesionless microbreccia into various, cm- to mm-large rock fragments, no failure plane.

Fig. 3.10. Different failure modes of the samples of the shear zone in granite (strongly foliated zone, fractured zone, cohesionless zone, foliation, macroscopically visible fracture, macroscopically visible, filled fracture). Diameter of samples is 90 mm, height is about 180 mm.
intact sample | failure mode
--- | ---
**GN1** | Sample failed along several failure planes enclosing an area of crushed rock material. Failure planes appear to have developed through the interconnection of pre-existing microfractures and foliation planes.

**GN2** | Failure by desaggregation into various, cm-large rock fragments. Desaggregation occurred along pre-existing, macroscopically visible fractures oriented normal and oblique to the specimen axis and along foliation planes.

**GN3b** | Sample failed along two, near parallel failure planes. Failure planes have irregular surfaces. Failure appears to have developed through the interconnection of pre-existing, macroscopically visible fractures oriented oblique to the specimen axis.

**GN4** | Failure involved large deformations (i.e. barrelling) and the propagation and coalescence of pre-existing, macroscopically visible, filled fractures and probably also pre-existing, filled microfractures mainly oriented oblique to the specimen axis and of foliation planes.

**GN5a** | Failure by desaggregation of the microbreccia into several, cm-large rock fragments, no failure plane.

Fig. 3.11. Different failure modes of the samples of the shear zone in gneiss (strongly foliated zone, fractured zone, cohesionless zone, foliation, macroscopically visible fracture, macroscopically visible, filled fracture). Diameter of samples GN1 to GN4 is 90 mm, height is about 180 mm. Diameter of sample GN5a is 85 mm, height is 170 mm.
Fig. 3.12. a), b) Thin section photos of failed sample GR3. Photos each show a long, transgranular, open fracture intersecting quartz-rich areas (qz) and finer-grained foliation domains. Within foliation domains the fractures are oriented almost parallel to foliation indicating that the long fractures have developed by the interconnection of foliation planes and microfractures oriented oblique to foliation. Thin sections show a cut perpendicular to foliation and parallel to stretching lineation. Crossed polars.
Fig. 3.13. a), b) Thin section photos of failed sample GN4. Pictures each show a transgranular fracture intersecting quartz-rich areas (qz) and finer-grained foliation domains. Fractures zigzag with orientations being parallel and oblique to foliation indicating that the fractures have developed by the interconnection of foliation planes and microfractures oriented oblique to foliation. Potential fracture infill has been washed out due to preparation of the thin section. Rock material was taken from the sample area showing barrelling. Thin sections show a cut perpendicular to foliation and parallel to stretching lineation. Crossed polars.
Fig. 3.14. Thin section photo of failed sample GN4. Photo shows a transgranular fracture intersecting feldspar minerals (plag), quartz-rich areas (qz) and finer-grained foliation domains. Within foliation domains the fracture is oriented almost parallel to foliation indicating that the fracture has developed by the interconnection of foliation planes and microfractures oriented oblique to foliation. Potential fracture infill has been washed out due to preparation of the thin section. Rock material was taken from a sample area showing no barrelling. Thin section shows a cut perpendicular to foliation and parallel to stretching lineation. Crossed polars.
Fig. 3.15. Coulomb residual strength envelopes for tested samples from a) the shear zone in granite, b) the shear zone in gneiss.
Fig. 3.16. Variation of Young’s modulus ($E_t$), Poisson’s ratio ($\nu$), peak strength ($\sigma_{\text{peak}}$), residual cohesion ($c_r$) and residual angle of internal friction ($\phi_r$) with respect to the structural composition of the shear zones.
<table>
<thead>
<tr>
<th>shear zone unit</th>
<th>predominant minerals</th>
<th>rock material properties</th>
<th>density of macroscopically visible fractures</th>
</tr>
</thead>
<tbody>
<tr>
<td>GR1</td>
<td>~60 vol.% qz,</td>
<td>mylonite; cm- to mm-spaced foliation, fine-grained (most grains in foliation domains have diameters &lt; 0.2 mm)</td>
<td>no visible fractures</td>
</tr>
<tr>
<td></td>
<td>~30 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~10 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GR2</td>
<td>~60 vol.% qz,</td>
<td>mylonite; mm-spaced foliation, fine-grained (most grains in foliation domains have diameters &lt; 0.15 mm)</td>
<td>cm-spaced fractures oriented mainly parallel to foliation, fractures can have an up to 1 mm thick infill</td>
</tr>
<tr>
<td></td>
<td>~20 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~20 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GR3</td>
<td>~60 vol.% qz,</td>
<td>ultramylonite; mm-spaced foliation, extremely fine-grained (most grains in foliation domains have diameters &lt; 0.1 mm)</td>
<td>cm-spaced fractures oriented mainly parallel to foliation, high density of microfractures, fractures can have an up to 1 mm thick infill</td>
</tr>
<tr>
<td></td>
<td>~20 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~20 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GR4</td>
<td>~60 vol.% qz,</td>
<td>heavily fragmented material; cohesive microbreccia</td>
<td>about 1cm-large rock fragments in fine-grained, cohesive matrix, high density of microfractures</td>
</tr>
<tr>
<td></td>
<td>~15 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~25 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GR5</td>
<td>~60 vol.% qz,</td>
<td>heavily fragmented material; cohesionless microbreccia</td>
<td>about 1cm-large rock fragments in fine-grained, mm-thick, cohesionless matrix</td>
</tr>
<tr>
<td></td>
<td>~15 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~25 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GN1</td>
<td>~60 vol.% qz,</td>
<td>mylonite; cm- to mm-spaced foliation, fine-grained (most grains in foliation domains have diameters &lt; 0.05 mm)</td>
<td>1-2 discrete fractures oriented parallel and oblique to foliation</td>
</tr>
<tr>
<td></td>
<td>~20 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~20 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GN2</td>
<td>~60 vol.% qz,</td>
<td>mylonite; cm- to mm-spaced foliation, fine-grained (most grains in foliation domains have diameters &lt; 0.05 mm)</td>
<td>cm- to mm-spaced fractures oriented parallel and oblique to foliation, fractures can have μm-thick infills</td>
</tr>
<tr>
<td></td>
<td>~10 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~30 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GN3a, b</td>
<td>~60 vol.% qz,</td>
<td>ultramylonite; mm-spaced foliation, extremely fine-grained (most grains in foliation domains have diameters &lt; 0.01 mm)</td>
<td>cm-spaced fractures oriented mainly parallel to foliation, fractures can have μm-thick infills</td>
</tr>
<tr>
<td></td>
<td>~5 vol.% fsp,</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>~35 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GN4</td>
<td>~60 vol.% qz,</td>
<td>ultramylonite; mm-spaced foliation, extremely fine-grained (most grains in foliation domains have diameters &lt; 0.01 mm)</td>
<td>cm- to mm-spaced fractures oriented parallel and oblique to foliation, high density of microfractures, fractures can have mm-thick infills</td>
</tr>
<tr>
<td></td>
<td>~40 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GN5a, b, c</td>
<td>~55 vol.% qz,</td>
<td>heavily fragmented material; cohesionless microbreccia</td>
<td>cm- to mm-large rock fragments in fine-grained, mm-thick, cohesionless matrix</td>
</tr>
<tr>
<td></td>
<td>~45 vol.% mica</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>GR1</td>
<td>GR2</td>
<td>GR3</td>
</tr>
<tr>
<td>--------</td>
<td>---------</td>
<td>---------</td>
<td>---------</td>
</tr>
<tr>
<td><strong>$E_T$ [GPa]</strong></td>
<td>17.15</td>
<td>14.49</td>
<td>1.53</td>
</tr>
<tr>
<td><strong>$E_s$ [GPa]</strong></td>
<td>12.07</td>
<td>8.07</td>
<td>1.00</td>
</tr>
<tr>
<td><strong>$\nu$</strong></td>
<td>0.33</td>
<td>--</td>
<td>0.15</td>
</tr>
<tr>
<td><strong>$\sigma_{\text{peak}} / \sigma_{\text{confining}}$ [MPa]</strong></td>
<td>109.4/5</td>
<td>82.1/5</td>
<td>50.5/5</td>
</tr>
<tr>
<td><strong>$\beta$</strong></td>
<td>30</td>
<td>30</td>
<td>40</td>
</tr>
<tr>
<td><strong>$c_r$ [MPa]</strong></td>
<td>1.21</td>
<td>2.05</td>
<td>--</td>
</tr>
<tr>
<td><strong>$\phi_r$</strong></td>
<td>27.0</td>
<td>21.5</td>
<td>--</td>
</tr>
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</table>

1 Stress at yield point

<table>
<thead>
<tr>
<th></th>
<th>GN1</th>
<th>GN2</th>
<th>GN3a</th>
<th>GN3b</th>
<th>GN4</th>
<th>GN5a</th>
<th>GN5b</th>
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<tr>
<td><strong>$E_T$ [GPa]</strong></td>
<td>6.47</td>
<td>0.05</td>
<td>0.50</td>
<td>1.32</td>
<td>0.63</td>
<td>0.24</td>
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<tr>
<td><strong>$E_s$ [GPa]</strong></td>
<td>4.63</td>
<td>0.02</td>
<td>0.46</td>
<td>1.42</td>
<td>0.65</td>
<td>0.27</td>
<td>0.30</td>
<td>0.66</td>
</tr>
<tr>
<td><strong>$\nu$</strong></td>
<td>0.24</td>
<td>0.43</td>
<td>0.23</td>
<td>0.34</td>
<td>--</td>
<td>0.35</td>
<td>0.38</td>
<td>0.25</td>
</tr>
<tr>
<td><strong>$\sigma_{\text{peak}} / \sigma_{\text{confining}}$ [MPa]</strong></td>
<td>53.9/5</td>
<td>10.8 1/5</td>
<td>--</td>
<td>7.5/1</td>
<td>29.3/5</td>
<td>16.6 1/5</td>
<td>31.5 1/10</td>
<td>54.1 1/20</td>
</tr>
<tr>
<td><strong>$\beta$</strong></td>
<td>60</td>
<td>--</td>
<td>--</td>
<td>40</td>
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<td>--</td>
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<tr>
<td><strong>$c_r$ [MPa]</strong></td>
<td>--</td>
<td>1.38</td>
<td>--</td>
<td>0.69</td>
<td>1.70</td>
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<tr>
<td><strong>$\phi_r$</strong></td>
<td>--</td>
<td>10.5</td>
<td>--</td>
<td>30.0</td>
<td>36.1</td>
<td>--</td>
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</tr>
</tbody>
</table>

1 Stress at yield point

Tab. 3.2. Testing results. $E_T$ = Tangent Young’s modulus, $E_s$ = Secant Young’s modulus, $\nu$ = Poisson’s ratio, $\sigma_{\text{peak}}$ = peak strength, $\sigma_{\text{confining}}$ = confining pressure, $\beta$ = angle of failure plane with respect to core axis, $c_r$ = residual cohesion, $\phi_r$ = residual angle of internal friction.
<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
<tr>
<td>intact material</td>
<td>22.5 (^4)</td>
<td>-</td>
<td>10.0 (^5)</td>
<td>17.1 (^2)</td>
<td>6.5 (^2)</td>
</tr>
<tr>
<td></td>
<td>22.0 (^4)</td>
<td>8.7 (^2)</td>
<td>8.0 (^5)</td>
<td>14.5 (^2)</td>
<td>-</td>
</tr>
<tr>
<td>intensely fractured</td>
<td>17.0 (^4)</td>
<td>5.6 (^2)</td>
<td>-</td>
<td>1.5 (^2)</td>
<td>1.3 (^1)</td>
</tr>
<tr>
<td>material</td>
<td></td>
<td>0.8 (^3)</td>
<td>7.5 (^5)</td>
<td>2.4 (^2)</td>
<td>0.6 (^2)</td>
</tr>
</tbody>
</table>

\(^1\) \(\sigma_{\text{radial}} = \text{ca.} 1 \text{ MPa}\), \(^2\) \(\sigma_{\text{radial}} = \text{ca.} 5 \text{ MPa}\), \(^3\) \(\sigma_{\text{radial}} = \text{ca.} 13 \text{ MPa}\),
\(^4\) \(\sigma_{\text{radial}} = \text{ca.} 50 \text{ MPa}\), values for E-modulus are estimated from axial stress-strain curves,
\(^5\) \(\sigma_{\text{radial}}\) not given

Tab. 3.3. Comparison of E-moduli for fault zone and fractured shear zone materials.
<table>
<thead>
<tr>
<th></th>
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<th>cohesion-less zone, granite</th>
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<td>122</td>
<td>122</td>
</tr>
<tr>
<td>$m_t$</td>
<td>33</td>
<td>33</td>
<td>33</td>
</tr>
<tr>
<td>GSI</td>
<td>blocky-good</td>
<td>blocky-very blocky/good</td>
<td>very blocky/good</td>
</tr>
<tr>
<td></td>
<td>70</td>
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<td>55</td>
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<tr>
<td>$m_b$</td>
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<td>1.23</td>
<td>0.82</td>
</tr>
<tr>
<td>$s$</td>
<td>0.000464</td>
<td>0.000036</td>
<td>0.000010</td>
</tr>
<tr>
<td>$a$</td>
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<td>0.50</td>
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<td>$\sigma_{peak}$</td>
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<td>$(\sigma_{peak}, \text{Laboratory})$</td>
<td>109.4$^1$</td>
<td>82.1$^1$</td>
<td>50.5$^1$</td>
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<tr>
<td></td>
<td>(10.64$^2$)</td>
<td>(30.4$^2$)</td>
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<table>
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<th>cohesion-less zone, gneiss</th>
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<td>15.39</td>
<td>20.69</td>
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<td>(13.49$^3$)</td>
<td>(29.3$^1$)</td>
<td>(16.6$^1$)</td>
</tr>
</tbody>
</table>

$^1 \sigma_{radial} = 5 \text{ MPa,} \quad ^2 \sigma_{radial} = 4 \text{ MPa,} \quad ^3 \sigma_{radial} = 1 \text{ MPa}$

Tab. 3.4. Input parameters and peak strengths estimated by the Hoek-Brown failure criterion for a) the granite hosted shear zone, b) the gneiss hosted shear zone.
4. Concluding Remarks

Detailed mapping at surface and in existing tunnels and the detailed examination of borhole cores presented in paper 1 suggest that the structural characteristics of the shear zones in the eastern Aar massif are similar, independent of their strike and the lithology they affect.

The shear zones examined are primarily ductile showing a characteristic progressive evolution from the host rock to mylonite and locally ultramylonite. Rock compositions simultaneously exhibit an increase in the mica content. Towards the mylonitic and ultramylonitic zones, the network of small-scale shear zones anastomosing around more competent, lozenge-shaped lenses increases in density.

The shear zones also show a higher fracture density than the host rock. The fractures can be open but are mainly filled with a fine-grained, often cohesionless, structureless arrangement of micas and quartz. Highest fracture densities are restricted to strongly foliated and fine-grained parts of the shear zone and are mainly developed along transitions from mylonite to ultramylonite. Along with the fracture density, the thickness of the fracture infill increases rendering the mylonite more and more friable towards the zone of highest fracture density. In the heavily fractured zones, structureless and almost cohesionless microbrecciae consisting of mylonite fragments enclosed in a fine-grained, cohesionless matrix occur.

It is proposed that, with respect to different fracture densities and mica contents, the shear zones can be subdivided into 3 different textural subzones. With increasing fracture density and mica content these are (1) a strongly foliated zone, (2) a fractured zone and (3) a heavily fractured, cohesionless zone. The heavily fractured, cohesionless zone represents a cohesionless, structureless microbreccia. The zones are oriented approximately parallel to the rock foliation and often form a symmetric sequence on both sides of the heavily fractured, cohesionless zone which may anastomose within the fractured zone. The cohesionless zone may eventually be absent.
As the fractures mainly developed by opening, it is argued that the majority of the fractures occurring within the shear zones resulted from hydrofracturing. The fractures are superimposed on the ductile deformation structures of the Alpine deformation phase and therefore have been formed later.

Results from laboratory triaxial tests performed on shear zone samples and presented in paper 2 could reveal that rock strength and deformability characteristics of the rock material within shear zones in the eastern Aar massif are related to their structural characteristics. The geomechanical properties are mainly influenced by fracture density, foliation intensity, thickness of fine-grained fracture infill and mica content of the material.

As fracture density increases and the influence of discrete, persistent discontinuities on rock mass strength decreases, behaviour of the test samples becomes more and more representative of rock mass behaviour. Samples with rock pieces between fractures being small compared to the overall size of the sample may be regarded as isotropic and representative for the rock mass.

It seems to be possible to assign each of the shear zone subzones defined in paper 1 a typical rock mass constitutive behaviour and a corresponding failure mode. The strongly foliated zone, similar to the neighbouring host rock, behaves in a brittle fashion, failing along a single, discrete shear plane and having a relatively high peak strength and low residual strength. The fractured zone represents a transition from brittle to ductile behaviour and therefore tends to fail along a failure zone as opposed to a plane. As a result, the peak strength is generally lower but the residual strength is higher than that of the strongly foliated zone. Material from the heavily fractured, cohesionless zone behaves in a fully ductile manner and fails through large deformations in the form of barrelling over the whole volume of the sample. Rock strength increases with increasing consolidation (strain hardening).

The structural and geomechanical characteristics of the shear zones give reasons for significant stability problems when the shear zones are encountered during tunnelling. The high density of foliation planes and fractures parallel and oblique to the foliation, particularly within the fractured and the cohesionless parts of the shear zone, implies that in low stress environments in the area of the shear zones an increased number of ground control problems in the form of unravelling of interlocking blocks and wedges (i.e. structure controlled instability) may be expected. In high stress environments, the low strength and high deformability characteristics of the shear zones examined would suggest that the sheared rock mass would predominantly fail by sliding along discontinuities and crushing of intact rock pieces (i.e. stress controlled instability).

To prevent significant stability problems, the design of adequate support measures may require the consideration of the specific structural characteristics established in this
study and the typical rock mass constitutive behaviour of each shear zone subzone. The factors should be included in numerical modeling calculations.

In this respect, results from this study may help to improve estimations regarding the impact of the shear zones described on site characterization of rock mass behaviour. The study may also contribute to the improvement of general characterizations regarding the relationship in between structural and geomechanical properties.
5. References


Habimana, J., Labiouse, V. & Descoeudres, F. 1998. Influence of Tectonisation on Geomechanical Parameters of Cataclastic Rocks: Experience from the Cleuson-


Appendix A

Geophysical Logging

In this part of the Appendix the results of the geophysical logging performed in boreholes Ia (borehole across the shear zone in granite) and IIa (borehole across the shear zone in gneiss) are presented. Structural properties and the composition of the rock material encountered in these boreholes are described in detail in chapter 2.

The raw logging data usable for input into "WinLogger"- or "Viewlog"-Programs is contained on the CD attached at the end of the thesis. An overview about the structure of the CD is provided in Figure B1.

A.1. Caliper Logging

Figures A1 and A2 illustrate the results of the caliper logging. The caliper logs are presented in comparison to the different shear zone subzones differentiated in the cores and in comparison to selected structural properties of the shear zones.

In the caliper logs of both boreholes it is visible that one caliper pair always shows larger values than the other, however X- and Y-calipers vary their diameter readings approximately parallel.

The caliper logs of both boreholes clearly show increases in borehole sections having a high fracture density in addition to a high foliation intensity. Such sections are almost always contained in the fractured and the cohesionless zones of the shear zones. Within the strongly foliated zones generally having low fracture densities, only one
isolated increase can be observed in the log in the borehole in gneiss. The increase is associated with a thin layer of higher fracture density.

Decreases in the caliper log, i.e. decreases to readings smaller than 120 mm, can not be correlated to structural features.

Increases in the caliper readings may show borehole diameter increases indicating small borehole collapses or a widening of the borehole due to erosion within intensely foliated and heavily fractured borehole sections. Decreases in the caliper readings imply a decrease in the borehole diameter. As the decreases can not be connected to any structural feature, the decreases can not be explained by borehole deformation due to a reaction of the structure or the composition of the rock material to the borehole, i.e. squeezing or swelling. Borehole diameter increases or decreases due to a change of drilling techniques, i.e. changes of the diameter of the drill pipe, can also be excluded.

However, as the caliper logs were performed in almost horizontal boreholes using a tool recommended for usage only in boreholes with a declination from vertical of at maximum 30°, it has to be assumed that the sonde did not work accurately, e.g. the sonde could have been lying on one arm during logging and one pair of linked arms therefore could not open properly. While twisting around the tool during logging the situation of the arms could have changed from time to time allowing the arms to open a little bit more or closing them. The sonde lying on one arm during logging could also explain the observation that one caliper pair always read larger values than the other.

The appearance of the caliper logs therefore, in summary, has to be explained by a combination of influences from the properties of the logged material and influences resulting from logging techniques.
<table>
<thead>
<tr>
<th>Core/Depth [m]</th>
<th>Average Number of Fractures per Cm of Core Length</th>
<th>Maximum Thickness of Fracture Infill [mm]</th>
<th>Mica Content [Vol. %]</th>
<th>Average Number of Foliation Planes per Cm of Core Length</th>
<th>CALIPER X [mm]</th>
<th>CALIPER Y [mm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>6</td>
<td>12</td>
<td>35</td>
<td>16</td>
<td>110</td>
<td>136</td>
</tr>
</tbody>
</table>

Fig. A1. Caliper logs of the borehole in granite (■ isotropic rock, ■ foliated rock, □ strongly foliated zone, □ fractured zone, ■ cohesionless zone).
Fig. A2. Caliper logs of the borehole in gneiss (□ isotropic rock, ■ foliated rock, ◆ strongly foliated zone, □ fractured zone, ■ cohesionless zone).
A.2. Full Wave Sonic Logging

Compressional wave velocities and variable density logs processed by full wave sonic logging are illustrated in Figures A3 to A6. All measurements are compared to the different shear zone subzones differentiated in the cores. The compressional wave velocities are additionally presented in comparison to selected structural properties of the shear zones.

The figures showing the variable density logs (Figs. A5, A6) are composed of the records of wave trains in a near and a far receiver. Yellow and red colours represent negative and greater negative amplitudes, respectively, light blue and dark blue colours stand for positive and greater positive amplitudes.

As full wave trains were recorded only every 20 cm, the number of records along the length of the boreholes in the variable density logs is less than in the compressional wave velocity logs. However, data of both logs have the same vertical resolution.

The compressional wave velocities and the variable density logs recorded in the borehole in granite clearly show significantly increased transit times and a strong attenuation of the wave trains, i.e. a strong decrease of the amplitudes, over the whole area of the shear zone. The increase in transit times and the attenuation starts at a borehole length of about 3.3 m in an area where the rock becomes strongly foliated and fracture density increases, i.e. at the beginning of the fractured zone of the shear zone. Transit times and attenuation further increase with borehole length and reach lowest values in between 4.3 to 5.4 m in an area of heavily fragmented and extremely strongly foliated material including the cohesionless zone. Further onwards in the borehole, towards the strongly foliated zone and the slightly ductilely overprinted granite, transit times and attenuation decrease again. From a length of about 9.4 m onwards, compressional wave velocities show values of about 4000 to 6000 m/sec which are typical for intact, isotropic granite (e.g. Talebi & Young 1992, Schön 1996). Over the whole range of the log, deflections showing maximum and minimum values occur.

In comparison to the logs from the borehole in granite, trends within the logs from the borehole in gneiss are almost only visible in the compressional wave velocity log. And also in this log, trends are relatively indistinct due to a high amount of deflections towards maximum and minimum values occurring over the whole range of the log. Nevertheless, in the logs recorded in the borehole in gneiss it is visible that transit times and attenuation of the wave trains are also relatively high over the whole area of the shear zone. Velocity values within the shear zone in gneiss can be compared to those determined within the shear zone in granite. Relatively high transit times and a high attenuation already occur at the upper limit of the logs, lying within a heavily fractured area of the fractured zone. Local, slight decreases in transit time and attenuation can be correlated to local, slight decreases in fracture density, e.g. at a borehole length of about 5 m. However, in areas where fracture density and foliation intensity are high, i.e. within the fractured and the cohesionless zones of the shear zone, transit times almost keep the values occurring at the upper limit of the log. From the beginning of the
strongly foliated zone onwards, transit times then decrease towards the less fractured, intact gneiss at the lower end of the borehole. The deviation in the decrease in transit times at a borehole length of about 10 m can be correlated with a local, strong increase in fracture density.

The observations show that variations in transit time and attenuation of the wave trains may be connected to a changing fracture density, i.e. increases in transit times and attenuation may be caused by an increased fracture density. Correlation of log variations to sudden changes in fracture density are limited due to the low vertical resolution of the logs of only 40 cm. Apart from fracture density, also an increase in foliation intensity seems to increase transit times and attenuation. The influence of foliation is especially visible in the lower part of the compressional wave velocity log in granite (5.5 to 9.4 m). In this part, the rock has a strong foliation, almost no fractures occur and velocity is still lower than in the approximately isotropic granite at the lower limit of the log. The influence of foliation can be explained by the fact that the waves are partitioned at each interface between dissimilar materials, e.g. foliation planes, into transmitted, reflective and refractive waves (e.g. Brady and Brown 1993). It can be expected that the influence of foliation on wave velocities and amplitudes is at a maximum, when the waves caught by the receiver have to propagate perpendicular to the foliation planes. The influence of foliation on wave propagation is also described by Huenges et al. (1997).

It has to be noted that, when displayed in the amplitude-time mode, the full wave trains rarely resembled the idealized wave train illustrated in Figure 1.9. In almost every wave train logged, the frequencies and amplitudes did not vary much and the amplitudes stayed at a low level over the whole length of the wave train. Apart from this, especially within the area of the shear zone in gneiss, amplitudes often occurred right from the beginning of the wave train indicating the immediate arrival of a wave. This frequently observed immediate arrival of the wave train may explain the deflections within the compressional wave velocity logs showing maximum values. Deflections showing minimum values may have been caused by the stop of the wave propagation along open fractures. It may be assumed that oscillations induced by the logging tool superimposed the signals resulting from waves travelling through the rock material. It is also possible that the fractured areas of the shear zones represented "slow" formations not producing any shear waves. However, there may be other reasons for the unusual appearance of the wave trains.

Despite the wave trains mainly not being similar to the wave train illustrated in Figure 1.9, compressional wave transit times could be determined from the full wave train by picking the transit time at the first clearly visible arrival. However, it had been impossible to clearly identify the arrival of the shear waves. Shear wave transit times therefore could not be determined clearly and dynamic Young's moduli could not be established.
<table>
<thead>
<tr>
<th>Core Depth [m]</th>
<th>Average Number of Fractures per Cm of Core Length</th>
<th>Maximum Thickness of Fracture Infill [mm]</th>
<th>Mica Content [Vol. %]</th>
<th>Average Number of Foliation Planes per Cm of Core Length</th>
<th>P-Wave Velocity [m/s]</th>
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Fig. A3. Compressional wave velocities within the borehole in granite ( ■ isotropic rock, ▲ foliated rock, □ strongly foliated zone, ■ fractured zone, ▼ cohesionless zone).
Fig. A4. Compressional wave velocities within the borehole in gneiss (isolropic rock, foliated rock, strongly foliated zone, fractured zone, cohesionless zone).
Fig. A5. Variable density log of the borehole across the shear zone in granite, raw data ( ■ isotropic rock, ■ foliated rock, □ strongly foliated zone, ■ fractured zone, ■ cohesionless zone).
Fig. A6. Variable density log of the borehole across the shear zone in gneiss, raw data (isolated rock, foliated rock, strongly foliated zone, fractured zone, cohesionless zone).
A.3. Spectral Gamma Ray Logging

Results of the spectral gamma ray logging compared to the shear zone subzones differentiated in the cores and to selected structural properties of the shear zones are shown in Figures A7 and A8.

The gross gamma ray log of the borehole in granite shows one significant change at a borehole length of about 3.2 m where gross gamma radiation clearly decreases. The decrease is caused by a decrease in the amounts of potassium and thorium. The amount of uranium simultaneously increased. The mineralogical composition of the rock material does not change in this area and also fracture density, compared to the surrounding rock material, is not especially increased.

Looking at a larger scale, the logs of the borehole in granite show slightly decreased contents in potassium and thorium over the whole range of the shear zone. The content in uranium is simultaneously increased pointing towards an increased accumulation of uranium in an area of increased fracture density.

The gross gamma ray log of the borehole in gneiss shows a higher number of variations within the area of the shear zone than the log across the shear zone in granite. Variations in gross gamma radiation of the gneiss material and variations in the quantities of potassium, uranium and thorium always occur in the same way, i.e. all logs show increases or decreases at the same borehole length. Increases in radiation can be correlated with the occurrence of a high foliation intensity, i.e. the existence of ultramylonite having an increased content in biotite. However, as this correlation is only clearly visible in the area in between 4.9 and 7.1 m of borehole length, the gamma ray logs may additionally be influenced by other factors than fracture density, foliation intensity and mica content.
|---------------|-----------------------------------------------|------------------------------------------|------------------------|----------------------------------------------------------|-------------------------|-----------------------------|----------------------------|

Fig. A7. Spectral gamma ray logs of the borehole in granite (地下水: isotropic rock, foliated rock, strongly foliated zone, fractured zone, cohesionless zone).
### Table

<table>
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### Diagram

Fig. A8. Spectral gamma ray logs of the borehole in gneiss (● isotropic rock, ■ foliated rock, □ strongly foliated zone, △ fractured zone, ▼ cohesionless zone).
A.4. References


Appendix B

DATA

This part of the Appendix contains all data which was collected, but not presented in any of the previous chapters. The data is shown in form of figures and/or is contained on the CD attached at the end of this thesis. The structure of the data recorded on CD is illustrated in Figure B1.

B.1. Structural Geology

B.1.1. Spatial Distribution of the Shear Zones

Shear zones mapped in the eastern Aar massif were plotted on topographical maps having a scale of 1:25'000. The maps are contained as scanned sections on the CD attached. Figure B2 shows the position of the scanned map sections within the area of the eastern Aar massif. Shear zones, shear zones suspected and shear zones suspected but only seen as lineaments on air photos were differentiated on the maps with different colours. For legend see Figure 2.5.

B.1.2. Core Logging

Complete illustrations of all borehole cores, which were already partly presented in Figures 2.12 and 2.20, are shown in Figures B4 and B6 and are also contained on the CD attached. In the illustrations of the cores the positions of the samples taken for thin section analyses and geomechanical testing are marked. Location and the orientation of
the boreholes within the security tunnel of the Gotthard highway tunnel were described in chapters 1 and 2 and are shown in Figures 1.2., 1.3., B3 and B5.
Fig. B1. Structure of data on CD.
Fig. B2. Naming and arrangement of scanned topographical map sections contained on the CD.
Fig. B3. Starting points of boreholes la and lb within the security tunnel of the Gotthard highway tunnel. Borehole la is oriented perpendicular to shear zones structures and in this respect is drilled with a 3 to 5° dip towards SSE. Borehole lb is oriented parallel to shear zone structures and is drilled with a 3 to 5° dip towards ENE. Illustration shows a section of the tunnel records taken during tunnel construction.
Legend for profiles of borehole cores illustrated in Figures B4 and B6:

- **isotropic Aar granite**
- **ductilely overprinted rock,**
  - coarse-grained, foliated in cm- to mm-scale;
  - protomylonite
- **strongly ductilely overprinted material:** foliated in cm- to mm-scale,
  - minerals are fine-grained, high amount of mica;
  - mylonite and ultramylonite,
  - fractures with cm- to mm-spacing
- **strongly ductilely overprinted material:** foliated in cm- to mm-scale,
  - minerals are extremely fine-grained, high amount of mica,
  - mylonite and ultramylonite,
  - material is characterized by fractures,
  - material is friable, soft, cohesive
- "cohesionless, heavily fractured zone",
  - breccia zone, microbreccia,
  - heavily fragmented, cohesionless, structureless material
- **quartz-veins**
- **foliation**
- **fractures**
Fig. B5. Starting points of boreholes IIa, IIb1 and IIb2 within the security tunnel of the Gotthard highway tunnel. Borehole IIa is oriented perpendicular to shear zone structures and in this respect is drilled with a 3 to 5° dip towards SSE. Boreholes IIb1 and IIb2 are oriented parallel to shear zone structures and are drilled with a 3 to 5° dip towards NE. Illustration shows a section of the tunnel records taken during tunnel construction.
B.2. Geomechanical Testing

Complete data sets recorded during the laboratory triaxial tests are contained on the CD attached. Within the data sets $\sigma_1$ corresponds to $\sigma_{axial}$ and $\sigma_3$ corresponds to $\sigma_{radial}$. Full axial stress-strain curves constructed from these data sets are illustrated in Figures B7 and B8. Figures B9 and B10 show Coulomb residual strengths envelopes constructed from the residual strengths established. Axial stress-strain curves and residual strengths were discussed in chapter 3.
Fig. B7. Axial stress-strain curves for tested samples from the shear zone in granite (samples GR1, GR2, GR3, GR4, GR5):
Fig. B8. Axial stress-strain curves for tested samples from the shear zone in gneiss (samples GN1, GN2, GN3b, GN4, GN5a):
Fig. B9. Coulomb residual strength envelopes for samples from the shear zone in granite. For sample GR3 an envelope could not be constructed as no data is available.
Fig. B10. Coulomb residual strength envelopes for samples from the shear zone in gneiss. For samples GN1 and GN5a-c envelopes not could be constructed as no data is available.
Curriculum Vitae

Personal Data

Name: Laws
First Name: Susanne
Date of Birth: October 10th, 1969
Place of Birth: Koblenz (D)
Nationality: German

Education

1976 – 1980 Elementary school Altkarthause, Koblenz (D)
19.05.1989 Final degree: Abitur

1989 – 1991 Studies in Electrical Engineering at the RWTH Aachen (D)
1991 – 1997 Studies in Geology at the University of Mainz (D)
Main Subjects: Applied Geology, Structural Geology, Petrology, Soil Sciences
07.05.1997 Final Degree: Diploma in Geology
1997 – 2001 Doctoral thesis at the Institute of Geology/Engineering Geology, ETH Zürich (CH)