CHAPTER 4.
MICRO-STRUCTURAL ANALYSIS OF FAULT MOVEMENTS: FAULT KINEMATICS AND PALEOSTRESS

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In this chapter we will see how a detailed micro-structural investigation of the structures identified using remote sensing can give information about the kinematic and dynamic aspects of superficial, brittle and recent deformation. We will discuss how to distinguish different kinematic phases in multi-phase deformed terrains, how to attribute a relative timing to the structures, and how to use them to obtain information concerning the active tectonic regime, and its characteristics. The problem of deducing dynamic information (stress) from kinematic data (strain) is addressed in its theoretical context and we will be concerned with the extends to which it is applicable in practice. For this, we need to discuss the physical background and fundamental processes responsible for the generation of the brittle faults and their kinematic markers, and evaluate to what extend these markers can be representative for the regional kinematics, and the deduced local stress field for the regional stress.

4.1 Problem statement: stress and strain

Although the principal concepts of kinematic analysis of rock deformation comprise brittle- as well as plastic deformational regimes, the discussion in this chapter will be limited to the brittle mode of deformation, as this is the main style found while studying active tectonics. This whole work is generally limited to the brittle tectonic environment, only briefly touching the brittle-plastic transition zone when discussing the depth of faulting and shear zones. As stress and strain are concepts whose use in structural geology remains highly debated, special attention will be payed in this chapter to a thorough discussion of the related concepts, their use and their limits in tectonic studies. Furthermore, the inverse problem of deducing (paleo)-stresses from geological structural data will be discussed for the brittle regime.

Strain describes the deformation of a (rock) body in terms of its final shape relative to its initial shape. Stress describes the forces acting on every point of this body. Providing high enough stresses, rocks will eventually accommodate to these stresses by deforming. Stress is not directly measurable, but nevertheless it is important to know the stress situation for reconstructing and estimating tectonic regimes, and for assessing fault kinematics.

In general, strain is a measurable parameter. Both for ancient zones (orogens, basins,...) by means of the structural geological tools, and for active deformation zones by means of geological and geodetic techniques (laser, GPS, ..). Stress on the other hand, can only be deduced, and special care has to be payed to consider the local situations and to estimate the deformational system in terms of coaxiality. This means that one has to assume a parallelism between the principal strain axes, and the principal stress axes [Twiss and Moore, 1992].

It is generally assumed that, in the plate-tectonic framework, large scale deformations occur due to local response of the lithosphere to induced stresses. Understanding the origine and distinguishing the different types of these stresses, as well as knowing their orientations and magnitudes is therefore a crucial tool in analysing and understanding tectonic deformation. Knowledge about the paleostress field which was active during the development of the investigated tectonic feature can give crucial information about the kinematic history of an area. Knowledge of the active stress field is important in the frame of seismic hazard management, and the estimation of the stress field evolution in time can result in fundamental conceptions about the earth dynamic system.

In spite of its importance, stress remains a confusing and somewhat esoteric concept, leading to lively discussions about its position (generic or dependant, [Marrett and Peacock, 1999]), measurability [Tikoff and Wojtal, 1999] and consistency in tectonic studies [Pollard et al., 1993; Pollard et al., 1999].

Over the years, two different ways of viewing the deformation of geological structures developed. One view is that in which the stresses are the independent variable, acting on geological structures, whose deformation is the dependent material response to the applied stresses [e.g. Pollard et al., 1999]. Another view treats the material displacements as the independent parameter, causing stresses to build up in the structures, making them the dependent material response to strain [Tikoff and Wojtal, 1999].

An example of this ongoing discussion is illustrated by Tikoff and Wojtal [1999], who question the possibilities of deducing information about stresses from deformed rocks. They argue that for non-coaxial deformation, such as simple-shear, the principal movement directions are parallel neither to the infinitesimal nor the finite strain axes. This is due to the interaction between rock anisotropy and minor perturbations in the principal movement directions. The velocity boundary conditions control the deformation, whereas the stresses adopt orientations and magnitudes to conform with these conditions.

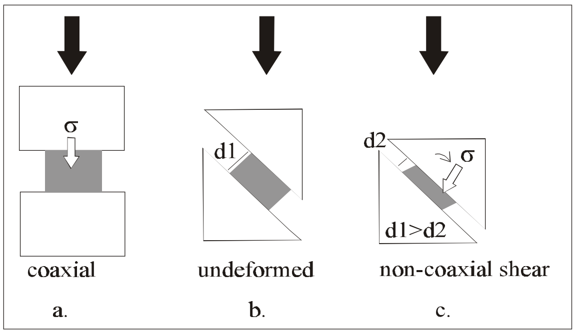

Fig.4.1 Cartoon of experimental deformation. Black arrows show movement direction, white arrow stress orientation. a. a system of coaxiality. b. a non-coaxial system (transpression) in undeformed state, and c. this system during deformation, where movement and stress are no longer parallel [Tikoff and Wojtal, 1999].

In analog modelling the boundary conditions, although referred to as stresses, consist of velocity (piston, plate,...), so that displacement rather then stress is the independent variable in most tectonic experiments. In such cases, the stresses develop as a response to the imposed displacements (as for example in fig. 4.1).

Another prerequisite for the deduction of stresses from strains following the constitutive equations is the isotropy of the deformed material. In non-isotropic rocks, analysis becomes much more complicated. Non-coaxial deformation due to these anisotropies can lead to contradicting results in the interpretation of paleostress deformations.

In our opinion, paleostress reconstruction from fault slip data suffers from another disadvantage. In order to collect enough fault- and slip orientations to be able to constrain the paleostress axes, it is often necessary to include data from several exposures, for which the local stress situation could have been different, as stress orientations seem to vary, induced by anisotropy and fault movements on other faults [e.g. Pollard et al., 1993]. This implies a statistical error in the obtained results, which is difficult to estimate.

It has also been argued that, because larger faults often have smaller splays with syn- or antithetic behaviour, which on their turn again show minor splays with varying orientation, the orientation of the stresses one infers from fault geometry and kinematics is scale dependent [Tikoff and Wojtal, 1999]. The same authors argue that the relatively good consistency and continuity of the stress orientation on a continental scale [Zoback, 1992] is due to the relatively high isotropy of such regions, leading to coaxial deformations and a parallelism between shortening (displacement) directions and principal stress directions. In regions with strong anisotropy's, such as the Central-Asian (amalgamated terrane) deformation zone, non-coaxial deformation should lead to local perturbations in the deduced stress field. It is part of this study to evaluate this hypothesis and to define the use and limits of kinematic fault slip analysis and related paleostress reconstructions.

Marrett and Allmendinger [1990], proposed a method of kinematic analysis of fault slip data, considering strain instead of stress. They argue that dynamic (stress) and kinematic (strain) analysis should be combined in the analysis of fault slip data. The uncertainty concerning the applicability of the basic assumptions of dynamic analysis (see Appendix B) on specific field data is thought to be reduced by their kinematic analysis, where the consistency of the assumptions can be tested in the field [Marrett and Allmendinger, 1990]. Indeed, graphical determination of the kinematic axes of shortening and extension are basically an alternative representation of the field observation data, equivalent to fault plane solutions P and T axes. This would imply that kinematic analysis of fault slip data yield kinematic (strain) principal axes, in stead of stress axes.

Despite all this opposition against the use of stress as a quantifying indicator of tectonic activity, its use has proven to be valid and powerful over the years [e.g. Hancock, 1985; Angelier, 1989; Engelder, 1993a; Fry, 1999; ...]. This study uses fault kinematics and deduced stress in order to define the tectonic regimes affecting the study area, and to estimate their evolution with time. To study rock deformation in terms of stresses, it is important to define the types of stress acting on rock bodies, and their analytical and physical meaning.

Appendix B offers a short overview of the basic concepts used, related to (paleo)-stress reconstructions. They are discussed and analysed in the frame of the specificity of the present study, namely in order to provide a basis that allows a sound evaluation of the application of the method in the study of recent tectonics.

4.2 Paleostress reconstructions

On of the goals of structural geology is to relate the observed geological structures to the causative tectonic forces, responsible for the formation of the structures. One way of describing the tectonic forces, is to quantify the stress pattern active during the formation of the observed structure. This can lead to a model of the tectonic history and - evolution of the investigated area.

As discussed in Appendix B, the stress tensor controls the slip of a fault plane with a specific orientation. We will use these theoretical considerations to evaluate the inverse problem, namely the possibility to determine the principal stress directions and stress ratio (so the reduced stress tensor) from observations of fault movements. Although the theoretical background has been discussed and elaborated in detail, [e.g. Angelier and Mechler, 1977; Angelier, 1994] we find it opportune to briefly recapitulate the basic assumptions and discuss them in the light of the recent publications on their behalf, as this can't be found in any text-book. Once the basic assumptions for stress inversion are qualified to their validity and their limits, we give a short outline of the inversion methods.

The basic hypothesis used in the inversion problem is the so-called *Wallace-Bott* assumption [Wallace, 1951; Bott, 1959], stating that *any slip on a fault plane is parallel to the shear* *stress acting on the plane*. This is a conceptual model, on which many authors based their methodology to determine paleostress orientations from fault slip data [a.o. Angelier and Mechler, 1977; Etchecopar et al., 1981]. Another basic assumption used in paleostress reconstructions is that the *regional stress field is homogeneous in space and time* (during the period of faulting). By ‘regional’ is meant, in this sense, a large area relative to the fault scale.

Being a difficult problem to solve analytically, due to the dependency on many different parameters, theoretical studies never brought clarity on the validity of the basic assumptions. Nevertheless, during the last decade, some studies using numerical methods showed that, in general, the observed slip on a plane remained sub-parallel to the direction of maximum shear stress on the plane [Dupin et al., 1993]. Another study examined several parameters possibly causing deviations from the Wallace-Bott assumption [Pollard et al., 1993]. It showed that the direction of slip on a fault plane can deviate from the direction of maximum shear stress in cases of fault interactions, in function of the depth of faulting and in function of the length-width ratio of the fault [Pollard et al., 1993]. Fault interactions have the highest influence on fault slip directions. This means that the movement on one fault plane accommodates the stress state and causes a change in regional stress orientation in the vicinity of that fault, so influences the movement on another fault.

Although experiments prove violations of the Wallace-Bott assumption caused by fault interactions with angular discrepancies between shear stress and fault slip direction up to 40° (in their experiment), the statistical mean and standard deviation of the discrepancy angle remains low and within the precision of field data measurements and stress inversion methods [Pollard et al., 1993]. This illustrates the importance of the statistical character of the method. A large fault population is necessary to statistically eliminate the (statistically few) deviating slips and obtain a reliable estimate of the (reduced) paleostress tensor and related tectonic forces.

Faults are weakness zones in the brittle part of the lithosphere, along which movement can take place in response to induced stresses. When faults undergo displacement, depending on geological and rheological conditions (see § 4.3), strain markers can be formed on the fault surface. Very often such strain markers are striations (lineations) elongated parallel to the shear direction of the fault, sometimes indicating the shear sense. Fault planes with shear-parallel striae are called slickensides. Following the basic assumptions, a slickenside surface has a one by one relationship to the regional stress tensor acting on the surface. Equation (4) in appendix B shows that every stress vector (known by the normal of the fault plane and the shear direction and sense) has a relation to the stress tensor, and thus constrains a range of possible orientations of the principal stress axes.

In the case of conjugated neoformed faulting, two conjugated faults completely define the reduced stress tensor. As discussed above, the statistical reliability of the basic assumptions requires a much larger set of fault measurements constraining the reduced stress tensor. Also, in cases of movements on inherited faults, a statistical approach is needed to limit the zones of possible stress axes’ orientations, and a large number of faults is needed.

Superimposing the dihedra of all the considered fault planes, the zones of possible orientations of the stress axes are limited. The use of dihedra graphical calculation of the reduced stress tensor from a population of faults is extensively discussed elsewhere [Angelier and Mechler, 1977; Angelier,1994; Delvaux, 1993; Dehandschutter, 1996].

As the reduced peleostress tensor has 4 variables, and the number of fault slip observations is much larger, a statistical model to obtain the best fit between the actual slip and theoretical shear stress is needed. The graphically obtained reduced stress tensor has to be modified in function of the correlation between observed slip and calculated slip on each fault. This method is explained in Appendix C.

Because there still is a misunderstanding of what exactly means the Andersonian stress and faulting regimes, and how the transition between the different regimes occurs, we refer to fig. 4.1.4 in appendix B. This figure shows the relationship between the stress and faulting regimes, and the way the transition occurs. This transition is basically due to a change in relative magnitude of the different principal stresses, and not due to a change in physical orientation of these principal stresses, as was supposed by [Petit et al., 1996].

It is important to bear in mind that the violation of the stress-strain coaxiality in simple-shear and transcurrent tectonics (see problem statement) hampers the validity of the paleostress reconstruction in such particular cases. This inference, together with the theoretical complications discussed above in this chapter will be of importance for understanding the discussion and interpretation of the field observed structures in the second part of the study (Part II). Also crucial for the understanding of the applied interpretation of Part II is the mechanical background of faulting and its consequences for the expression of faulting and relative timing in the field. This is discussed in the next paragraph.

4.3 Fault rocks, rheology and depth of faulting

Depth of faulting, fluid circulation, mineralogy and rheological conditions (P, T, strain rate, fluid pressure, stress, etc.), determine the type of fault rock and slickensides produced in the brittle deformation regime. The expression of faulting generally changes with depth (P-T conditions), and establishing a relative timing of faulting therefore depends on these phenomena, which will consequently be discussed here. Next, we will discuss the fault types developing generally with increasing depth. This is important, because it enables us to distinguish the different fault rock types in the field, and relate them to different tectonic regimes, on their turn allowing a relative timing and modelling of the stress field evolution.

4.3.1 Rheology: Model of a shear zone

Rheology is the branch of physics dealing with the deformation and the flow of materials. Rheological behaviour of a body is the deformational response of that body to applied stress. The understanding of the rheological behaviour of rocks is very important for understanding the tectonic meaning of observed structures, because the rheology of a rock body determines the way it deforms. Without going into great detail, this section discusses the generalities of upper crustal material response to tectonic stress, focussing on the elastic part of the deformation spectrum, which is related to brittle faulting (and recent tectonics).

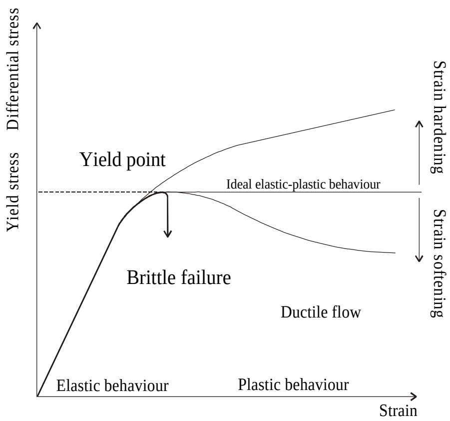

Fig. 4.2 Stress-strain relations for brittle and ductile deformation.

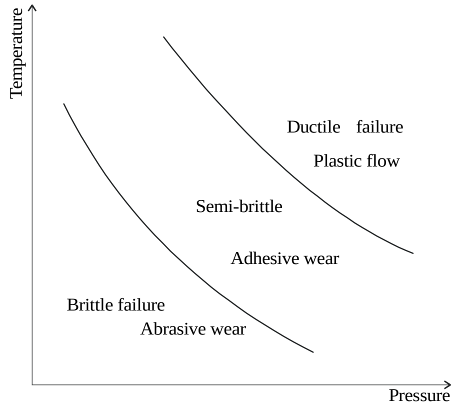

Fig. 4.3 Temperature-Pressure relations for brittle, transitional and ductile behaviour.

Faulting is controlled by the rheology of the country rock. There are basically two types of response of rocks to stress: brittle (elastic-frictional plastic) and ductile (viscous-plastic, fig. 4.3). The brittle deformational regime is controlled by elastic stress-strain relations. Deformation with stress relief (total drop in the ideal case and intense strain-weakening in real circumstances of so-called coulomb-materials) takes place at stresses exceeding the yield stress (rock strength, fig. 4.2), following the deformational regimes of crack propagation, shear rupture and frictional sliding (appendix B). As indicated by the Mohr-Coulomb failure criterion, brittle failure depends mainly on the confining pressure, and is relatively slip-rate or load-rate independent. Regions in the lithosphere following this behaviour are called ‘brittle’. The rock strength increase with depth (confining pressure) is analogous to the Mohr-Coulomb failure envelop, and can be presented as follows:

$$\tau_{max} = (\sigma 1 - \sigma 3)/2 = \mu \sigma_v$$

It was stated by Byerlee [Byerlee, 1978], that the value of μ is approximately 0.75-0.85 for σ_n<200 MPa and varies with rock type and with lithostatic pressure. For high effective normal stresses (σ_n>200 MPa) the relation becomes

$$\tau = 60 + 0.6\sigma_n$$

These relations are known as ‘Byerlee’s law’. It generally and roughly holds until mid-crustal levels (up to about 15 km).

Ductile deformation on the other hand, is insensitive to normal stress, but highly temperature and strain-rate dependant, as indicated by the dislocation creep failure criterion:

$$\tau = B\varepsilon^{(1/n - 1)} \exp^{(Q/RT)}$$

where B, n and Q are constants that depend on rock type, ε is the strain rate, T is temperature and R is the gas constant [e.g. Thatcher, 1995].

For ductile deformation, the stress at the yield point ideally remains constant for increasing strain. Discrete fractures (of Mode I, II or III) are associated to brittle deformation, while plastic flow generates bulk zones of ductile flow (ductile shear zones). Ductile flow can result in strain hardening, in which case the stress (friction coefficient) increases with increasing strain, while strain weakening reduces differential yield stress with increasing strain. Strain hardening can occur in high confining pressure zones with relatively low temperature, whereas strain softening is temperature dependent. The actual ductile behaviour will be governed by P, T, strain-rate and chemical composition. It should be noted that some authors prefer to talk about ductile deformation in a non-mechanistic, but phenomenologic way, relating it to the ability of a rock body to undergo non-localised strain, rather then relating it to intracrystalline plasticity deformation mechanism [Rutter, 1986].

The transition zone between brittle and ductile deformation (fig. 4.4) forms a zone in the lithosphere where stress is accommodated partly by shear rupture and frictional sliding, and partly by ductile flow (bulk deformation). This zone is referred to as the semi-brittle field of deformation (fig. 4.3) or the transition zone (fig. 4.5).

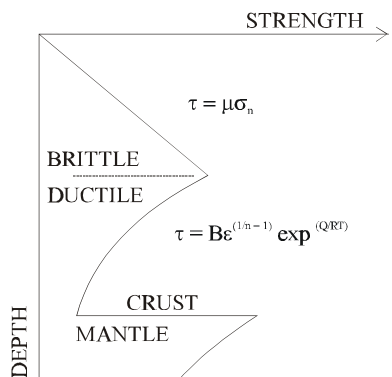

Fig 4.4 Maximum shear stress as a function of depth for the continental lithosphere.

The degree to which the laboratory derived generalised strength profile shown in fig. 4.4 applies to the continental lithosphere is uncertain, as we discussed in chapter 1 (§ 1.3.4). Brittle failure seems to occur at much lower stresses than predicted by Byerlee’s law on fig. 4.4. This could be due to an increased pore fluid pressure [e.g. Axen, 1992].

Figure 4.5 shows a scheme of the relations between depth of faulting and various fault rock. With increasing P and T, there is an evolution in wear products [Sibson, 1977]. At low temperatures, deformation occurs by pure mechanical wear, i.e. fracturing with production of loose wear material (abrasive wear, gauge). With increasing P-T conditions, abrasive wear keeps forming gauge, but depending on fluid circulation and composition of the rock, mineralisation inside the fault contact can cause preferentially oriented minerals, forming slickenlines. Generally, abrasive wear formed fault-rock is classified as cataclasite. As temperature increases, the wear becomes more plastic, and the semi-brittle field of adhesive wear is entered. There is local shear of asperities. There can be reorientation of crystals on the fault plane and crystallisation of quartz, carbonate, or iron-oxide on the fault plane. From about 300°C on, when quartz (generally) becomes plastic, the rocks forming in a fault zone become more mylonitic. As rock metamorphic facies follows the same depth-temperature dependency, it is possible to associate the different fault styles with ranges of metamorphism.

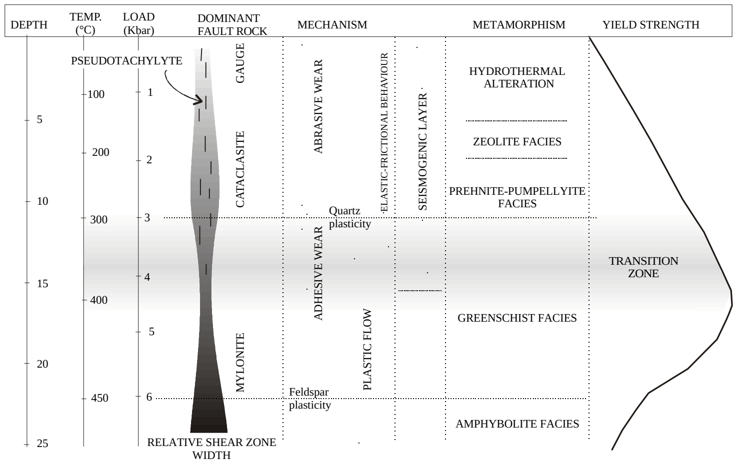

Fig. 4.5 Model of a shear zone of quartzo-feldspathic composition. Modified after Sibson, [1977] and Scholz, [1988]. Geotherm and yield strength are only illustrative, as they highly vary for different regions. Pseudotachylyte occurrence indicated by vertical lines.

Quantification of the P-T conditions of the faulting styles is based on mechanical experiments and modelling of geothermal gradients in the crust. The 300°C marks the onset of plasticity of Quartz, and 450°C this of feldspar. Depth and T for the transitions should give an order of magnitude rather than a constant value, because of the varying geotherms in the crust.

Also, the rock yield-strength profiles are very dependent from the strain rate in the plastic regime, the mineralogical composition and the applied geotherms [Sibson, 1983; Ziegler et al., 1995]. They therefore are case-sensitive and should be treated here only indicatively. The yield strength increases approximately linear with depth, as long as the rock shear is dominantly frictional, with either abrasive or adhesive wear mechanisms. Below the transition zone, rock strength follows the high temperature flow law, and decreases with depth in a way which highly depends on the strain rate.

The fault type, yield strength and rheological profile of the crust is also related to the depth of the seismogenic zone [Sibson, 1983; Swanson, 1992]. Most shallow earthquakes occur at a depth shallower than 20 km, corresponding generally to the zone in which quartz is deforming elastically in response to differential stress, so to the brittle deformation regime [Scholz, 1988].

These observations led to the two-mechanism rheologic model of the earth crust, which is approached as a two-layered upper part of the lithosphere, with an upper, elastic zone, and a lower, ductile zone [e.g. Sibson, 1983]. These models describe an increase in frictional strength with pressure (depth) in the upper crust, following Byerlee's law, until dislocation creep becomes the dominant deformation mechanism and strength decreases with increasing temperature at mid-crustal levels. The brittle-ductile transition is often supposed to be quite abrupt. However, recent studies show the important influence of strain-rate dependency of strength, and the fluid assisted mechanisms, seriously influencing the strength of the lower part of the upper, brittle crust [Chester, 1995; Ziegler, 1995]. The slip velocity can strengthen or weaken the fault zone, depending on the fault mechanisms, by influencing the coefficient of friction of the fault. This brings important complications to the strength profile of the crust, as Byerlee's law uses a fixed value for the coefficient of friction, which seems to be not always applicable. The strength profile of the crust, on its turn, influences the type and dynamics of upper crustal (and surficial) tectonics, and is therefore important to tectonic studies.

A very important factor weakening a fault (zone) is the presence of water and other pore fluids in the fault plane(s). It can favour pressure solution, lower the temperature of quartz plasticity, and reduce the effective stress relative to a dry fault. These processes have a weakening effect, decreasing the differential stress (and thus the maximum shear stress) required for shear-failure in the elastic-frictional regime [Sibson, 1977]. Also, the width of the fault zone influences the evolution of the shear-strength with depth, with thick shear zones behaving significantly weaker then thin zones [Chester, 1995].

For the purpose of the present study, a description of different fault styles as the expressions of the rheological circumstances at the original location of faults observed in the field will be evaluated in the following section.

4.3.2 Fault rock and slickenlines

As described above, the type of rock occurring in a fault depends on the depth and rheology of the location of faulting. In the same way, slickenlines, which sometimes form on the fault plane, express the local conditions of faulting in their texture, type and composition. Slickensides are polished fault mirror planes that can form during frictional-slip. The striae on the slickenside planes indicate the local displacement direction of the fault. The shear sense criteria on faults, often involving slickenlines and slickensides are discussed elsewhere [e.g. Petit, 1987; Dunne and Hancock, 1994; Bolbas et al., 1997; Means, 1987]. Here we only discuss the relation between type of slickenside and faulting environment.

The relation between fault rocks and slickenlines is described in table 4.1. Temperature, depth and strength profile are only indicative.

temperature (°C)	depth (km)	fault rock	slickenlines	yield strength
100	5	gauge and fault breccia	Ridge-in -groove iron coating (hematite)	
200	10	cataclasite	Q-lineations Chlorite-Epidote lineations	
400	15	mylonitic rock		

Table 4.1 Fault rock types in function of the faulting depth.

Although fault mechanism and associated fault rock vary with depth, the type of fault rock and slickensides is also controlled by the slip-rate or strain rate, and by its variations during the seismic cycle [Power and Tullis, 1989]. Aseismic creep can ductilely (by adhesive wear mechanism) deform the same fault that in another period of the seismic cycle can form cataclasite (by abrasive wear), slickensides (a.o. by polishing) or pseudotachylyte (by frictional melting, see further). Note also that the width of a shear zone increases with confining pressure, so with depth (fig. 4.5) [Scholz, 1987]. The fault zone width is also slip-rate dependent. The morphologic expression and the gauge thickness linearly increases with slip-rate.

Gauge and fault breccia

In upper crustal conditions with near-surface fault movements, low-temperature fault rock can develop. A general name for the unconsolidated wear material in a fault, originating from the frictional sliding of the fault, is *gauge*. Gauge is displaced material, composed of crushed fault- wall material that can be hydrothermally altered. Gauge is differentiated from fault breccia, by the criteria of the amount of visible fragments (respectively <30% and >30%; Sibson, 1977). Clay gauge is a very fine-grained gauge. At the surface, clay minerals can result from gauge weathering in atmospheric conditions. At depth, hydrothermal activity can cause the weathering.

Experimental studies of fault gauge showed that the size of gauge fragments decreases with confining pressure (depth). Porosity reduction, strain hardening and compaction of the gauge can occur with increasing strain. Gauge dilation is associated with high strain rates, whereas compaction of the gauge layer, and pore volume decrease occurs at low strain rates [Morrow and Byerlee, 1989].

Fault breccia and gauge often indicate shear sence, since pinnate joints, Riedel shears, structural slickenlines and fault striae easily form in the soft breccia.

Cataclasite

Cataclasite is a general name for the consolidated, cemented fault rock that forms in upper crustal, brittle (elastic behaviour, frictional-slip regime) conditions, with fragmentation of wall rock material, grain fragment rotation and frictional grain boundary sliding [Sibson, 1977]. Cataclasite is differentiated from gauge by the amount of matrix (respectively >30% and <30%). Cementation by fluid circulation often occurs in cataclastic rocks. Cataclasite is often enriched in quartz relative to the wallrock, due to hydrothermal fluid circulation. Cataclasite can contain striated surfaces (slickensides), possibly indicating or corresponding to (relatively) high-velocity slip events in (relatively) dry conditions.

Pseudotachylyte

Pseudotachylyte is a glassy fault rock formed by frictional melting under high slip rates (0.1-2 ms^{-1}) during coseismic slip events in the seismogenic, brittle layer of the crust [Sibson, 1977; Grocott, 1981; Spray, 1992], or in the brittle-plastic transition zone in mid-crustal levels [e.g. McNulty, 1995]. Pseudotachylyte is mostly associated to cataclastic deformation, although it can form in a mylonite association in high strain-rate conditions [Swanson, 1988].

Micro-structural characteristics of the pseudotachylyte associated with syn-deformational brittle fractures can reveal the kinematics of the faulting associated with pseudotachylyte formation [Grocott, 1981; Swanson, 1988; Theunissen et al., submitted]. The possibility of pseudotachylyte to be dated by absolute methods [Kelley et al., 1994; Spray, 1995; Thompson et al., 1998], together with its kinematic indicators makes this fault rock an important potential for dating and quantifying brittle deformations[1].

Slickensides

Slickensides are non-penetrative, shiny, reflective fault surfaces. Like fault rock, they form due frictional slip on the fault plane and hence they usually occur in zones of gauge, fault breccia or cataclasite. The thickness of the fault rock they develop on can vary widely, from mm to m scale. The generation of a slickenside can be a two-phase process. First, gauge is created by frictional slip. Polishing and striation of the fault surface occurs during or after cementation of the gauge, with further slip [Engelder, 1974].

There are several types of slickensides, such as polished surfaces, coated fault planes, fibrous mineral striae, streaked fault gauge, ridges and grooves,... Slickensides can carry striations on their surface, originating from a wide set of processes, such as asperity ploughing, debris streaking, erosional sheltering, and fibre growth [Means, 1987]. Different types of slickensides develop in different faulting regimes, depending on depth of faulting, fluid pressure, hydrothermal fluid circulation in the fault plane, slip rate and composition of the fault walls. A few studies estimated the depth of particular slickenside formation, as well as the temperature of formation, chemical composition of the slickenside material, and the slip-rate [Power and Tullis, 1989]. The latter proposed a general alteration of high strain-rate

[1] The role of pseudotachylyte in the formation and evolution of the study area has been discussed in a separate paper [Theunissen et al., submitted], and will not be discussed in the present study.

cataclasis with dilation and fragmentation, and low strain-rate slip with precipitation of hydrothermal fluids, and formation of a preferred crystallographic orientation. Repetitions of this alteration would eventually form the slickensides. Other types of slickensides seem to form under high strain-rate with frictional melting and concentration of iron on the fault surface [Spray, 1989]. Near-surface movements can result in gauge-smearing and ridge-in groove lineations. The surface of slickensides can vary in roughness from very rough to very smooth. It is often composed of Si-Al-Fe combination, where the top of the surface is composed of FeO and Fe_2O_3, and Al occurs in kaolinite flakes, dispersed in the quartz matrix.

Iron-coated striae

Under conditions of relatively dry faulting, high rate coseismic slip can result in surface heating of a thin surface layer (<150 micron) showing partially melted zones, resulting in polished slickenside formation. Devitrification processes however generally prevent the direct observation of remnants of frictional melt in natural outcropping slickensides. In oxidising environments, there can be an Fe enrichment on the polished slickenside surface due to temperature dependent diffusive transport. This was shown for artificial slickensides by SEM, TEM and microprobe investigations [Spray, 1989]. Although they can form at the surface, such slickensides generally form at shallow depths, in and under the zone of dilatant gauge formation where the confining pressure is high enough. The formation of such high strain-rate frictional melting surfaces with concentration of Fe, relates the iron-coated slickensides to fast slip events, as opposed to the precipitated minerals related to creep-slip.

Quartz-Calcite mineral fiber striae

Slickensides composed of quartz or calcite mineral fibres generally form by pressure solution slip or slip with fluid circulation. The mineral fibers grow by precipitation of hydrothermal fluids or by pressure solution. The elongate crystals grow parallel to the local displacement direction [Means, 1987]. The formation of quartz-gauge with both cataclastic zones and mineral fibers seems to be related to a different behaviour and response at different stages in the seismic cycle. Cataclasite with grain size reduction, and boundary sliding occurs during high slip-rate deformation, whereas quartz lineations with preferred optical axes develop during the inter-seismic low slip-rate period [Power and Tullis, 1989]. Thus, quartz (and calcite) lineations can form in a wide depth range of the crust, ranging from several kilometres up to the brittle-ductile transition zone at roughly 10 km.

Chlorite-Epidote mineral fiber striae

Although also mineral fibre striae, chlorite and epidote fibres are believed to belong to a different rheological environment then the quartz-chlorite striae. Fault planes showing such type of striations generally reflect movements in the brittle-plastic transition zone, and can therefore be related to relatively deep movements in the transition zone, starting at lower green-schist facies conditions. In field conditions these slickensides always predate iron-coated and clay striae, by cross cutting relations.

4.4 Fabrics in fault zones

The mechanical behaviour of brittle fault zones is of great importance for accurate regional structural reconstructions and tectonic modelling. Recognising fault and fracture patterns related to this mechanical behaviour allows kinematic reconstructions both in remote sensing surveys and in field observations. Early laboratory investigations on fracture development during shear [e.g. Cloos 1955; Byerlee et al., 1978] and field observations already showed some interesting general observations [Logan et al., 1992]:

- A repeatable fabric was produced under a wide range of confining pressures, strain rates, gauge thicknesses and material types.
- R1, R2, P and Y fractures (fig. 4.6) are the most abundant elements developing during shear. The type of element forming depends on the amount of shear strain, so the fabric changes with shear strain.
- The fabric array seems to be self-similar over a wide scale-range from microscopic to outcrop, for a wide variety of materials.

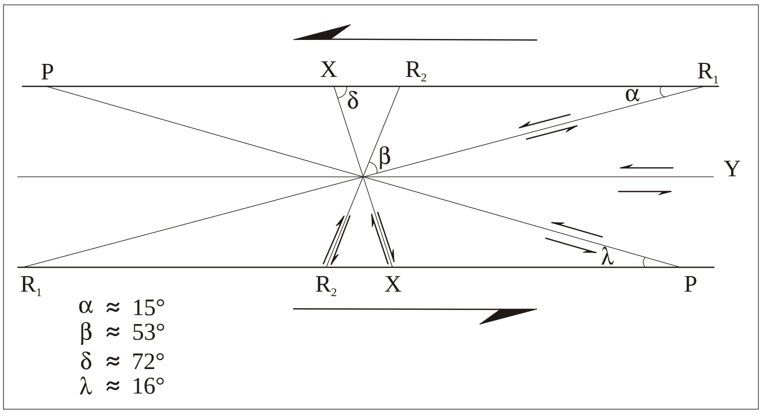

Fig. 4.6 Fracture array and geometrical relationships in fault gauge during triaxial compression. After [Logan et al., 1992].

The study of the fabric types developing during brittle failure of rock with gauge formation elucidated the mechanism of fracture development during shear by showing the evolution of shear zone fabrics with increasing shear strain. The following sequence was observed: at low strain, R_1 form first, followed by R_2 and P fractures (fig. 4.6). With strain increase, Y fractures start to form at various levels inside the gauge. Eventually, new low shear strain zones start to form between two Y fractures, reproducing the initial situation at a sub-zone inside the original shear zone. At very high strains, finally, Y fractures parallel to the moving fault blocks take up all shear [Logan et al., 1992].

The fabrics described here are a very useful instrument for determining shear sense of deformed rock at various scales, from thin-section to outcrop.

4.5 Application: use and limits

The traditional analysis of fault systems uses a 2D Coulomb criterion for failure for inferences about stresses responsible for the faults. The application of this theoretical concept to real-world fault systems should take into account that local stress variations, rock anisotropy, non-frictional slip and non-coaxial deformation can restrict the applicability of such analysis.

There is a growing perception that stress near major continental fault zones is not the simple uniform field predicted from basic Coulomb theory. By now, it is well known that the remote stress field acting on a rock body can be influenced by existing cracks and fractures, locally modifying its orientation. Local rotations of the stress axes at the vicinity of the pre-existing fracture walls can cause a change in orientation of the neoformed fractures, influenced by the orientation of the pre-existing fractures, although this influence occurs only in the close vicinity of the older fractures [Dyer, 1988]. Although the regional stress field might be relatively constant [e.g. Zoback, 1992], stresses in rocks between faults are not always uniform due to fault interaction, leading to local deviations in paleostress orientations.

Many large continental faults are weak, as they move under low resolved shear stresses at high angles to the σ_1, rather than at high shear stress predicted by Byerlee’s law and with small angle to the σ_1, predicted by Andersonian mechanics [e.g. Zoback and Zoback, 1990]. Fault weakness implies that failure occurs at low differential stress, corresponding to low shear stress, possibly imposed by a high pore fluid pressure reducing the effective stress [e.g. Axen, 1992]. For weak faults, anisotropic rocks affected by faulting will influence the local stress field and the deformation near a curvilinear fault, as the principal stress directions will tend to parallelise with the principal anisotropy directions. This tendency increases with increasing degree of anisotropy [Chester and Fletcher, 1997]. Restraining and releasing bends along weak fault segments (ramps for thrust faults and fault bends and undulations in strike-slip faults) cause a change in stress state in the fault blocks sliding past the bends. The maximal principal stress direction near the fault in a fault block passing a restraining bend tends to change from fault-parallel just before the bend, over fault-normal at the bend, again to fault parallel after the bend. This causes secondary shear-fractures to deviate in stress state (and hence in shear sense) from the imposed regional stress field [Chester and Fletcher, 1997]. Such conditions, not unlikely to occur in many continental fault systems, can cause deformation in the vicinity of bends opposite to the regional deformation. For the San-Andreas fault system, left lateral shear on faults parallel to the San-Andreas, discrepant with the overall right lateral shear movements on this fault are observed [Shamir and Zoback, 1992], and explained by a 2D analytic elastic plate model [Saucier et al., 1992]. This model indicates zones of opposite shear sense on secondary shear fractures near bends in a moving, undulatory (sinusoid) strike-slip fault. Equivalent settings are likely to exist in the presented study, and therefore the kinematic and dynamic analysis has to be carried out with care.

Slip on large non-planar faults causes the bounding fault blocks to rotate, and with them the minor faults inside the blocks. This results in rotated paleostress directions, not coinciding with the actual directions at the time of formation. Care has to be taken to recognise possible fault block rotations.

Apart from the natural variations in the local stress field and the related variations in measured paleostress orientations, focal mechanisms and in-situ stress measurements discussed above, several methodological considerations concerning the stress-strain relations and other basic assumptions in paleostress reconstructions have been proposed.

A first consideration deals with non-coaxial deformation. Non-coaxial deformation implies that certain fault-slips occur in directions not conform to the contemporary paleostress orientations, leading to misfits and incompatibility of the considered slip data. Wojtal and Pershing illustrated the possibility of an orientation difference between the computed paleostress directions and the paleostrain orientations in a fault zone [Wojtal and Pershing, 1991], indicating non-coaxial deformation. Non-coaxial strains generally results from slip partitioning onto one of the several fault sets favourable for reactivation in the paleostress field. There can be a spatial variation in displacements within a constant stress state, producing strain gradients leading to non-coaxiality.

Although the regional stress field might be relatively constant [Zoback, 1992], stresses in rocks between faults are not always uniform due to this fault interaction, leading to local deviations in paleostress orientations and the existence of local transient paleostress regimes. Slickensides mainly define the directions of bulk shortening and elongation for discontinuous (frictional) deformation. The deduced paleostress directions therefore reflect the principal directions of shortening and elongation for deformation increments on individual fault sets.

Thus, In the context of the micro-structural analysis of fault slip data, we propose to use the paleostress reconstructions on individual fracture zones to be used as an estimator of the local 3D displacement field in the fracture zone, which can be used to estimate the fault kinematics of the related macro-structure, for which direct strain or kinematic observations are often obstructed by a high erosion of the fault, soil cover, vegetation,... Regarding the principal stress directions inferred from fault-slip movements as indicators for the local velocity (displacement) field, could create a conceptual tool in the kinematic analysis of brittle tectonic movements and upper crustal deformation.

In the present study, homogenous tectonic regimes (directions of principal stresses) could be attributed to many faults in the studied Altai region (see Part II), conform to their style of faulting (deep brittle versus shallow (mechanic, brecciated). Their relative timing and tectonic style are also recognized and confirmed by seismic and geomorphologic analysis, suggesting their realistic nature...

Thus, careful use of micro-structural data, distinguishing different faulting phases, can contribute to local tectonic reconstructions, which can be extended comparing the tectonic styles for a given period of adjacent regions.

Distinction of faulting phases should be made on basis of slickenside style, ranging from semi-ductile mineral lineations over cataclasis, pseudotachylyte formation to chlorite/epidote-coated faults, iron-coated (hematite) planes, ridge-in-groove lineations, and ultimately superficial movements with clay-gauge and breccia fill. In that way, kinematic analysis of the different faulting events can be useful for a relative reconstruction of the different tectonic events in the region.