Chapter 16

Towards a Genetic Classification of Fault Rocks: Geological Usage and Tectonophysical Implications

S. M. SCHMID and M. R. HANDY

ABSTRACT

A simple genetic classification of fault rocks into cataclastic and mylonitic fault rocks is proposed. The fundamental criterion underlying this major subdivision is the transition from frictional to viscous flow. Frictional flow includes those deformation mechanisms that involve fracture and dilatancy and that are shown experimentally to depend strongly on effective pressure. Viscous flow encompasses non-frictional, thermally activated deformation mechanisms. The classification of a fault rock is based on the identification of the mineral(s) and deformation mechanism(s) that accommodated most of the strain and are inferred to have controlled the rheology of the rock. It is emphasized that this genetic fault rock classification is complementary to petrological rock classification. The frictional to viscous transition represents a first-order microstructural and mechanical discontinuity in the lithosphere and is expected to occur over a progressively narrower depth-range with increasing strain. This transition occurs below the onset of crystal plasticity, within a broader zone of transition from unstable (seismic) to stable or conditionally stable (aseismic) slip.

INTRODUCTION

Fault rocks (Sibson, 1977) or fault-related rocks (Wise *et al.*, 1984) form as the result of strain concentration within a tabular or planar zone. This strain concentration may involve substantial displacement of undeformed or less deformed wall rocks across the planar zone. There is a wide range of names used for such high strain zones such as ductile shear zone, mylonite belt, ductile fault zone, gouge zone, or shatter zone. All these names imply a finite width of what may be considered a planar discontinuity at a much larger scale.

If geologists or geophysicists wish to be more specific about the character of the fault rocks within a fault zone, then they need a well-defined terminology and several attempts to provide such a terminology have been made in the past (Table 16.1). The topic of fault rock classification is and will remain highly controversial because there are differing views about the purpose of such a classification in the

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first place. Some people seek a strictly non-genetic classification based on descriptive criteria which do not require a deeper understanding of deformational processes involved. The clear advantages of such classifications are that (i) the non-specialist can readily use the classification, and, (ii) that the classification need not be continuously changed as our understanding of processes improves. The trouble with existing non-genetic classifications is that they incorporate genetic terms (e.g. cataclasis, cataclasites) or use terms that have acquired a genetic meaning (e.g. mylonite; Higgins, 1971). Such hidden genetic connotations betray the desire to employ genetically meaningful criteria, even though this is rarely admitted. As discussed below, another major problem with non-genetic fault rock classifications is that purely descriptive criteria do not suffice to classify the entire spectrum of fault rocks. While we agree that debates and controversies on nomenclature tend to be uninteresting, we feel that a better consensus in the nomenclature of fault rocks is needed unless we are willing to remain unspecific or to include long descriptions in map legends and papers not primarily dealing with processes and conditions of deformation in fault zones.

The criteria presently used for classifying fault rocks range from purely descriptive (microfabric; e.g. Bell and Etheridge, 1973; Heitzmann, 1985) to genetic (rate of recovery and strain-rate; Wise *et al.*, 1984). No general consensus was reached at a Penrose conference devoted to mylonites, although most participants agreed that criteria 1, 2 and 3 (Table 16.1) are the most significant (Tullis *et al.*, 1982). Foliation (criterion 1 in Table 16.1) was originally thought to be restricted to mylonites (Sibson, 1977), and so was regarded as an excellent descriptive criterion for distinguishing cataclasites and mylonites. Since then, Chester *et al.* (1985) have shown that foliation involving rigid-body rotation and alignment of platy fragments also develops in cataclastic fault rocks at shallow crustal levels. Therefore, foliation only remains a valid criterion for distinguishing cataclasites and mylonites if the mechanisms of foliation development are taken into account.

The reduction of grainsize (criterion 2 in Table 16.1) is another problematic non-genetic characteristic of many, but not all, fault rocks. For a long time, grainsize reduction by cataclasis was the only deformation mechanism envisaged for fault rock genesis (e.g. Higgins, 1971). The experimental discovery of grainsize reduction by syntectonic recrystallization during crystal plastic deformation (Christie *et al.*,

Table 16.1 Criteria used by various authors for fault rock classification.

	References					
Criteria	(1)	(2)	(3)	(4)	(5)	(6)
1. Foliated-unfoliated	XX	X	XX	X	XX	
2. Grainsize reduction		X		X	X	
3. Strain localization	XX		XX			
4. Brittle vs. ductile			X		X	
5. Deformation mechanism	X	XX	X	X		XX
6. Rate of recovery and strain-rate				XX		X

XX – crucial for classification; X – other criteria used.
(1) Bell and Etheridge (1973).
(2) Sibson (1977).
(3) White *et al.* (1980); White (1982).
(4) Wise *et al.* (1984).
(5) Heitzmann (1985).
(6) This chapter.

1964) took an amazingly long time to reach the wider geologic community. This discovery introduced the need for microstructural criteria to distinguish cataclasis from dynamic recrystallization. Grainsize reduction by syntectonic recrystallization is indeed common in many mylonites deformed at relatively high differential stresses and low homologous temperatures, where subgrain rotation recrystallization dominates. However, deformation at high homologous temperatures promotes grain growth associated with syntectonic grain-boundary migration recrystallization, as observed in rocksalt (Guillope and Poirier, 1979), ice (Huddleston, 1980), calcite (Schmid *et al.*, 1987), quartz (Lister and Snoke, 1984), and olivine (Nicolas and Poirier, 1976). As long as we continue to believe that grainsize reduction always accompanies mylonitization, mylonitic fault rocks that exhibit syntectonic grain growth will not be recognized as such. Finally, strain concentration (criterion 3 in Table 16.1) is obviously a characteristic of all fault rocks and so is not a satisfactory criterion for classification.

Having illustrated the serious problems with all three descriptive criteria generally considered to be most important, we are left only with genetic criteria (criteria 4 to 6 in Table 16.1) to distinguish and classify fault rocks. For this reason, and because existing classifications contain hidden genetic connotations anyway, we propose a genetic classification for fault rocks. We feel that such an attempt is justified in light of the progress made in recent years on inferring conditions and mechanisms of deformation from microstructural studies of experimentally and naturally deformed rocks. A genetic classification not only names rocks, it also enables us to infer deformation mechanisms and conditions of deformation. Because this contribution emphasizes the tectonophysical implications of a genetic classification, concepts such as the brittle to ductile transition and switches in the active deformation mechanisms and rheology will be discussed first. Some use of specialized terms is unavoidable in our discussion, so a glossary of these terms is provided at the end of the paper to aid the uninitiated reader.

THE BRITTLE–DUCTILE TRANSITION

Imprecise use of the terms "brittle" and "ductile" still creates a lot of confusion. The following definitions closely follow the excellent discussion by Rutter (1986). Ductility is a measure for the capacity of a material to deform by uniformly distributed flow. Of course, flow may be uniform on one scale and non-uniform or localized on another scale. Consider, for example, a fault zone that offsets an undeformed parent rock. The entire system (i.e. fault and parent rock) exhibits localized flow, the fault rock itself deforms ductily, whereas on the scale of grains within the fault rock, the flow may be considered localized again (e.g. grain boundary sliding in cataclasites or very fine-grained mylonites). So irrespective of the deformation mechanism, all fault rocks are ductile if the scale of observation is chosen appropriately. Therefore, ductility is not a mechanistic concept. In contrast, brittle or brittleness are both terms that are associated with the formation of cracks and so clearly reflect a mechanistic concept.

Unfortunately for the field geologist, the rock-mechanics literature introduced the concept of a brittle–ductile transition. This term combines the mechanistic concept of brittleness with the non-mechanistic concept of ductility. Further confusion arises because many experimentalists implicitly or explicitly attach a mechanistic significance to the term ductile (for example, see Handin, 1966, p. 226). In the laboratory, brittle deformation is associated with a stress drop from ultimate to residual strength after negligible amounts of non-elastic deformation. The strain in the sample is concentrated along a discrete fracture. Towards higher temperatures and/or confining pressures, materials undergo increasing amounts of permanent strain before they fail and a stress drop occurs. At these conditions, deformation becomes more uniformly distributed (ductile) on the scale of the specimen. Increased ductility in this dual microstructural and mechanical sense (i.e. increasing permanent strain prior to a stress drop and increasing homogeneity of deformation) represents a transition from discrete fracturing (brittle in the sense of the experimentalist) to cataclastic flow associated with brittle microcracking (brittle in a mechanistic sense only), and/or to crystal plasticity (Mogi, 1974).

Clearly, the terms brittle and ductile are of very limited use or even useless for the field geologist when employed in the dual sense of the rock-mechanics literature because the stress–strain history of naturally deformed rock is difficult if not impossible to reconstruct. What is fundamental for a genetic classification is the distinction made below between deformation mechanisms that are associated with frictional and viscous mechanical behaviour.

DEFORMATION MODES AND THE FRICTIONAL-VISCOUS TRANSITION

Cataclasis is a deformational process involving initial granulation of grains by microcracking, leading to frictional sliding, dilatancy, and rigid-body rotation among grain fragments, grains, or groups of grains. Generally, both the mean grainsize and the degree of grain sorting decrease with strain during cataclastic flow (Engelder, 1974). In the laboratory, grain sliding is frictional (i.e. strongly dependent on normal stress) and dilatancy (i.e. volume change) is strongly pressure-sensitive. Thus, the strength of a cataclastically deforming rock depends primarily on the effective pressure (the difference between lithostatic and fluid pressures) and hence also on depth and fluid pressure. Temperature and strain-rate are of secondary importance. The frictional strength of rocks is largely independent of rock composition (Byerlee, 1978), although fine-grained gouge produced in zones of intense cataclasis typically displays a lower coefficient of friction than does a clean (i.e. gouge-free) sliding surface (Shimamoto and Logan, 1981). Based on these experimental observations, cataclasis is classified as a frictional mode of deformation.

Viscous creep behaviour is characteristic of all known, thermally activated deformation mechanisms (see Schmid, 1982 for a brief introduction and Poirier, 1985 for a more complete treatment). This group of deformation mechanisms comprises dislocation glide and dislocation creep, solid-state diffusional creep (Nabarro–Herring or Coble creep), diffusional mass transfer involving solution-precipitation via fluids (pressure solution creep) and viscous (i.e. non-frictional) grain boundary sliding (a mechanism associated with superplasticity). Temperature, strain-rate, and often also grainsize control rock strength while effective pressure is of minor importance. The relationship of stress to strain-rate can be linear (Newtonian), non-linear (power law), or exponential. Sibson (1977) collectively referred to these non-cataclastic mechanisms as quasi-plastic whereas Scholz (1988) called them plastic. The reason to avoid the term plastic is because this normally only implies crystal plasticity, and refers specifically to dislocation creep when used in connection with rheological stratification (e.g. Ranalli and Murphy, 1987; Ord and Hobbs, 1989).

What really ought to interest geologists and geophysicists is the structural expression of the transition from frictional to viscous modes of deformation. This transition is important because it is associated with a complex interplay of various deformation mechanisms and, more importantly, coincides with rheological instabilities that mark the base of the crustal seismogenic zone (e.g. Sibson, 1982).

How can we recognize the frictional–viscous transition in exhumed fault rocks? To answer this, we need microstructural criteria for determining which mineral(s) and deformation mechanism(s) governed the flow of a rock during deformation. This is quite easy in monomineralic rocks if the deformation is homogeneous on the scale of observation. We simply identify the deformation mechanisms according to the criteria discussed in the next three sections. Problems arise if the deformation is localized on the observational scale. This is almost always the case in polymineralic rocks where the constituent minerals have contrasting flow strengths.

The solution to this problem lies in distinguishing the deformation mechanism within the interconnected matrix from the mechanisms in the less-deformed minerals or grain aggregates that form clasts or boudins. Investigations of experimentally (e.g. Jordan, 1987) and naturally deformed bimineralic aggregates (Handy, 1990) indicate that the weakest phase controls the rheology of a rock, even where present in low volumetric proportions (<20%), provided that the strain is sufficiently large to produce a foliation in which the weaker phase is interconnected (domains 2 and 3 in Fig. 16.1). The strong phase(s) can deform by other mechanisms than the matrix, but these mechanisms usually accommodate local strain incompatibilities and so do not significantly influence the rheology of the rock. The reader is referred to Handy (1990) for a more comprehensive discussion of solid-state flow in polymineralic rocks.

By analogy with experiments, therefore, the interconnective matrix phase(s) in naturally deformed rocks accommodate most of the strain and are inferred to have governed the rheology of the rock. If this matrix phase deforms by cataclasis, then the strength of the rock at the time of deformation is described by constitutive equations for frictional sliding. On the other hand, if the matrix phase shows signs of syntectonic recovery and syntectonic recrystallization (which is the case in many mylonites), then the strength of such a mylonitic rock is adequately described by constitutive equations for nonlinear viscous creep. Similar argumentation pertains to fault rocks with other deformation mechanisms in the matrix phase (e.g. grainsize-sensitive creep) and leads to different mechanical inferences.

When this method of determining the rheologically dominant deformation mechanism is applied to microstructures in naturally and experimentally deformed rocks, it becomes apparent that the frictional–viscous transition occurs over a progressively narrower depth-interval with strain. Cataclastic flow predominates to high strains in rocks comprising strain-hardening minerals that are present in volumetric proportions of 80% or more. In contrast, rocks that contain 20 volume % or more of a strain-softening, viscously deforming phase show a strain-dependent switch from frictional to viscous behaviour. In low-strain experimental deformation of such rocks, cataclasis and crystal plasticity occur simultaneously (semibrittle behaviour, Carter and Kirby, 1978) and transient creep depends strongly on effective pressure, temperature and strain-rate. However, much higher strains accrue in natural fault rocks and their microstructures indicate that continued cataclasis eventually results in a matrix grainsize which is sufficiently small to induce grainsize-sensitive creep (Mitra, 1984; Wojtal and Mitra, 1986). Pressure solution in a fluid at the grain boundaries can accompany cataclasis and experiments have shown that this can lead to linear viscous creep behaviour (e.g. slow creep experiments of Rutter and White, 1979).

The main message we wish to convey here is that the transition from frictional to viscous flow occurs over a narrow depth range for fault rocks accumulating large

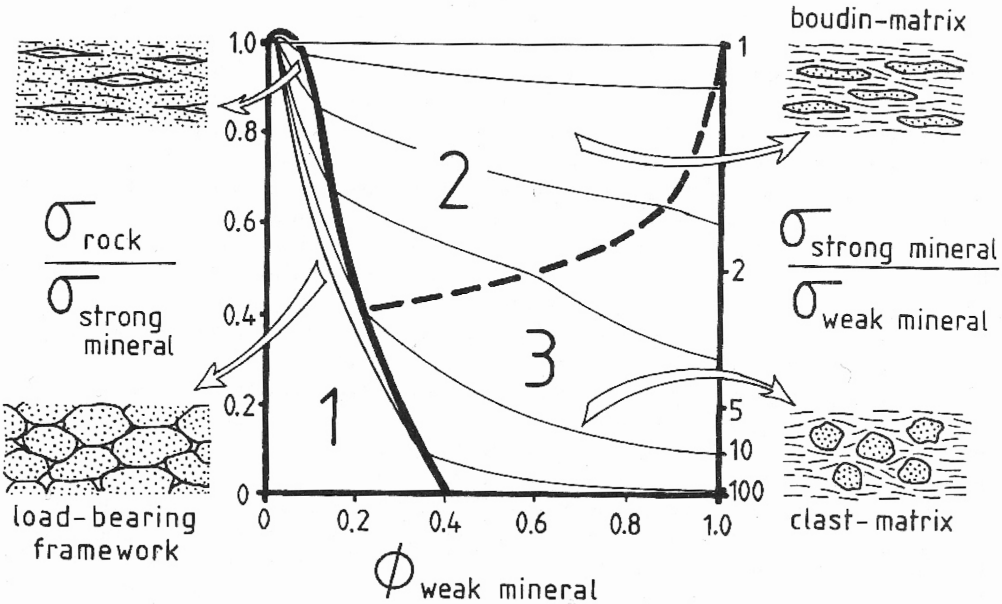


Figure 16.1 Diagram modified from Handy (1990) showing the strength of a bimineralic rock as a function of the volume proportion of the weak mineral (ϕ_w). The left-hand vertical axis indicates the normalized rock strength (strength of the stronger mineral set equal to one). The thick curve separates the load-bearing framework rheology (domain 1) from the matrix-controlled rheologies (domains 2 and 3). The thick dashed curve represents the gradational transition between boudin–matrix and clast–matrix rheologies (respectively, domains 2 and 3). The thin curves are contours of normalized rock strength at different mineral strength contrasts. They indicate the compositional dependence of rock strength for different mineral strength ratios shown on the right hand vertical axis. Arrows point to the dynamic steady-state microstructure for each domain (stippled: strong mineral; unstippled: weak mineral; lines: (foliation). This diagram is based on microstructural and mechanical evidence from experimentally and naturally deformed rocks (see Handy, 1990 for further explanation).

strains, whereas this transition will be much broader at small strains. This frictional–viscous transition is of prime rheological importance and is reflected in the micro-structural appearance of fault rocks. Table 16.2 summarizes the relationship of the fault rock fabric and inferred deformation mechanisms to the conditions of deformation.

MYLONITES

Definition and the genesis of mylonites

Mylonites are foliated rocks in shear zones, in which most or all of the strain occurs within minerals that show evidence of viscous deformation (Table 16.2). The criteria for recognizing various deformation mechanisms associated with viscous creep are discussed below in the second part of this section (see also Knipe, 1989). In polymineralic mylonites, the weakest phase(s) accommodate the bulk strain if these phases form an interconnected matrix. Stronger phases form boudins or clasts depending on the strength contrast between strong and weak phases (domains

Table 16.2 Summary of proposed classification and tectonophysical implications.

	Cataclastic fault rocks		Mylonites, mylonitic fault rocks
	Fault breccia Fault gouge	Cataclasite	
General nature of deformation mechanism	Frictional (brittle and dilatant)		Viscous in minerals that govern bulk rheology: matrix mineral(s) in domains 2 and 3, and load-bearing framework minerals in domain 1 of Fig

Weaker phase(s)
dislocation creep
mechanisms

2 and 3 in Fig. 16.1; Handy, 1990). More rarely, the stronger phase(s) form a viscously deforming, load-bearing framework (mineral strength ratios of less than 10:1 in domain 1 of Fig. 16.1). In all mylonites, a penetrative foliation develops after moderate to large strains and stretching lineations are frequently, but not always, observed. Sometimes, the foliation is only detectable under the microscope (e.g. flinty ultramylonites).

Stretching lineations in most mylonites form as a result of rotational (i.e. noncoaxial) strain combined with viscous strength contrasts between the constituent minerals or monomineralic aggregates (Lister and Price, 1978). When the shape of the strain ellipsoid is prolate the more competent minerals or grain aggregates rotate and streak out into parallelism with the direction of shear. However, stretching lineations are weak or non-existent if the strain ellipsoid is strongly oblate, the viscosity contrast among the constituent minerals or grain aggregates was small, or if post-tectonic annealing was sufficiently strong to obliterate the mylonitic microstructure (see below).

Mineral reactions during mylonitization can lead to significant changes in the volume proportions, relative strengths, deformation mechanisms, and the types of minerals, and so exert a strong influence on both microstructure and rheology (review in Brodie and Rutter, 1985). Syntectonic mineral reactions may accompany prograde or retrograde regional metamorphism outside of the shear zone, or they may be confined to the shear zone for kinetic or fluid permeability reasons. In the latter case, the mylonite is usually easy to distinguish from its protolith because of its contrasting mineralogy and grainsize. The strain-localization is then said to be reaction-enhanced (White and Knipe, 1978).

The causes of strain localization leading to mylonitization are currently the subject of much debate, but strain localization in many natural settings is probably related to initial rheological instabilities along pre-existing structural or lithological discontinuities in the protolith. The localized microstructural changes associated with incipient unstable deformation induce strain-softening and amplify strain-localization (White *et al.*, 1980). The positive-feedback nature of the localization process therefore leads to overprinting of the early strain history by the subsequent strain-softening deformation. If the localization involved initial strain-hardening, then the shear zone may have broadened prior to the onset of strain-softening. In this case, structural evidence of early strain-hardening may be preserved in the country rock adjacent to shear zones or within clasts in the mylonite.

We point out that our definition of a mylonite not only includes the typical mesoscopic features of mylonites recognized by Bell and Etheridge (1973), it also accounts for all other known microstructural effects of the identifiable deformation mechanisms in the rock (e.g. syntectonic grainsize reduction and syntectonic grain growth in the case of crystal plasticity). Interestingly enough, it was Lapworth (1885), one of the pioneer mappers of fault rocks, who first recognized the possibility that mylonitization involves viscous (plastic) behaviour in some of the minerals (review in White, 1982). Past classifications of mylonites into non-mylonitic tectonites, proto, ortho- and ultramylonites (Sibson, 1977; White, 1982; Wise *et al.*, 1984; Heitzmann, 1985) are only applicable to a limited range of lithologies and deformational conditions (typically, greenschist facies quartzo-feldspathic mylonites, in which syntectonic recrystallization and mineral reactions are associated with grainsize reduction). A descriptive classification of mylonites that are derived from all kinds of protoliths and are deformed under the full range of physical conditions rapidly becomes a semantic exercise and, in our opinion, leaves us further from understanding the processes of mylonitization than does a classification which is based on the identification of rheologically dominant deformation mechanisms.

Diagnostic features of mylonites

There are two complementary approaches for recognizing mylonites. First, the existence of a mylonite belt or crustal-scale shear zone in the field is established by tracing the undeformed or weakly deformed lithologies (protoliths) into the highly deformed areas. The amount of deformation needed to cause mylonitization depends on the magnitude of the pre-existing strain in the wall rocks. So when a fault rock is referred to as a mylonite, this sometimes needs to be qualified with a description of the strain gradient into the shear zone. This is particularly important in high-grade metamorphic terrains containing schists or gneisses that have undergone complex deformational histories. In the case of a crustal-scale mylonite belt, an area of several square kilometres must be mapped and studied, otherwise individual specimens and outcrops cannot be interpreted in the context of regional strain gradients. Second, microstructural studies, preferably on oriented samples, must indicate that viscous flow was responsible for deformation within the matrix of the rock. While it is often difficult to identify the exact deformation mechanism(s), it generally suffices to show that cataclasis was not the dominant mechanism (see criteria for cataclasis in the next section). Microstructural criteria for recognizing the various deformation mechanisms associated with viscous behaviour are described below and summarized in Table 16.3.

Dislocation glide and dislocation creep

These deformation mechanisms are referred to collectively as crystal plastic (or intracrystalline plastic) because strain is produced by the movement of dislocations within individual crystals or grains. At low strains and low homologous temperatures, dislocations move or glide within crystallographic planes. Undulose extinction, deformation bands, kink bands and deformation twins are common microscopic features of low-strain crystal plasticity. However, at higher strains and homologous temperatures, dislocations climb between different crystallographic planes within the lattice (dislocation creep). The climb of dislocations is a necessary condition for large strains to accrue. Thus, syntectonic recovery (indicated by microscopically visible subgrains) and syntectonic recrystallization are diagnostic of high-strain crystal plasticity, not the observation of intracrystalline optical strain features alone. After large strains, crystal plastic deformation always produces a strong crystallographic preferred orientation. Syntectonic recrystallization is usually observed to enhance this preferred orientation (e.g. Schmid and Casey, 1986).

Syntectonic recrystallization involving subgrain rotation often leads to a decrease in grainsize compared to the protolith (Fig. 16.2a). At higher homologous temperatures, syntectonic recrystallization primarily involves grain boundary migration and so potentially leads to an increase in grainsize (Fig. 16.2b). Syntectonically recrystallized grainsize depends largely on stress (and indirectly also on temperature and strain-rate). Thus grainsize will increase or decrease depending on initial grainsize and on the differential flow stress. Static recrystallization may lead to further grain growth after deformation (discussion below).

Diffusion creep and pressure-solution creep

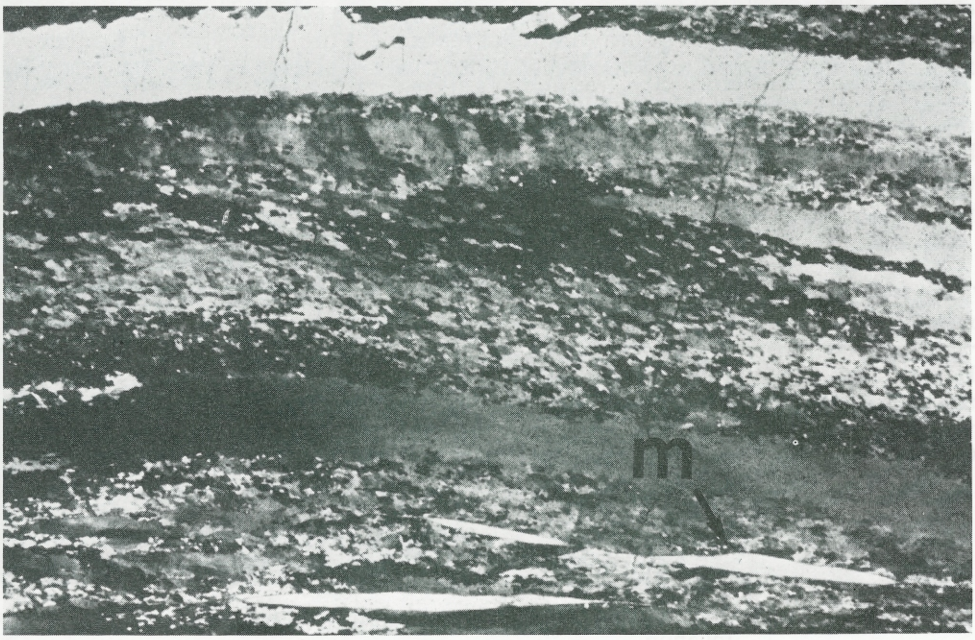
These mechanisms are often referred to collectively as grainsize-sensitive creep. They accommodate most of the strain in the rock when the grainsize is sufficiently small for diffusive mass transfer processes acting within grains or at grain boundaries to be rate-competitive with dislocation mobility within grains (crystal plasticity). In experiments, decreased grainsize dramatically increases the strain-rate at constant stress or leads to large stress drops at constant strain-rate (strain-rate proportional to d^{-2} or d^{-3}, where d is the grain diameter). Additional factors like fluid permeability and the rock- and fluid-composition are important, particularly for pressure-solution creep. Small grainsizes may result from syntectonic recrystallization, cataclasis, or syntectonic reaction. Examination of strain gradients and mineralogy within a thin section or on a larger scale usually reveals which is the case (e.g. fig. 10 in Handy, 1989). If grain growth is inhibited during mylonitization, strain softening may result from such transitions to grainsize-sensitive creep (e.g. Schmid, 1982). In the particular case of syntectonic metamorphic reactions, a feedback relationship between localized reactions and localized deformation (often referred to as reaction-enhanced ductility; White and Knipe, 1978) can induce a transition to grainsize-sensitive creep. Alternatively, the metamorphic reactions may simply produce minerals which are inherently more deformable by dislocation creep.

Weak or non-existent crystallographic preferred orientation in the absence of grain growth in fine-grained aggregates is indicative of grainsize-sensitive creep. However, directly inferring the existence of grainsize-sensitive creep in naturally deformed rocks is often very difficult (see Behrmann, 1985), if not impossible. Evidence of grainsize-sensitive creep sometimes comes indirectly from the observation of displacements along fine-grained microshears derived from syntectonically recrystallizing porphyroclasts (SP-mylonites of Boullier and Guegin, 1975).

THE TRANSFORMATION OF MYLONITES INTO GNEISSES AND SCHISTS

Up to this point, we have discussed the microstructural and mechanical effects of mylonitization on undeformed or less-deformed rocks. Often, however, high temperature conditions outlast deformation, resulting in static recovery and recrystallization, and associated grain growth (processes collectively referred to as annealing; see Glossary). Annealing obliterates many of the diagnostic strain features of

Figure 16.2 (a) Syntectonic recrystallization involving predominantly the progressive rotation of subgrains leads to grainsize reduction in this quartz mylonite. The old grains are flattened into ribbons that define a foliation parallel to the macroscopic foliation, marked by mica flakes (m). The new, syntectonically recrystallized grains are comparatively small (ca. 60 μm) and display serrate, non-linear boundaries. The recrystallized grains define a shape fabric (NW–SE oriented) oblique to the macroscopically visible foliation and leaning in the direction of shear. The crystallographic preferred orientation is strong. Greenschist facies mylonite taken from the footwall of an exhumed low-angle extensional fault in the Ivrea Zone, Southern Alps (Handy, 1987). Thin section cut parallel to the stretching lineation and perpendicular to the schistosity (frame dimensions: 9 × 6 mm). (b) Syntectonic recrystallization by grain boundary migration leads to grain growth in this quartz mylonite. The grain boundaries are extremely lobate, indicating that this mylonite did not undergo post-tectonic (i.e. static) recrystallization (see text and Fig. 16.3 for comparison). As in (a) above, an extremely strong crystallographic preferred orientation and the preservation of a grain boundary fabric (NW–SE oriented) oblique to the macroscopically visible foliation (EW), marked by mica flakes (m), indicates the dynamic (i.e. syntectonic) nature of this microstructure. Mylonite taken from an originally deep part of a major thrust in the Eastern Alps (Schmid and Haas, 1989), where deformation occurred under amphibolite facies conditions. Thin section orientation and frame dimensions are as in (a).



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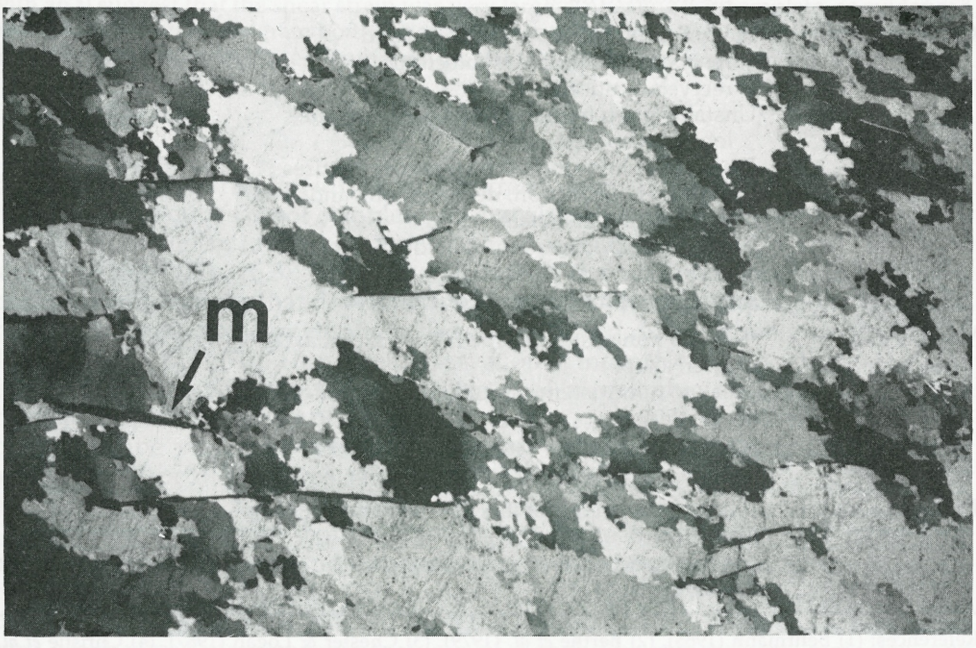

(*b*)

mylonitization and makes it difficult to recognize mylonites. Where the post-tectonic overprint is very strong, a mylonite may no longer be identified as such and is completely transformed into a fine-grained gneiss or a schist. The former existence of a mylonite may still be inferred if relict features of the microstructure or curved foliation patterns indicate strain-localization.

Distinguishing static recrystallization from dynamic recrystallization is difficult in mylonites that were deformed at high homologous temperatures because grain boundary migration can occur both syn- and post-tectonically (e.g. White, 1977). The typical features of dynamic recrystallization have been described above. Annealed microfabrics contain grains with straight, equilibrated boundaries, low dislocation

Table 16.3 Microstructural features used to identify deformation mechanisms by optical microscopy.

Deformation mechanism	Microstructural criteria	References
Cataclasis	Angular fractured grains, grain aggregates, or grain fragments Microcracks (intergranular, intragranular, and transgranular) Discrete sliding surfaces parallel and oblique to the shearing plane Stringers and trails, sometimes defining a crude foliation Glass or ultrafine grained groundmass (in pseudotachylites)	3,9,13,15,17
Crystal plasticity	Subgrains (syntectonic recovery) and recrystallized grains (syntectonic recrystallization) Unstable grain boundaries (jagged, arcuate, lobate–cuspate) Frequently, low strain intragranular features (undulose extinction, deformation bands and deformation lamellae) Foliation, sometimes multiple foliations and shape fabrics (S–C fabrics) Crystallographic preferred orientation	2,4,10,11,12,16,18
Viscous grain-boundary sliding	Weak to nonexistent shape fabric (± equiaxed grain shape) Weak to nonexistent crystallographic preferred orientation Straight to slightly arcuate grain boundaries Usually, very fine grainsize	1,7,16,19
Pressure-solution	Precipitation overgrowths in voids and pressure shadows Indented and sutured grain contacts Stylolites	5,6,8,14

References: (1) Behrmann (1985), (2) Berthé *et al.* (1979), (3) Chester & Logan (1987), (4) Christie *et al.* (1964), (5) De Boer *et al.* (1977), (6) Durney (1976), (7) Edington *et al.* (1976), (8) Elliot (1973), (9) Engelder (1974), (10) Gapais & White (1982), (11) Groshong (1988), (12) Lister & Snoke (1984), (13) Mitra (1984), (14) Rutter (1983), (15) Rutter *et al.* (1986), (16) Schmid *et al.* (1987), (17) Sibson (1975), (18) White (1976), (19) White (1977).

Figure 16.3 Post-tectonic grain growth has eliminated most direct microstructural evidence of previous viscous creep in this quartz-rich gneiss (compare with Fig. 16.2a and b). All traces of a shape fabric in quartz are gone, even though the boudinaged feldspar aggregates (marked with an f) suggest that the shear strain was very large. Annealing is incomplete, as evidenced by the locally non-linear grain boundaries and irregular grainsize. Amphibolite facies gneiss from the Strona–Ceneri Zone, Southern Alps (Handy, 1987). Frame dimensions: 9 × 6 mm.

densities, and few traces of intragranular strain. Often, post-tectonic equilibration of the grain boundaries is not complete, so that although grain boundaries can appear somewhat nonlinear (a characteristic of dynamic recrystallization), in detail they are locally straight (Fig. 16.3). Annealing also tends to weaken strong crystallographic preferred orientations that developed during mylonitization, although experiments have shown that some orientations are strengthened during grain growth (Hobbs, 1968; Green *et al.*, 1970). Thus, the preservation of incomplete fabric girdles is a good criterion for inferring the former existence of a mylonitic microstructure. If the crystallographic preferred orientation in a rock is completely obliterated, then the oblique shape preferred orientation of platy or acicular minerals (e.g. micas or amphiboles), or of fractures within these phases, is another indicator of earlier mylonitization (e.g. fig. 10d in Lister and Snoke, 1984).

Deformed rocks have both a petrological name and a structural name (e.g. muscovite–chlorite schist and fine-grained metabasic gneiss, respectively, for granitic and metabasic mylonites). Our nomenclature for fault rocks addresses the active deformation mechanisms inferred from the microstructure. Petrological nomenclature emphasizes metamorphic features of a rock like mineralogy, grainsize, and structure (called texture in the petrological literature). These features may be pre-, syn- or post-tectonic. When used in the sense above, the two nomenclatures are complementary and we see no difficulty in calling the same rock a mylonite and a schist/gneiss in the field. To avoid confusion, however, we recommend that schists and gneisses be described as mylonitic if they contain microstructural evidence that viscous creep mechanisms controlled the flow of the rock. Conversely, the terms mylonite or mylonitic should be avoided once direct microstructural evidence for the deformation mechanisms has been destroyed by static recrystallization and associated grain growth.

CATACLASTIC FAULT ROCKS

Definitions and the cohesiveness of cataclastic fault rocks

Cataclastic fault rocks are fault rocks in which cataclasis is the dominant deformation mechanism. Cataclasis has been defined above and its microstructural characteristics are described below. Outcrops of cataclastic fault rocks are variably fractured and jointed, and usually are extensively altered along cracks. This damaged and altered rock grades into very narrow, fine-grained zones that transect the outcrop or sometimes even juxtapose different structures and lithologies (e.g. Chester and Logan, 1986).

Following Sibson (1977), we differentiate between cataclastic fault rocks that are cohesionless (fault gouges and fault breccias) and those that are cohesive cataclasites). However, contrary to Sibson (1977), we use the adjectives cohesive and cohesionless to describe the inferred cohesiveness of the fault rocks during faulting and not the cohesive state of the fault rocks when exposed in outcrop. Whereas the latter usage is of prime importance to the engineering geologist, it is of limited value to geologists interested in the conditions at the time of faulting. Cohesionless cataclasis involves a loss of cohesion during faulting among a sufficiently large proportion of the constituent grains for the whole fault rock to lose cohesion. During cohesive cataclasis, cohesion is only lost among a small (but constantly changing) population of grains at a given time, so that the fault rock as a whole maintains cohesion. The cohesiveness of a cataclastically deforming fault rock depends strongly on effective normal stress. Obviously, the interaction of other factors such as the mineralogy and the syntectonic circulation of fluids can also influence cohesion during cataclasis.

Cataclasites sometimes show evidence of limited and localized dislocation glide (e.g. undulose extinction, kinking, twinning), pressure-solution (precipitation overgrowths), or even melting (glass). These mechanisms act primarily to accommodate strain and maintain cohesion during cataclastic flow. If a foliation is present in the matrix (Fig. 16.4a), then careful microscopic work is necessary to establish whether cataclasis (rather than crystal plasticity or pressure-solution) is the dominant deformation mechanism.

Fault gouge (i.e. rock flour, powder, or paste) and tectonic breccia are products of cohesionless cataclasis, but syn- or post-tectonic cementation associated with infiltrating hydrothermal fluids sometimes causes these fault rocks to regain cohesion (hence the important distinction between Sibson's use of "cohesive" and ours). Late hydrothermal cementation is usually characterized in thin section by microcrystalline matrix groundmass or by free-growth crystals occupying former cavities in the fault rock (Stel, 1981). Figure 16.4b shows that in some cases, more than one cycle of cracking and sealing of the gouge or breccia can be inferred.

We consider pseudotachylites to be cohesionless cataclastic fault rocks, even though they are cohesive when exposed in outcrop. Pseudotachylite commonly occurs in a complex network of narrow veins that truncate structures in the country rock. They are often, though not always spatially and genetically related to cataclasites (Sibson, 1975) and even to mylonites (Passchier, 1982). Evidence for a primary loss of cohesion comes from their often glassy matrix and from the angular to rounded, variably rotated clasts of country rock. Seismic loss of cohesion and local decompression are believed to trigger rapid melting (Sibson, 1975) or vaporization (Wenk, 1978), followed by rapid cooling and solidification of the fluid phase. The matrix may be either amorphous (glassy) or microcystalline. The latter occurs when cooling was slower or when a strong metamorphic or hydrothermal overprint followed cataclasis.

Fault rock stability

The stability or instability of cataclastic flow is another important genetic characteristic of cataclastic fault rocks that may be inferred from their microstructure. Fault stability is strongly dependent on the effective normal stress, the sliding velocity, and the properties of the sliding surfaces (e.g. gouge, asperities, anisotropies; Jackson and Dunn, 1974). Frictional sliding is stable in experiments when the frictional coefficient of a fault surface increases with displacement during a given increment of time. The friction is then said to be velocity strengthening. However, if the frictional coefficient decreases with displacement (velocity weakening) then faulting either becomes unstable or remains conditionally stable, depending on whether the stiffness of the fault rock exceeds or falls below the stiffness of the surrounding intact rock (Dieterich, 1978). Conditionally stable sliding becomes unstable if the sliding velocity decreases or the normal stress increases above a certain critical level. Unstable sliding is characterized by periods of little or no slip (stick) alternating with periods of rapid slip and associated stress drop. Stick-slip is widely believed to cause seismicity on active fault planes in nature, whereas stable or conditionally stable sliding is aseismic (Brace and Byerlee, 1966).

Unstable cataclasis can be inferred from microstructural evidence that crack growth was critical (i.e. unstable and therefore rapid) at the time of faulting. Critical cracks transect or even offset grains and mineral phase boundaries, both in clasts and in the matrix (e.g. figs 3a and 5b in Mitra, 1984; fig. 11 in Wojtal and Mitra, 1986). There is no discernible crystal plasticity associated with these cracks, suggesting that fracture is controlled by pre-existing inhomogeneities on the grain scale (e.g. cracks, cleavage twins, and grain boundaries; Gandhi and Ashby, 1979). Stable microfracturing is associated with subcritical crack growth (Atkinson, 1984). Such cracks are restricted to individual grains or grain-aggregates and terminate at clast-matrix interfaces (figs 3b and 5d in Mitra, 1984). Limited crystal plasticity or pressure-solution may be observed at crack tips or adjacent to the cracks, indicating either intragranular creep fractures or cleavage fracture assisted by plasticity (Gandhi and Ashby, 1979). Thus, although cataclasis remains the dominant deformation mechanism, the onset of local plasticity as a strain-accommodation mechanism coincides with a switch from abrasive to adhesive wear (Scholz, 1988) within the frictional regime. We emphasize that incipient crystal plasticity occurs above the frictional–viscous transition and does not necessarily coincide with a change from cataclastic to mylonitic deformation as proposed by Scholz (1988). The recognition of incipient crystal plasticity ("crystal deformation front" of Groshong, 1988) in cataclastic rocks has great potential for indicating the temperature of cataclasis, because dislocation mobility is highly temperature-dependent at low homologous temperatures.

The basic problem with all microstructural criteria for fault stability lies with the fact that stability is time- and strain-dependent. In addition, cataclastic structures have similar configurations on different scales of observation (Tchalenko, 1970), so that cracks that appear stable on one scale can appear unstable on another. Therefore, the microstructures observed in a cataclastic fault rock may only reflect the local conditions at the moment that the deformation stopped. Establishing whether or not faulting was stable on the outcrop scale or on the scale of an entire cataclastic fault zone requires careful mapping and sampling of a large area, as recommended for mylonite belts in the section above.

Cataclastic fault rocks can also form during unstable viscous shearing (Hobbs *et al.*, 1986). In principle, the mechanical criteria for stability during viscous shearing are analogous to the velocity-dependent stability of frictional faulting, except that transient creep within mylonitic shear zones depends strongly on temperature and strain-rate (Hobbs *et al.*, 1986). Below a critical homologous temperature, the transient stiffness of a mylonite can exceed the elastic stiffness of the country rocks. This results in an unstable cyclic loading history in the shear zone and may account for field observations of mutually overprinting relations among pseudotachylites and mylonites (Passchier, 1982).

TECTONOPHYSICAL IMPLICATIONS

Existing mechanical models of crustal faulting consider the transition from frictional to viscous flow solely in terms of the transition from cataclasis to crystal plasticity (e.g. Scholz, 1988). Accordingly, the frictional–viscous transition is believed to occur within a strain- and time-invariant depth range that is controlled primarily by the geothermal gradient and regional strain-rate. Yet as we pointed out earlier, progressive strain narrows this frictional–viscous transition zone because grainsize reduction via cataclasis in the frictional regime and syntectonic recrystallization in the viscous regime induces a switch to grainsize-sensitive creep mechanisms. Syntectonic metamorphic reactions involving pressure-solution can enhance this process in both the frictional and the viscous flow regimes. This suggests that both the transition from abrasive to adhesive wear and the change from frictional to viscous flow may occur at shallower levels than predicted by Scholz (1988). How much shallower these levels are probably depends on the local and regional fluid flux because fluids

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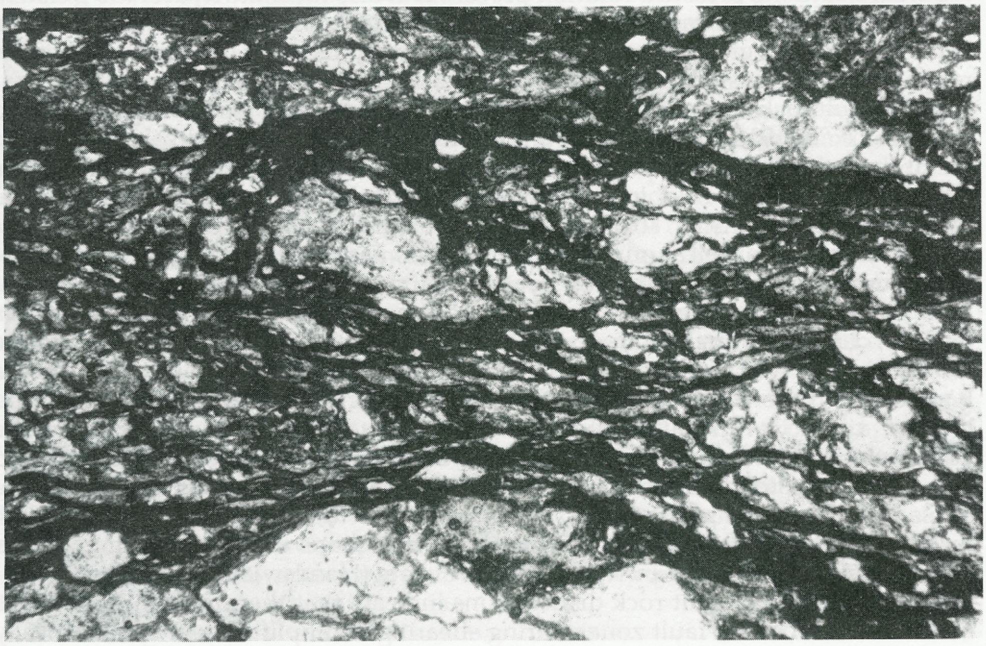
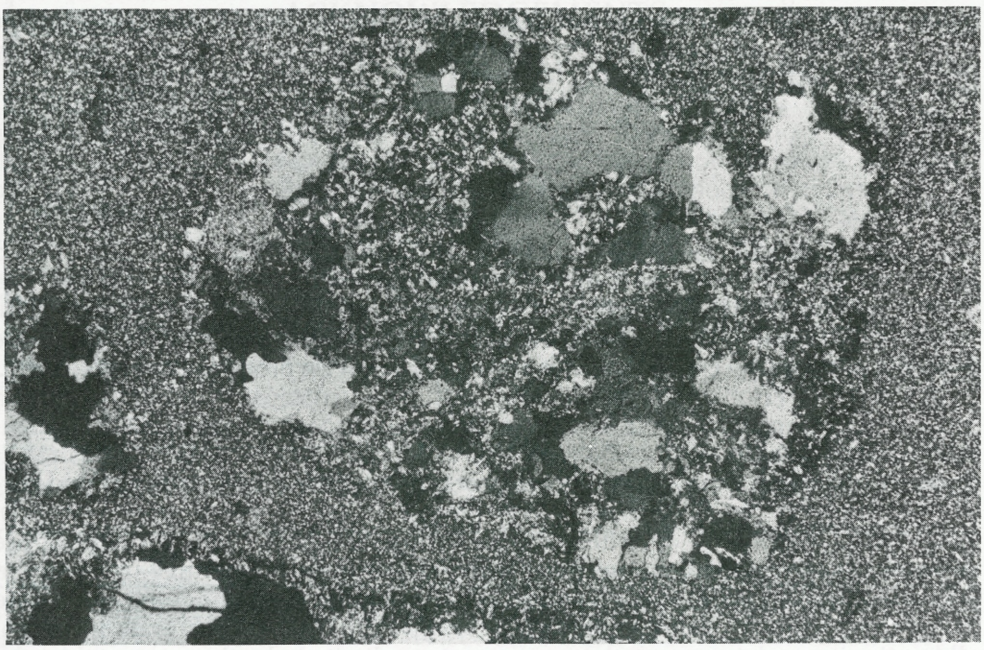
Figure 16.4 (a) Foliated cataclasite derived from an augengneiss. Composite foliation contains Riedel shears. Under higher magnification, quartz locally exhibits undulose extinction and deformation lamellae but no signs of recovery or syntectonic recrystallization. Note the wide spectrum in the size of grains, grain fragments, and grain aggregates. Cataclasis is the predominant deformation mechanism, although pressure-solution may locally act as an accommodation mechanism about the clasts. Specimen taken from the frontal (unmetamorphic) part of the same major thrust as in Fig. 16.2b (Schmid and Haas, 1989). Frame dimensions: 3.5 × 2.3 mm. (b) Fault breccia with two generations of clasts and hydrothermal quartz–illite cement. Note that the large clast in the centre contains smaller quartz clasts that are cemented by an earlier generation of hydrothermal quartz and illite. Thus, repeated loss of cohesion due to unstable cracking and frictional sliding alternated with periods of fluid infiltration, mineral precipitation, and cementation. The final cementation represented by the matrix surrounding the large clast is what gives the rock its present cohesion. Sample taken from borehole in the Kaisten basement rock beneath the tabular Jura Mountains (courtesy of M. Mazurek, Bern). Frame dimensions: 11.5 × 7.7 mm.



(*a*)



(*b*)

are essential to the pressure-solution process and also facilitate hydrolytic weakening of fine-grained silicates (e.g. Jaoul *et al.*, 1984). A possible consequence of this evolution is that a greater proportion of the seismogenic zone lies within the depth range for viscous instability than previously assumed.

Many fault rocks formed during the same shearing event which brought them to the surface. Transitions from mylonitic to cataclastic fault rocks in the field often result from cataclastic overprinting of the mylonites during uplift and cooling. The distribution of fault rocks in ancient, exhumed fault zones thus yields valuable information on the rocks' structural and mechanical evolution during decompression and cooling, as well as on the large-scale kinematics of the fault zone. Generally, stress and strain are progressively localized during uplift as decreasing temperatures and syntectonic grainsize reduction lead to pronounced rheological contrasts (Handy, 1989). Normal faults exhume mylonites and high-grade rocks in the footwall into tectonic contact with cataclasites that are simultaneously forming in the hangingwall whereas thrust faults transport mylonites in the hangingwall into tectonic contact with cataclasites forming in the footwall. In either case, the asymmetrical distribution of fault rocks reflects the fact that the rate of uplift and shearing was faster than the rate of thermal equilibration across the fault zone. The sense of the asymmetry can be used together with sense-of-shear criteria to distinguish normal from thrust faulting, even in regions where the fault zone has been passively reoriented by later tectonic events (Handy, 1987). Fault zones that display symmetrical fault rock distributions indicate that thermal equilibrium was maintained across the fault zones during shearing and uplift. These symmetrically developed fault zones are particularly interesting because the zonation of structures from their margins to their centres offers a sequential view of structural development during shearing. The advantages of a genetic fault rock nomenclature are manifest when one attempts to interpret the mechanical history of such sequentially developed structures in the field.

CONCLUDING REMARKS

Our discussion of the genesis of fault rocks has probably fuelled rather than quelled the long-standing debate on fault-rock terminology. We certainly do not claim to have found definitive solutions to all problems in this controversy. Nevertheless, several conclusions may direct us towards a flexible, usable fault-rock nomenclature that also incorporates current understanding of grain-scale processes during faulting.

A strictly descriptive classification of fault rocks suffers serious deficiencies in light of the recent discoveries of foliation in cataclasites and grain growth in high-temperature mylonites. As an alternative, we propose a genetic classification of fault rocks into cataclastic and mylonitic fault rocks. The fundamental criterion underlying this division is the transition from frictional to viscous flow (Table 16.2). Classifying a fault rock according to this scheme entails an identification of the mineral(s) and deformation mechanism(s) that accommodated most of the strain and therefore controlled the rheology of the rock (Fig. 16.1 and Table 16.3). The frictional–viscous transition represents a first-order microstructural and rheological discontinuity and is expected to occur over a limited depth-range within the lithosphere. We emphasize that this transition occurs below the onset of crystal plasticity at depth.

A basic shortcoming of any genetic classification is that the genetic criteria used in the classification may change as new discoveries are made and the general understanding of deformational processes improves. Admittedly, an unequivocal identification of the rheologically dominant deformation mechanism(s) is not always possible, especially in very fine-grained fault rocks. In such cases, we much prefer to describe the fault rock rather than to impose a genetic name on it.

Beyond any philosophical qualms one might have about genetic classifications in general, our classification scheme contains some unresolved problems and points to several avenues of future research. The genetic subdivision of cataclastic fault rocks into cohesive cataclasites and cohesionless fault breccias and fault gouges is problematic, because the evidence for determining whether a fault rock was cohesive or not during faulting is often ambiguous. The transition from seismic to aseismic deformation is still poorly understood from the mechanistic and microstructural points of view. As discussed above, mechanical instabilities leading to seismicity can occur within both the frictional flow regime or within the viscous flow regime. In the latter case, the cyclic overprinting relationships between cataclastic fault rocks and mylonites are not always easy to distinguish from the structures produced by cataclastic overprinting of mylonites during uplift and cooling across exhumed crustal-scale fault zones. Finally, the transformation of mylonites into gneisses and schists by annealing raises the vexing question of whether nature really shows us everything that occurs at depth. Are there perhaps some deformation mechanisms whose microstructural signature is preserved better than others? Future studies that answer this question will address the problem of microstructural stability.

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GLOSSARY OF SOME TERMS USED IN THE TEXT

Cataclastic flow: A mode of deformation in which fracture, loss of cohesion, and frictional sliding occur on all scales smaller than the width of the fault zone.

Crystal plasticity (or intracrystalline plasticity): Collective term for deformation mechanisms in which the deformation is intragranular and occurs by the movement of line defects, known as dislocations (see dislocation glide and dislocation creep). Dislocation mobility is a thermally activated process.

Diffusion creep: Deformation is achieved by diffusional mass transfer. Solid-state diffusion either occurs within the entire volume of a grain (Nabarro–Herring creep) or is confined to a narrow domain just within the grain boundaries (Coble creep).

Alternatively, diffusional transfer may involve solution into, precipitation from, and transport through an intergranular fluid (pressure-solution creep).

Dislocation creep: Nucleation, glide and climb of dislocations within the crystal lattice, with syntectonic recovery and syntectonic recrystallization (see below) continuously rearranging and eliminating dislocations during deformation. This enables the material to undergo large strains at constant deviatoric stress, because strain-hardening related to the tangling and pile-up of gliding dislocations is balanced by strain-softening induced by the rearrangement and elimination of climbing dislocations. This deformation mechanism is inferred to predominate in power-law creep deformation (non-linear viscous flow regime) at moderate to high homologous temperatures.

Dislocation glide: The nucleation and movement of dislocations within a single crystallographic plane of the crystal lattice during deformation. Dislocation glide is associated with strain-hardening in many materials because dislocations tangle and pile-up and therefore cannot rearrange into lower energy configurations. Consequently, dislocation glide leads to cataclasis at high strains. In experiments, dislocation glide predominates at lower homologous temperatures and/or at higher differential stresses than dislocation creep.

Homologous temperature (T_h): Temperature expressed as a ratio of the ambient temperature to the melting temperature of a material. Homologous temperature ranges from 0 to 1, where $T_h = 1$ is the melting point of a material at a specified pressure. In effect, temperature is normalized to the intrinsic properties of the material, because melting temperature depends strongly on composition and impurity content.

Static recrystallization (annealing): Recrystallization of a previously strained rock involving formation and growth of strain-free grains. During the first stage (primary recrystallization), the driving force is the reduction of strain energy associated with the elimination of dislocations. This results in a polygonal microstructure with straight grain boundaries. During the second stage (secondary recrystallization), the driving force is the reduction of surface energy by grain growth. The statically recrystallized microstructure is often referred to as granoblastic by petrologists.

Strain hardening: The increase in strength of the material with progressive deformation due to an increase in free (unbound) dislocation density. The mobility of free dislocations is limited by their interaction with other dislocations, grain boundaries, impurities, or other obstacles in the crystal lattice. Undulose extinction indicates that dislocations were unable to rearrange into low-energy subgrain walls.

Superplasticity: This term denotes an experimentally determined deformation regime in which viscous grain-boundary sliding is inferred to be the dominant (i.e. main strain-accommodating) deformation mechanism. "Viscous" sliding emphasizes the fact that the sliding is not frictional. Grain boundaries act like strain-producing planar defects, analogous to dislocations acting like strain-producing line defects.

Syntectonic recovery (or dynamic recovery): The rearrangement of dislocations into lower energy subgrain walls (optically visible as subgrain boundaries) during crystal plastic deformation. This process involves dislocation climb and so is favoured at higher temperatures. Like syntectonic recrystallization, syntectonic recovery maintains the density of free (ie. unbound) dislocations at a constant level, thereby facilitating the accumulation of large strains without hardening the material.

Syntectonic recrystallization (or dynamic recrystallization): Recrystallization that accompanies crystal–plastic deformation. The reduction of elastic strain energy associated with the rearrangement and elimination of dislocations is the prime driving force. The density of free dislocations is reduced and/or kept at a constant level by two different processes: (1) the climb of free dislocations into subgrain boundaries, causing increasing crystallographic misfit across subgrain boundaries (subgrain rotation recrystallization); and, (2) the migration of grain boundaries across the material from regions of lower dislocation density into regions of higher dislocation density. These processes are not mutually exclusive and both reduce the dislocation density in grains or parts of grains. Process (1) usually leads to grain-size reduction (Fig. 16.2a). Process (2) leads to either a grainsize decrease or increase, depending on the initial grainsize. In the latter case, the grain boundaries are lobate and cuspate (Fig. 16.2b).

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