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On the development of zones of reverse shearing in mylonitic rocks

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Abstract

Most shear zones contain a minor proportion of domains (recognizable from thin-section to outcrop scale) which show an opposing shear sense relative to the overall kinematic framework. In this paper, we discuss how these universal features result from the evolving deformation/strain partitioning that generally accompanies progressive general shear, principally in low metamorphic grade shear zones. Because the rheologic properties and, consequently, the contributions of the different deformation mechanisms change continuously with progressive strain, an individual domain with a predominant deformation mechanism much faster or slower than the average may depart from the general kinematics, enhancing the viscosity contrast between adjacent layers and causing locally reverse shear in the domain boundaries. Care is needed, therefore, in interpreting opposing shear senses in the same shear zone as indicating superimposed deformation events.

1. Introduction

In the last 20 years, experimental results and investigation of naturally deformed rocks have provided insight into the importance of grain-scale processes and have led to an increased interest in microstructural analysis as a tool for evaluating mylonites and ductile deformation (e.g. Berthé et al., 1979; White et al., 1980; Lister and Snoke, 1984; Knipe, 1989; Michibayashi, 1993; Pauli et al., 1996 among many others). Within this broad field, the search for reliable shear sense indicators at a microscale has become a major research subject (see review by Hanmer and Passchier, 1991), with a number of new indicators having been identified in the last few years (e.g. Hippertt, 1993; Doblas et al., 1997). The most commonly recognized shear-sense indicators include, in rough order of reliability, shear bands (see Simpson and Schmid, 1983), *S–C* foliations (Berthé et al., 1979; Lister and Snoke, 1984), rotated porphyroclasts (Passchier and Simpson, 1986) and asymmetric quartz *c*-axis fabrics (e.g. Bouchez et al., 1983). This paper is focused on the latter three of these main types of kinematic indicators.

Along with the documentation of each of these indicators has come an appreciation for the extreme heterogeneity of strain at the microstructural scale and a general agreement that a statistical approach is necessary for the accurate application of any indicator (Hanmer and Passchier, 1991). Aside from the documented cases of antithetical shearing (e.g. oblique microshears in domino-like microstructures or along *S* foliations; cf. Krohe, 1990), many individual domains appear to exhibit a shear sense which does not agree with the bulk shear derived from independent field relations and a more statistical analysis of the microstructure. At this point, it is essential to clarify the difference between antithetical shearing and what is referred to in this paper as *reverse shearing*. The former is consistent with the bulk shear direction as resolved across fractures or discrete oblique shear

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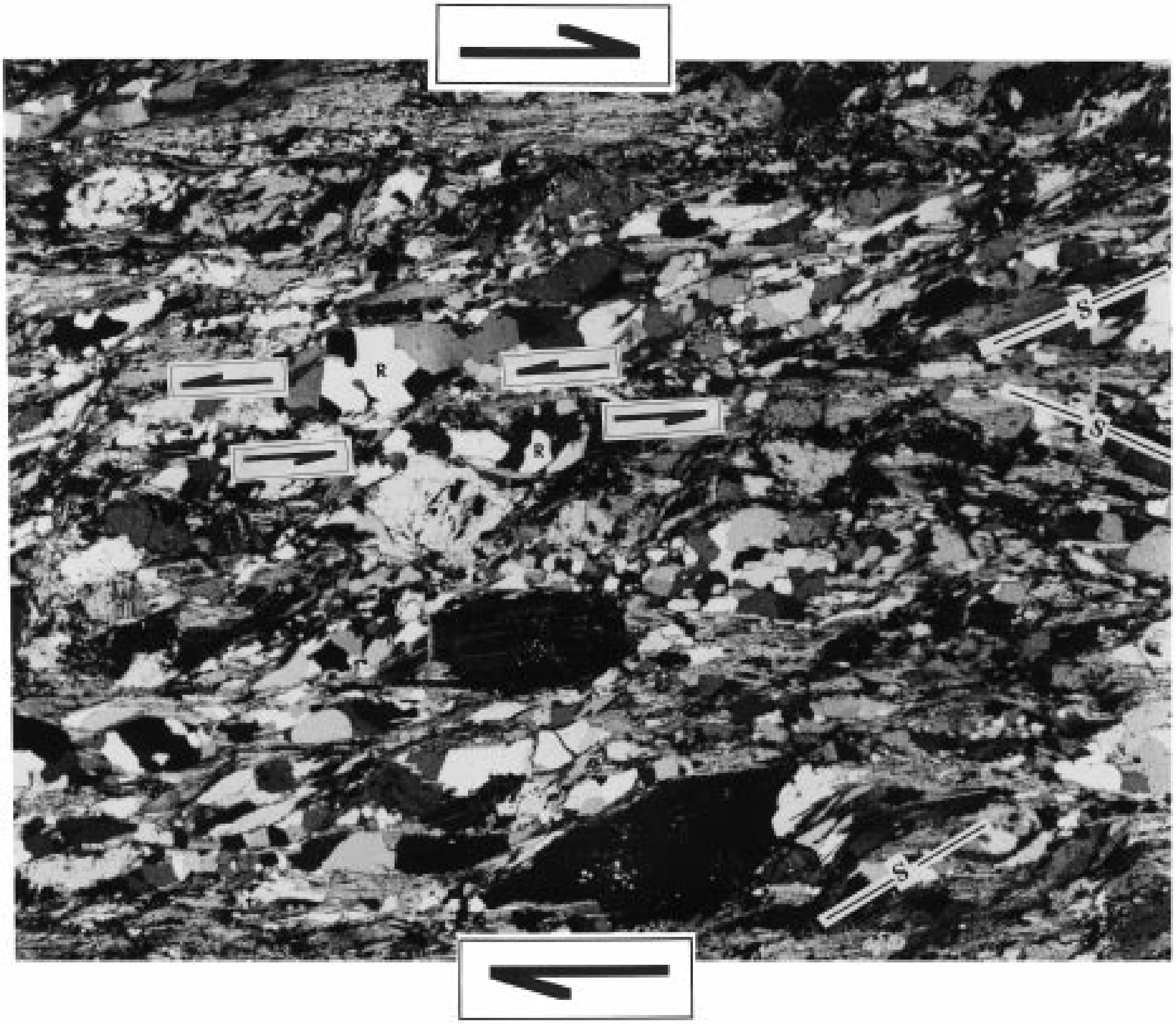

Fig. 1. *S*–*C* mylonite from the sheared margin of the granitic–gneissic Bação Complex, southeastern Brazil. The obliquity of the *S*-foliation defined by the polycrystalline quartz ribbons indicates a predominant dextral shear. However, a longitudinal domain showing the *S*-foliation in an opposing orientation is observable on the right margin of the photo. The domain boundary is marked by a shear fracture subparallel to the main shear plane which accommodates reverse shear, as indicated by displacement of the ribbon 'R'. Width of view is 4.5 mm.

bands which normally correspond to conjugate Riedel shears in brittle shear zones. Reverse shear, however, describes a movement on the main shear plane (usually defined by the shear zone boundaries) that is opposite to the overall shear sense displayed in most domains of the rock. The universality of these reverse shears has led to speculation about the geometry of ductile shear zones and has given rise to interpretations in terms of reactivation and overprinting of successive fabrics in mylonitic rocks (e.g. Bell and Johnson, 1992), and on the existence of non-simple shear zones (Passchier, 1996). Therefore, the correct interpretation of such kinematically opposing microstructures is essential to elucidate the deformation history of any shear zone.

The key to many of these opposing microstructures lies in the extreme strain partitioning at a microscale. Contrasts in both deformation mechanisms and strain rates at this scale can produce complex movements in the course of a single tectonic event. In this paper, an attempt is made to demonstrate how one progressive general shear deformational event can produce shear indicators which record opposite shear senses at the same geological instant and within the same thin-section. Examples for this paper are from natural mylonites formed in shear zones from southeastern and northeastern Brazil, most in the Quadrilátero Ferrífero granite–greenstone terrain of Minas Gerais. The scope of this paper considers the processes illustrated by these rocks rather than their implications for the regional geology. Readers interested in a comprehensive structural history of this region should consult the

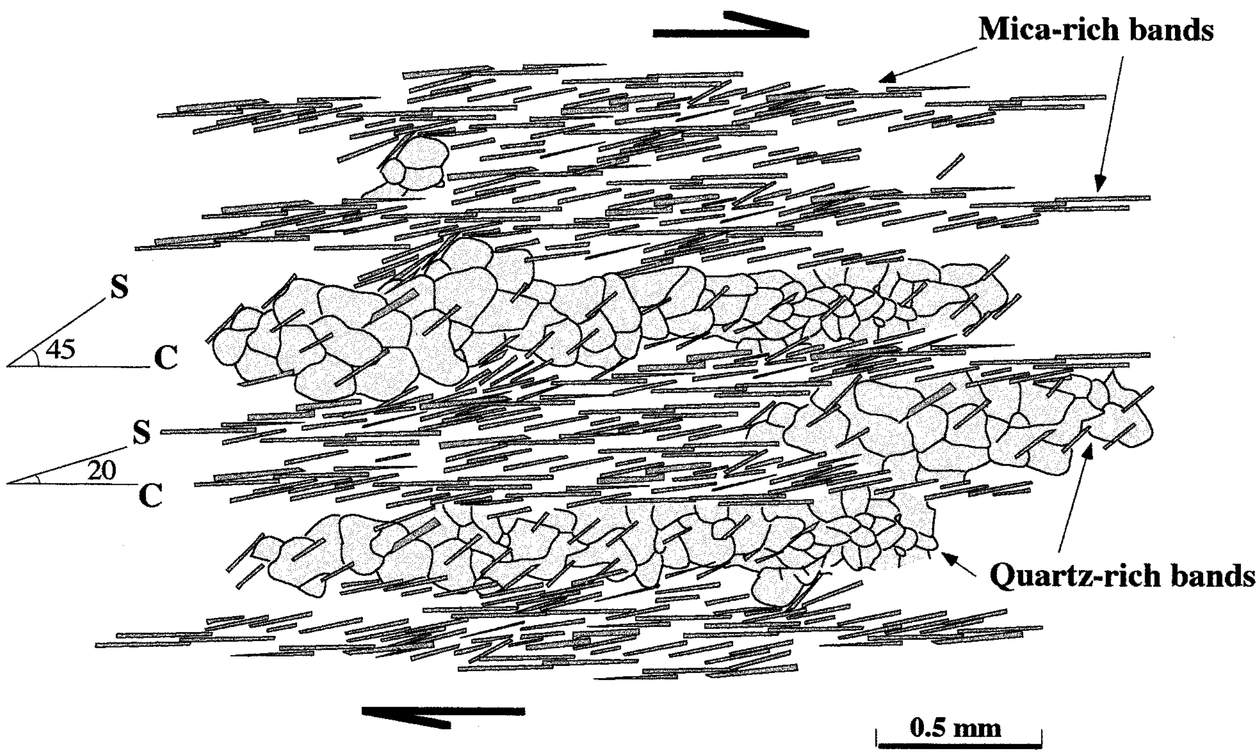

Fig. 2. Sketch illustrating a typical microstructure of micaceous quartzites and phyllonites sheared in the Moeda Bonfim shear zone. The two sets of *S*–*C* foliations are spatially related to the distribution of the mica-rich and quartz-rich bands. In the micaceous bands, the lower angle between *S* and *C* indicates a higher finite strain in these domains.

recent review by Chemale et al. (1994). Some corroborative examples were also taken from the literature. Most of the mylonite examples came from the low metamorphic grade Moeda Bonfim shear zone (Hippertt, 1994).

2. Observations

2.1. S–C fabrics

S–*C* fabrics (Berthé et al., 1979; Lister and Snoke, 1984) have been shown to develop in ductile and semiductile strain regimes with a strong non-coaxial strain component (e.g. Blenkinsop and Treloar, 1995). *S*–*C* fabrics form principally in rocks such as micaceous quartzites and granitoids, where there is a strong rheologic contrast between their rock-forming minerals. In these rocks, the two foliations are commonly defined by the alignment of platy minerals (or elongate quartz aggregates) around relatively rigid, lenticular domains (generally porphyroclasts of quartz or feldspar). There is a general agreement that *S* and *C* foliations are related to the same non-coaxial deformation event, as these fabrics commonly appear, for example, in shear zones cross-cutting undeformed, isotropic granitoids where no pre-existing fabric exists (e.g. Berthé et al., 1979). Also, the mineral characteristics of the *S* and *C* foliations are generally identical, reflecting their simultaneous formation under the same metamorphic conditions. The interpretation most commonly accepted is that the *C* foliations, which form parallel to the overall shear plane, accommodate most of shear strain in the rock (Berthé et al., 1979). The *S* planes, on the other hand, are considered to have formed perpendicular to the bulk shortening direction and accommodate only a small (if any), antithetical component of shear (e.g. Krohe, 1990). Thus, the obliquity of the *S* planes, when observed in an *XZ* section, directly indicates the orientation of the finite strain ellipse, and consequently the shear sense on the *C* planes. The oblique *S* planes form initially at 45° to the *C* planes. This angle decreases with progressive strain because the *S* foliations are rotated towards the shear plane which is the main fabric attractor during simple shear (Jiang and White, 1995; Passchier, 1997).

However, complex geometries are common and unambiguous cases of opposing *S* foliation orientation exist locally in most sheared rocks that we have observed, although one orientation is always largely predominant. Apart from representing a minor proportion of the deforming rock volume, the domains subjected to reverse shear generally show poorer developed fabrics than the domains depicting the overall

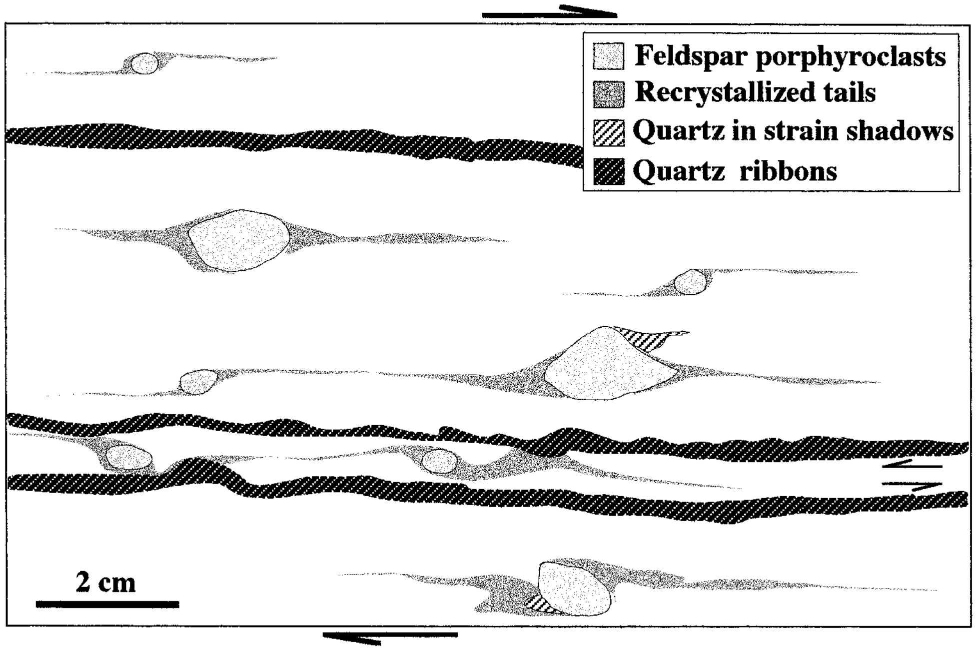

Fig. 3. Sketch showing a XZ view of a granitic mylonite from the high grade Além Paraiba shear zone, as observed in a road cut in Sapucaia town, Rio de Janeiro. Most porphyroclasts show a σ-type geometry indicative of predominant dextral shear. In the longitudinal domain which is limited by the two closely spaced quartz ribbons, however, the porphyroclasts appear to show an opposing asymmetry. Making an analogy with Fig. 1, these quartz ribbons might reflect shear fractures that were infilled by quartz.

shear sense. That is why reverse shear has commonly been overlooked in many shear zones. Fig. 1 shows an example taken from a granitic mylonite in the margin of the granitic–gneissic Bação Complex in southeastern Brazil. In this example, the sense of shear as inferred from the individual *S* foliations indicates locally opposite shear senses in adjacent microdomains. Some of these microstructures probably result from local flow perturbations due to the presence of porphyroclasts. However, we cannot invoke local perturbations in the example shown in Fig. 1 since the contiguous domains are laterally continuous and parallel to the main flow plane, at a scale much larger than that which can be affected by a single porphyroclast. In addition, many of these rocks showing opposing *S*–*C* fabrics are not porphyroclastic at all. Another example of antithetic asymmetries on a microscopic scale are the contiguous domains showing opposing grain shape fabrics (herringbone microstructure) which are common in some recrystallized quartz aggregates (e.g. fig. 3c of Pauli et al., 1996).

Aside from these opposing asymmetries, the angles between the *S* and *C* planes are also variable at a microscale in most sheared rocks that we have observed. These fluctuations directly follow the compositional differences between the different layers which reflect different viscosities/strain rates and different finite strains attained by each domain. These variations are tightly associated with production of reverse movements (discussed later). Fig. 2 illustrates an example of this phenomenon observed in micaceous quartzites and phyllonites of the Moeda Bonfim shear zone.

2.2. Mantled porphyroclasts

The geometry of rotated porphyroclasts with well-developed asymmetric tails of recrystallized grains or reaction-softening products is a common shear-sense indicator (e.g. Simpson and Schmid, 1983; Passchier and Simpson, 1986). Generally, the majority of porphyroclasts in different flow planes show a consistent asymmetry at both the outcrop and microscopic scale. The difficulty with this indicator is usually the interpretation of the tail geometry (σ or δ types) rather than with the presence of porphyroclasts with opposing asymmetries. Nevertheless, contiguous flow planes with porphyroclasts showing identical tails in an opposite orientation can be observed locally in some shear zones, principally at the outcrop scale, indicating that reverse shearing occurred in some domains. Fig. 3 illustrates such an example taken from the medium/high grade, strike-slip Além Paraiba shear zone from the Mantiqueira Province, southeastern Brazil.

The scarcity of antithetic porphyroclasts in porphyroclastic rocks does not necessarily indicate that local reverse shearing is not common. A possible explanation is that the harder, porphyroclast-rich domains accommodate only a minor component of the bulk strain, as in porphyroclastic rocks most of strain

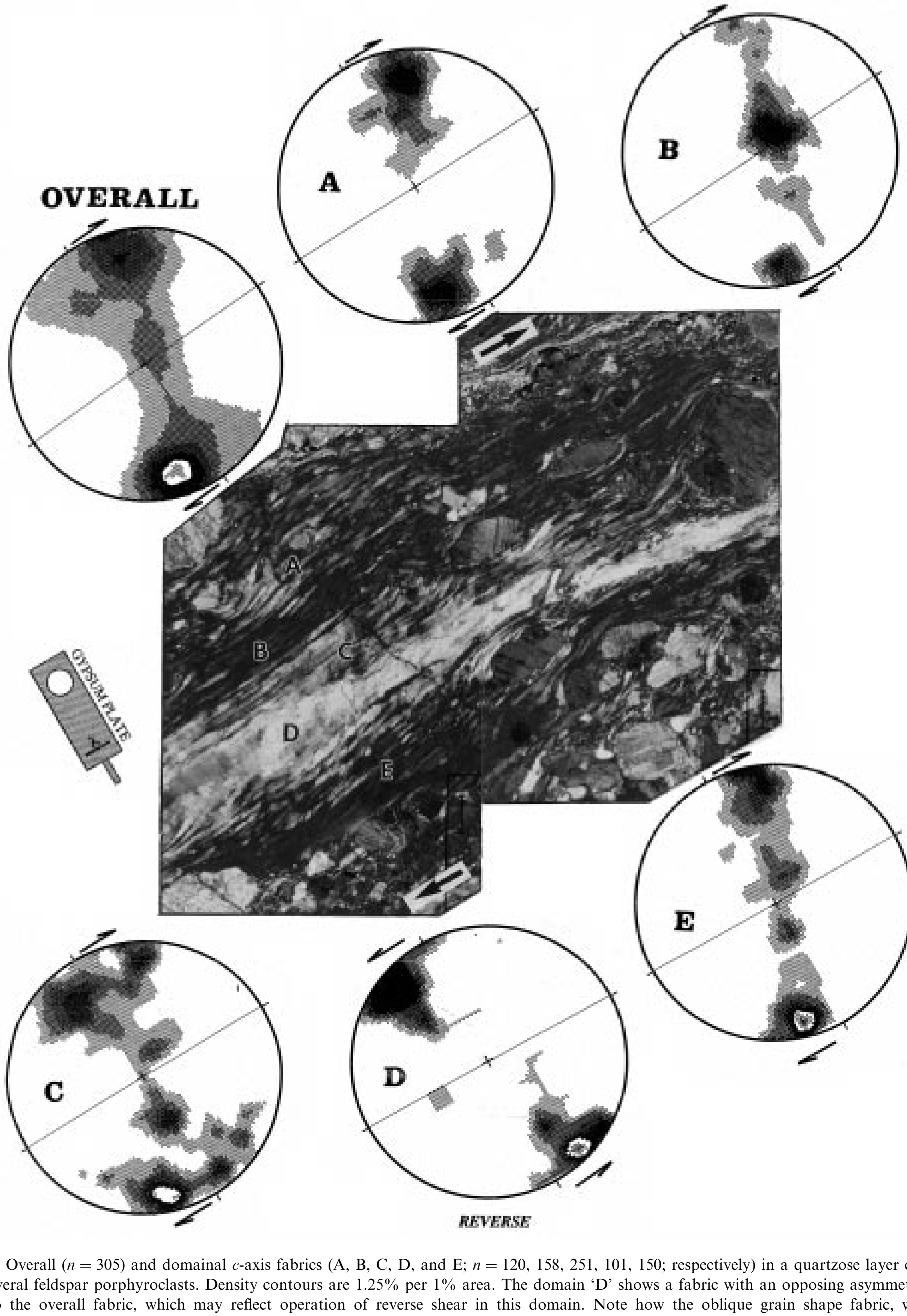

Fig. 4. Overall ($n = 305$) and domainal c-axis fabrics (A, B, C, D, and E; $n = 120$, 158, 251, 101, 150; respectively) in a quartzose layer containing several feldspar porphyroclasts. Density contours are 1.25% per 1% area. The domain 'D' shows a fabric with an opposing asymmetry relative to the overall fabric, which may reflect operation of reverse shear in this domain. Note how the oblique grain shape fabric, which is noticeable in the other domains, is not so clear in D. The orientation of the gypsum plate is indicated. Granitic mylonite from the low/medium metamorphic grade Senador Pompeu shear zone of northeastern Brazil. Width of view is 1.2 mm.

is accommodated in the softer, mica–quartz-rich domains between the porphyroclast-rich bands.

2.3. Quartz c-axis fabrics

Preferred orientation of *c*-axes develops as result of crystal-plastic deformation in response to the applied differential stress (Lister et al., 1978). *c*-Axis fabrics are characterized both by their skeletal outline and by the distribution of the *c*-axes over this skeleton. The fabric skeleton is believed to be principally related to the kinematics (Lister and Williams, 1979). The distribution of *c*-axis maxima, however, is less understood and appears to be controlled both by the operation of different slip systems (also dependent on pressure–temperature conditions) and by the pre-existing crystal orientations (e.g. Vissers, 1993).

The asymmetry of some *c*-axis fabric skeletons has been successfully used as a shear sense indicator in some shear zones (e.g. Bouchez et al., 1983; Law et al., 1986), although it is widely accepted that this indicator should only be used in conjunction with other independent criteria. That is because zones displaying opposing asymmetries can commonly coexist in the same thin-section, principally in rocks subjected to coaxial deformation (e.g. Fueten et al., 1991) where the operation of opposite shearings is expected. The situation is different in rocks deformed under a large component of non-coaxial strain. In this case, the fabric skeleton is expected to show a consistent asymmetry synthetic with the overall shearing. This asymmetry generally appears when composite fabrics including different microstructural domains of the same thin-section (or even from different thin-sections) are obtained. Our experience in obtaining these global fabrics in shear zones of different regions of Brazil indicates that asymmetric type-I cross-girdle fabrics (Lister et al., 1978) are the most common global fabrics. Fig. 4 presents an example of global and domainal fabrics in a granitic mylonite from the Proterozoic Senador Pompeu shear zone, northeastern Brazil. The overall *c*-axis fabric is an asymmetric type-I cross-girdle, contrasting with the single-girdle fabrics which are more common in the individual domains. Some individual domains, however, show a fabric with antithetic asymmetry relative to the bulk shearing. Similar situations have been recognized in other shear zones (e.g. Pauli et al., 1996). These opposing asymmetries, easily recognized by utilizing the gypsum plate (Fig. 4), indicate that opposing shears may have occurred in different parts of a shear zone or in different microdomains of a same thin-section. The weak branch of the global asymmetric cross-girdle fabric, therefore, may reflect local domains of reverse shear. Fueten et al. (1991) reached a similar conclusion when interpreting their domainal *c*-axis fabric as produced by coaxial strain.

3. Discussion

3.1. Dynamic deformation/strain partitioning and reverse shearing

Structural geologists currently recognize that the geometry of structures commonly records a local deformation that does not necessarily reflect the bulk deformation history viewed on a larger scale, and that the notions of coaxiality and non-coaxiality are basically scale dependent concepts (Hanmer and Passchier, 1991). The development of localized reverse shear zones can be explained according to two different models. In the general shear model, a component of pure shear is accommodated by the shear zone, leading to what Means (1989, 1990) has termed ‘stretching faults’. In this model, localized reversals in shear sense are caused by lateral gradients in the slip rate within the slip plane. The second explanation, which constitutes the bulk of this contribution, invokes viscosity contrasts between adjacent layers. Deformation of rheologically heterogeneous materials (*multilayers*) involves in most cases partitioning of both strain and vorticity between the material domains with different rheologic properties. Previous theoretical work in this field (Lister and Williams, 1983; Treagus, 1983) has already pointed out that, in a multilayered body, the more competent strong layers tend to deform coaxially while most of the vorticity is accommodated in the less competent, weak layers. More recently, numerical modelling of multilayer deformation by the finite element method (Ishii, 1992) has also demonstrated that adjacent, rheologically distinct layers may deform with a different sense of non-coaxiality in some situations. The ultimate cause for this is the larger rotation rate of the instantaneous stretching axis (ISA) relative to the rotation rate of the material lines.

Numerical modelling by Ishii (1992) has predicted the development of reverse shear in deforming rocks with alternating layers of contrasting viscosities [viscosity ratio $(\mu_a/\mu_b) > 2$] such as occur, for example, in a banded tectonite with alternating feldspar-rich, mica-rich and quartz-rich layers (e.g. Hippertt and Hongn, 1998). Previous microstructural studies (e.g. FitzGerald and Stünitz, 1993) have shown that in ductile/semi-ductile deformation regimes, a polymineralic rock tends to employ a number of deformation mechanisms to accommodate plastic strain. The operation of the different deformation mechanisms is controlled both by pressure–temperature conditions and material characteristics (modal composition, grain size, fluid content, etc.). The spatial partitioning of deformation mechanisms results from this tectonic banding normally developed in mylonitic rocks, principally at low to intermediate levels of strain (e.g. Hippertt, 1998). This banding is characterized by two or more alternat-

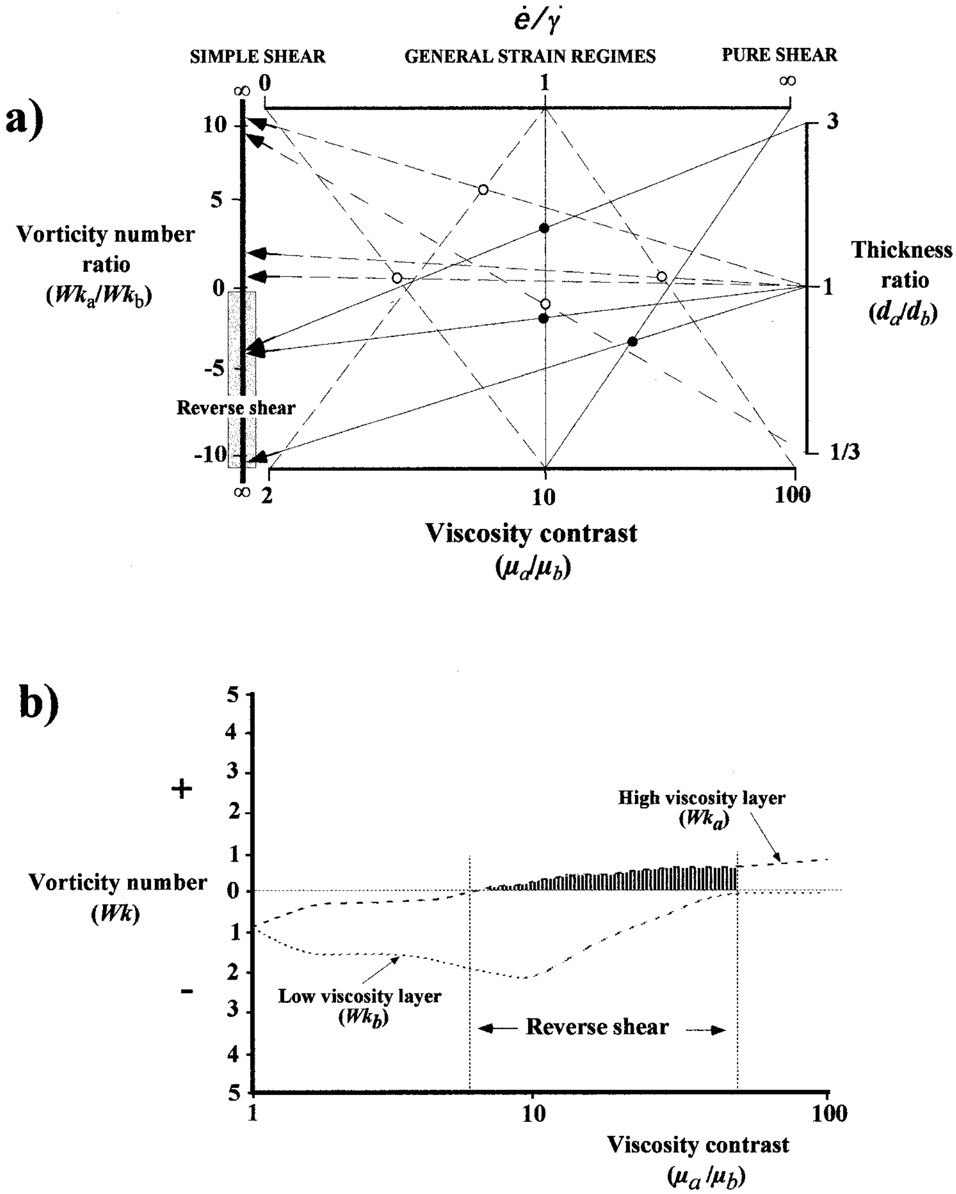

Fig. 5. (a) Four axes diagram summarizing the available numerical modelling results for multilayer deformation (layers 45° to the shear plane) at different proportions of pure shear and simple shear [indicated by the ratio extension rate ($\dot{e}$)/shear strain rate ($\dot{\gamma}$)], viscosity contrast (μ_a/μ_b) and thickness ratio (d_a/d_b), where 'a' and 'b' subscripts refer to layers with high and low viscosities, respectively. The black dots mark sets of conditions where reverse shear was produced [i.e. the resulting kinematic vorticity number ratio (Wk_a/Wk_b)<0]. The open circles mark sets of conditions that did not produce reverse shear (i.e. $Wk_a/Wk_b > 0$). Each dot or circle is in the crossing point between two lines which intercept the horizontal and vertical axes in the corresponding boundary conditions and resulting vorticity ratio (arrowed end). (b) Variation of vorticity number (Wk) in the high viscosity and low viscosity layers (Wk_a and Wk_b, respectively) with an increasing viscosity contrast (μ_a/μ_b). Layers 45° to shear plane, $\dot{e}/\dot{\gamma} = 1$, thickness ratio = 1. The two diagrams were constructed based on data produced by Ishii (1992).

ing domains with different composition and/or textural characteristics, hence with different rheologic properties. Each individual rheologic domain, though deforming simultaneously, permits strain through a different combination of deformation mechanisms and, as a consequence, at a different rate. However, an area dominated by a deformation mechanism significantly faster or slower than average may depart from this general shear sense with respect to its immediate vicinity by enhancing the viscosity contrast between adjacent layers, as predicted by Ishii's (1992) modelling. The selective competition between the deformation mechanisms and the consequent change in the material characteristics leads to a progressive homogenization of the microstructure with progressive strain, where only a few mechanisms should be dominant. Consequently, the deformation partitioning is also expected to decrease with progressive strain. The

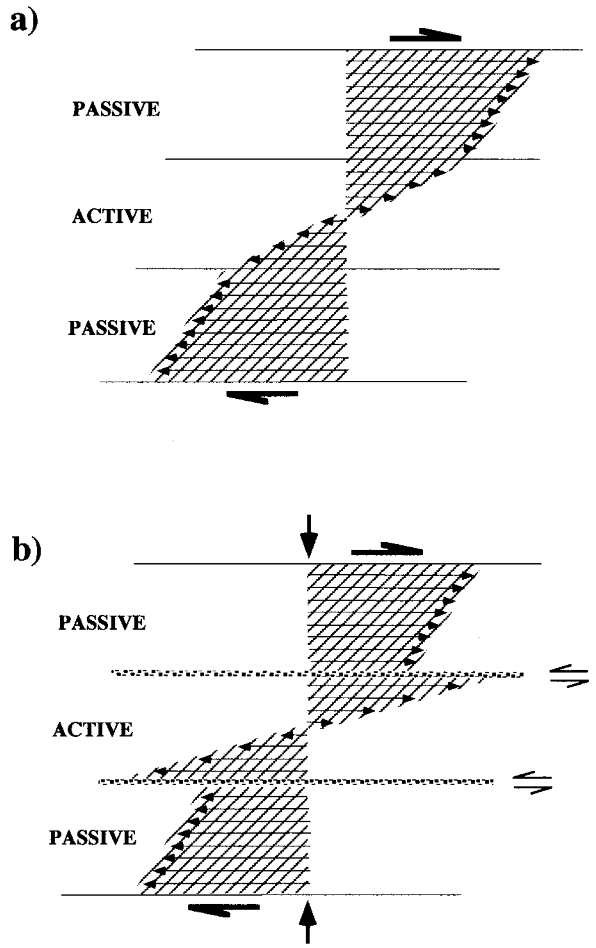

Fig. 6. (a) Sketch showing the displacement of particles in active and passive domains with a moderate strain rate contrast and continuity across the domain boundaries. (b) In general shear, a higher strain rate contrast between the active and passive domains may produce discontinuities and development of reverse shear on the domain boundaries (reverse shear fractures).

chance for development of local reverse shear, therefore, also decreases with progressive strain.

Apart from other controlling parameters, Ishii (1992) has shown that the occurrence of local reverse shear in the strong, competent layers is necessarily related to a component of coaxial deformation (Fig. 5a), a condition also inherent to the stretching fault model of Means (1989, 1990). An important question, however, regards the orientation of the rheologically contrasting layers relative to the overall shear plane. In Ishii (1992), reverse shear was shown to occur when the layers are oblique to the bulk shear plane, the obliquity angle varying with the strain regime, the thickness ratio and the viscosity contrast of the layers. This circumstance may constrain the direct application of his results to high strain tectonites where the banding is generally assumed to be subparallel to the bulk shear plane and, consequently, reverse shear would not be expected to occur. We infer, therefore, that features indicative of reverse shear should be more commonly observed in low strain domains of shear zones, where the tectonic banding is generally oblique to the overall shear plane defined by the shear zone boundaries (e.g. Berthé et al., 1979). With progressive strain, the rock fabrics tend to become parallel to the bulk shear plane (the main *fabric attractor*; Passchier, 1997), consequently impeding the occurrence of local reverse shear.

Local reverse shear can also be the result of switches in the strain rate within the individual domains due to abrupt changes in the operative deformation mechanisms, as a consequence of changes in the material characteristics with progressive strain (Figs. 6 and 7). For example, non-steady-state crystal plasticity and dynamic recrystallization may produce progressively finer grain aggregates up to a critical point where grain boundary sliding may be a more economic process, causing deformation to be accommodated at faster rates by granular flow (cf. Stünitz and FitzGerald, 1993). These switches generally correspond to strain-softening kinks in the strain–time path that enhance the viscosity contrast between the layers or even cause an inversion between the relative roles of passive (competent) and active (incompetent) domains in a banded mylonitic rock (e.g. Hippertt and Hongn, 1998; Fig. 8). This is consistent with Ishii's (1992) numerical modelling, where an increase in the viscosity contrast (μ_a/μ_b) of about one order of magnitude (from 2 to 10) has produced reverse shear in some particular layer orientations (Fig. 5b). This situation may also favour the formation of discontinuities (shear fractures) in the limits between domains deforming with contrasting strain rates, where reverse shear may occur (Figs. 3 and 6). Indeed, in banded tectonites, shear fractures commonly occur in the limits between bands with contrasting composition and rheologic properties. It should be noted, however, that in Ishii's (1992) numeric modelling, reverse shear has been produced *within* the stronger, competent layers, not in the layer boundaries as also commonly observed in naturally deformed rocks.

A natural example of these inversions in the rheologic behaviour of some rock layers is given by some granitic rocks deformed under low metamorphic grade conditions in the presence of an aqueous fluid (e.g. Handy, 1990; Hippertt, 1998). At low strain levels, the rock develops a deformational banding with alternating quartz-rich and feldspar-rich layers. Most of the strain is accommodated by plastic deformation accompanied by recrystallization in the less competent quartzose layers. The more competent feldspar-rich layers accommodate strain at slower rates via fracturing and rigid body rotation of the fragments. However, the presence of water may enable reaction softening with production of mica in parallel with progressive disappearance of feldspar (e.g. White and

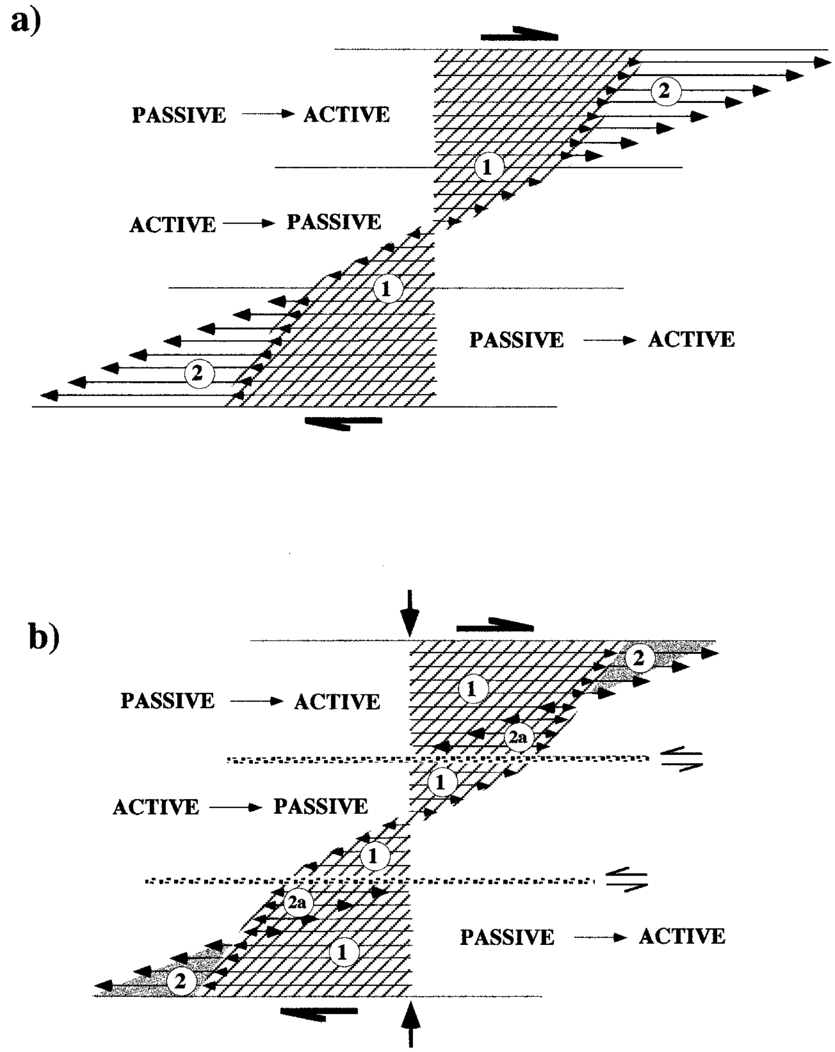

Fig. 7. (a) Sketch showing the displacement of particles during an inversion in the relative roles of active and passive domains, while maintaining the continuity across the domain boundaries. The displacement Field 1 correspond to the strain produced before the inversion (thin arrows). The Field 2 corresponds to the strain produced after the inversion (thick arrows). To simplify the diagram, zero strain is assumed in the passive domains after the inversion. No reverse shear is produced in the domain boundaries. (b) In a general shear regime, if discontinuities are formed after the inversion, the coaxial deformation component induces some particles to move backwards (Field 2a) causing reverse shear on the domain boundaries.

Knipe, 1978; Selverstone et al., 1991). Thus, the rock evolves towards a banded quartz–mica mylonite where, after a critical degree of reaction softening, the mica-rich layers (previously feldspar-rich layers) may become the main strain accommodator via slip on basal planes of mica, consequently preventing continued deformation in the quartzose bands, and eventually causing local reverse shear in the boundary domains. In the Moeda Bonfim shear zone, this inversion in the role of the quartzose bands is reflected in the weakening of the *c*-axis fabrics after a critical mica content of about 40% in the tectonites (Hippertt, 1994).

Based on the available data for material coefficients (A_0), activation energy (E) and exponent (n) for natural power-law rock flow (e.g. Shelton and Tullis, 1981; Kirby and Kronenberg, 1987; Koch et al., 1989), and assuming geologically reasonable differential stress values (in the range 10–100 MPa), we can roughly estimate the viscosity contrast between the different rheologic domains of a tectonite at a given temperature by using the equations (1) and (2), where μ is viscosity, σ_D is differential stress and $\dot{\varepsilon}$ is strain rate.

$$\mu = \sigma_D/\dot{\varepsilon} \quad (1)$$

$$\sigma_D = (\dot{\varepsilon}/A_0)^{1/n}\exp(E/nRT) \quad (2)$$

Our calculations show that this specific microstructural change (i.e. from a feldspar-rich to a phyllosilicate-rich multilayer; both containing quartz layers) will

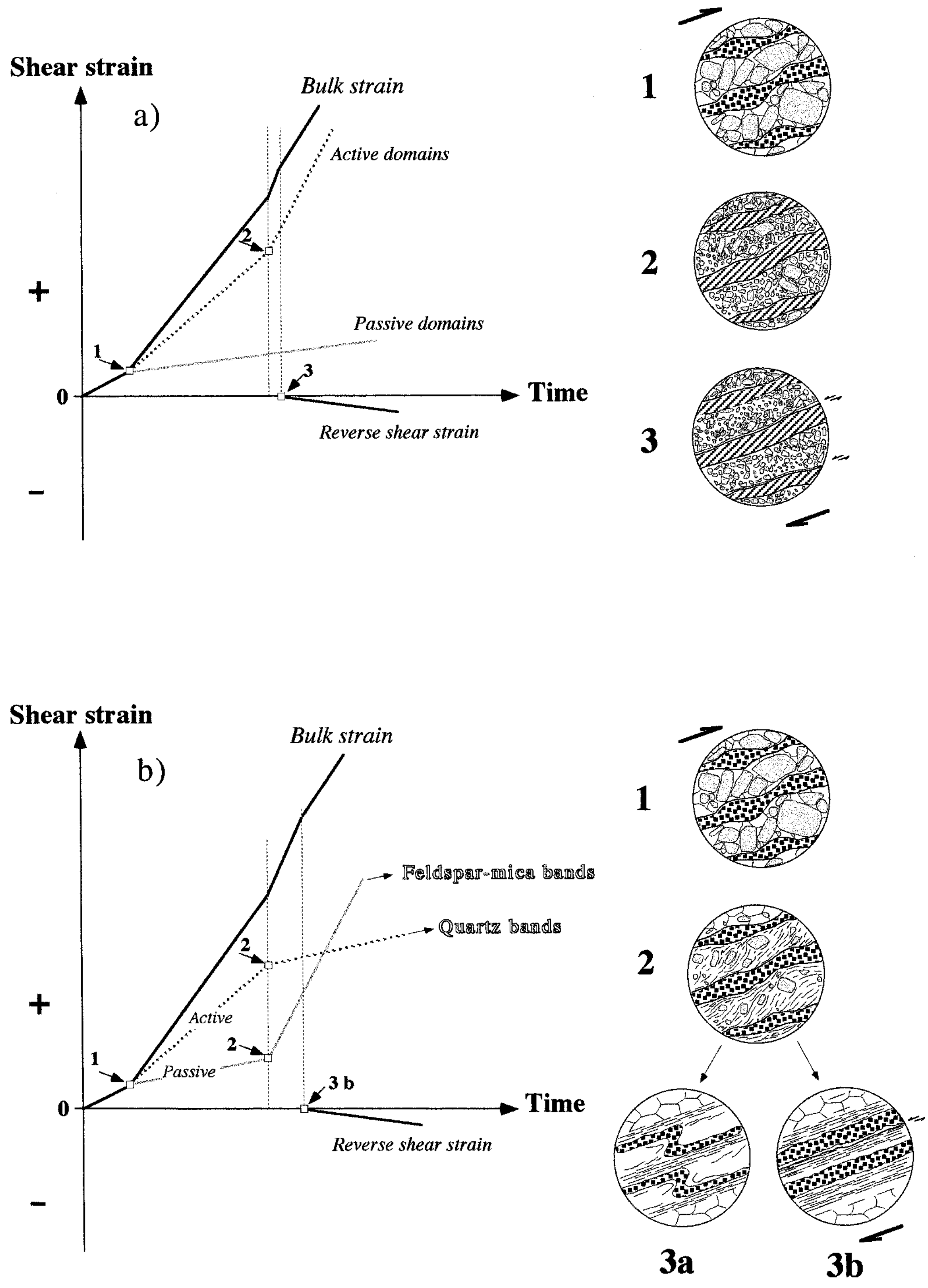

Fig. 8. (a) Strain–time path for progressive general shear of a granitic protolith. Kink 1 marks the onset of a tectonic layering with strain/deformation partitioning between active, quartz-rich layers and passive, feldspar-rich layers. Development of lattice preferred orientation induces softening and increases strain accommodation in the quartz layers (kink 2). If a critical strain rate contrast between the active and passive layers is attained, discontinuities and reverse shear may occur on the domain boundaries (kink 3). (b) Strain–time path for deformation of a granitic protolith in the presence of fluid. The increasing transformation of feldspar into mica leads to an inversion in the relative roles of the active and passive domains (kink 2). After inversion, if the continuity across the domain boundaries is maintained, folding of the quartz layers (microstructure 3a) may occur (Hippertt and Hongn, 1998). If discontinuities are formed, reverse shear may occur (kink 3b). The circles show the microstructures corresponding to the indicated kinks (see text).

increase the viscosity contrast between competent and incompetent layers (μ_a/μ_b) from about one order of magnitude (from approximately 10^2 to 10^3) which may induce local reverse shear, consistent with Ishii's (1992) numerical modelling results. Such changes are expected to be more common at low to medium metamorphic grades, where a larger number of deformation mechanisms generally coexist (FitzGerald and Stünitz, 1993), and fluid-assisted deformation processes (e.g. reaction-softening) can induce profound mineralogical changes. In contrast, a more homogeneous and continued deformation via pervasive crystal-plasticity is expected in high-grade shear zones. Hence, the presence of local domains with reverse shear should be more commonly found in low metamorphic grade tectonites. This is in fact true for the investigated shear zones of southeastern Brazil.

In summary, apart from other controlling factors such as a pure shear component, we suggest that the development of zones of local reverse shearing is a direct consequence of the evolving deformation/strain partitioning at a microscale. With increasing strain, the rheologic properties of the rock change, giving rise to the predominance of a new set of deformation mechanisms. Possible changes in metamorphic or fluid conditions are also a factor for potential mechanism switches. Whatever factors might influence switches in deformation mechanisms, the important conclusion is that these changes will not occur simultaneously nor uniformly throughout a deforming rock. Thus, we are left with a picture of a very dynamic deformation/strain partitioning at the microscale, with local differences in material characteristics, deformation mechanisms and associated strain rates evolving through time and ultimately causing localized reversals of shear sense.

4. Conclusion

This study has documented some common microstructures and textures indicative of operation of local reverse shear in mylonitic rocks. Because the presence of zones of reverse shear is a universal feature in most shear zones, this study confirms that a kinematic analysis based in any indicator must always be statistical, and that most shear zones depart from ideal simple shear strain paths. Based on our observations which are partially supported by some previous theoretical predictions, we conclude that reverse shear is a direct consequence of the evolving deformation/strain partitioning at a microscale, principally at low metamorphic grade conditions, where the rheologic properties of a deforming rock (and consequently their deformation mechanisms) are continuously evolving. The occurrence of reverse shear is more common at low to moderate levels of finite strain, where a compositional tectonic banding is generally present. In a given instant of the deformation path, an individual domain which is deforming with a faster or slower strain rate than its vicinity may induce local reverse shear in the domain boundaries. Detailed microstructural analysis can elucidate the heterogeneity of strain at a microscale, and identify the ultimate causes of reverse shear in each shear zone.

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