

## Elementary geometry of deformation processes

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**Abstract**—All rock transformations can be seen as essentially structural transformations. Particles of some size move, and a change in the arrangement or intensity of structural elements occurs. The structural elements themselves are mostly linear or planar regions, of various absolute sizes, at or around which there is some interruption in geometrical or compositional regularity. A fundamental problem, in structural geology and petrology alike, is to understand the geometry of the processes of structural change. A geometrical classification of deformation processes is presented, based on whether there is a velocity difference across a typical structural element, and on whether the local rates of strain and of structural and material migration are zero. In this classification, deformation mechanisms are a subclass of deformation processes. The classification may promote thinking about several general questions: What kinds of deformation processes lead to features with memory of movement history? Do distinct deformation processes necessarily have distinct structural signatures? Can processes with different strain rates be simultaneously active at a given point in a volume of deforming material?

### INTRODUCTION

THE focus of this Special Issue on the *geometry* of rock deformation celebrates John Ramsay's enormous contribution in this field. It also reminds us of the primacy of geometrical data and geometrical interpretation in our subject. We observe structural geometry and seek to reconstruct from this at least a piece-wise history of geometrical change, using various methods of kinematic or strain analysis, or section balancing (Ramsay 1967, Dahlstrom 1969, Ramsay & Huber 1983, Woodward *et al.* 1985).

The structural geometry of rocks can also be seen in a larger way. All rock transformations, structural or metamorphic, diagenetic or igneous, are rearrangements of particles, on one scale or another. Even 'chemical processes' can be seen as essentially geometrical changes. Like familiar mechanical processes, they rearrange particles, and they rearrange, introduce, or delete structural features. The purpose of this paper is to look at the geometry of deformation processes in a way that is elementary, yet scale-independent. The restriction to 'deformation' processes is not as great as it sounds. By deformation processes, I simply mean processes active in deforming rock bodies, including processes where the local strain rate is zero. A subclass of deformation processes are deformation mechanisms.

### STRUCTURAL ELEMENTS, PARTICLES AND PROCESSES

We start by making a clear distinction between structural elements and particles. *Structural elements* are interruptions or 'defects' in the geometrical or compositional regularity of a region. Examples are vacancies and dislocations in crystals, bedding planes and fold axial planes in fold trains, and porphyroblasts and plu-

tons in schist bodies. As these examples indicate, structural elements can be of any absolute size. When looked at closely enough, all structural elements extend over *volumes* of material. The volumes can be elongate in one, two or three dimensions, giving elements sometimes referred to as linear, planar and point defects, respectively. When structural elements are repeated, more or less rhythmically in space, they define morphological *fabrics* in the sense of Sander (1970, pp. 3–5). A complete description of the structure of a rock body, over a given range of scales, entails description of all the structural elements present within that range of scales.

*Particles* here are patches of material. They can be of any absolute size—atoms, grains or blocks. Particles fill space completely, for our purposes. At every point (in space), there is a particle (of material). Particles may be solid or fluid, or somewhere between these conditions. Particles can have boundaries that are defined by structural elements, but this is not necessary. Hence the word 'patch' above rather than 'piece'. Particles define a continuous material field, in contrast to structural elements, which are arrayed discontinuously through the material field.

A *process* is said to be acting in a subregion of a deforming material if particles are being rearranged there. This can take place with or without straining of the subregion and with or without structural change. An example of a process without straining is the shuffling and diffusion of atoms that takes place at a migrating grain or phase boundary (i.e. at a quartz–quartz grain boundary during static recrystallization, or at a quartz–garnet phase boundary during static garnet growth). A process without optically visible structural change is the sliding of grains past one another along a segment of their boundary that is aligned in the sliding direction. A full description of particle rearrangement provides a full description of any process. (You tell me how every atom has moved from one moment to the next, and I can tell

you exactly how each structural element has behaved.) The alternative description of a deformation process, a description of the structural change, is always incomplete and ambiguous. A complete description of structural change between two moments is *not* sufficient to establish a complete description of particle motion. So different deformation processes are most essentially *different patterns of particle motion*. The superiority of the material description of processes arises because structures are, by definition, discontinuous and non-material.

### Particle motion

Particles can be rearranged in two end-member ways (Fig. 1). There is ideal *co-ordinated motion*, in which no particle changes any of its nearest neighbors. Only particle-particle distances change. And there is full *mixing motion*, in which every particle changes all of its neighbors. Real changes in particle arrangement during flow fall between these two extremes, especially when small increments of rearrangement are considered. Examples of particle rearrangements with a strong mixing component occur during the formation of tectonic melanges and during atomic diffusion processes.

When the co-ordinated component of particle rearrangement is large enough, the geometry of rearrangement can be described by picking three representative particles which define a triangle enclosing a structural element of interest. The particles need to be more widely spaced, relative to their size, the larger the mixing component of motion and the larger the 'diffusion' distance. The change of configuration of this triangle from one moment to the next defines a *local strain rate* across the structural element (Fig. 2). Something similar can be said for three dimensions, but we limit discussion here to two dimensions. A strain rate can be defined for any triangular region, whether or not

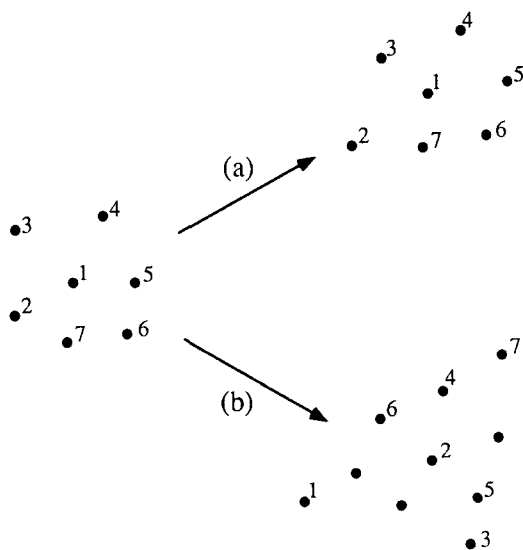


Fig. 1. (a) Co-ordinated motion of an array of particles. (b) Mixing motion of the same array. The unnumbered particles seen in the lower right are particles that were not nearest neighbors of particle 1 in the original state.

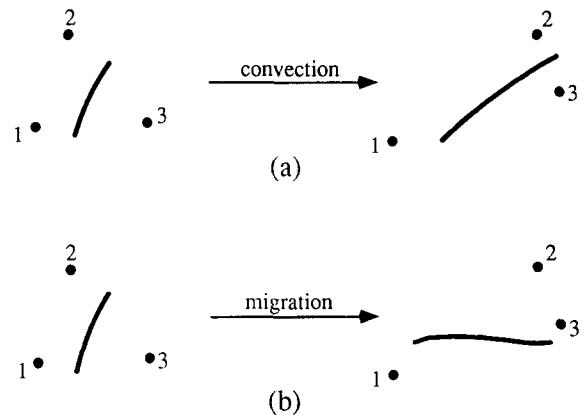


Fig. 2. Triangles defined by three particles enclosing a portion of a structural element of interest (line). The rate-of-change of configuration of the triangle defines a local strain rate. The structural element can convect with the material (a) or migrate through it (b).

straining is uniform within it. In fact straining can never be perfectly uniform over any real material region on every scale.

When the particles under consideration are individual atoms, it is necessary for our purposes to define particle velocities as the rate-of-change of the *mean positions* of the atoms. The strain rate of a material triangle is then defined by the differences between these averaged velocities of the atom, at its corners, and is independent of the thermal oscillations of the atoms about their mean positions.

### Structural changes

Two kinds of adjustments are possible for any structural element. It can move and/or it can change intensity. The motion can be *with* the surrounding particles, here called *convection* (Fig. 2a), or *through* the surrounding particles, here called *migration* (Fig. 2b). Intermediate behaviors between convection and migration are also possible and important in rocks. We return to this later.

Independently of whether it is moving, a structural element can change intensity. That is to say the compositional and/or geometrical contrast across it can strengthen or weaken. Folding and unfolding across an axial plane are examples of geometrical intensity changes. Strengthening or weakening of the compositional contrast across an exsolution lamella boundary are examples of chemical intensity changes across a structural element. A special case of strengthening is *nucleation*, where the earlier intensity is zero. A special case of weakening is *annihilation*, where the later intensity is zero.

The classification of processes to follow uses the above ideas of local strain rate and migration, and two other ideas. A process going on within a triangle is said to be *conservative* if there is no net gain or loss of material across the triangle's boundaries. This definition is not entirely satisfactory because it is not clear how the boundaries of the triangle are defined, but perhaps it will do for now. A structural element is said to be *active* if

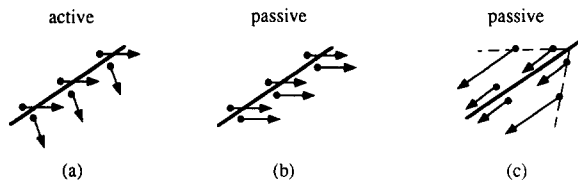


Fig. 3. Structural elements (heavy lines) and velocity distributions across them. Active element (a) is site of a velocity discontinuity. Passive elements (b) and (c) are not sites of velocity discontinuities, although the velocity gradient may be discontinuous (c). Dashed lines in (c) represent a fold profile.

there is a discontinuity in particle velocity across it (Fig. 3). Otherwise it is *passive*. Dilating cracks, migrating twin boundaries and faults on any scale are examples of active structural elements. The terms ‘active’ and ‘passive’ have been borrowed from Paterson & Weiss (1961) but their meaning has been changed somewhat. According to Paterson & Weiss (1961) and Turner & Weiss (1963, p. 384), any *imposed* structural element is active—that is to say any element introduced by deformation, like a fold axial plane. In my usage, a (non-migrating) fold axial plane is not active but passive. It is the site of a discontinuity in the particle *velocity gradient* (Fig. 3c), but it is not the site of a discontinuity in the particle velocity itself (see also Cobbold *et al.* 1984, Means & Jessell 1986).

**A CLASSIFICATION OF DEFORMATION PROCESSES**

The classification (Fig. 4) is meant to apply *locally*—that is to a triangular region small enough to straddle a

single structural element of some kind. On the other hand, no absolute size or size-range is implied. The triangular region could be tens of kilometers across, and the structural element could be a segment of a major fault zone. Or the triangular region could be only several unit cells across and the structural element could be a vacant lattice site.

The classes are distinguished from one another by applying the criteria of local strain rate, conservativeness, structural element activity, and structural element migration rate, in that order. Division by strain rate yields two classes. Subsequent division by the independent criterion of conservativeness creates a total of four classes (2<sup>2</sup>). One might expect that use of two further criteria, activity and migration rate, would result in a total of 16 classes (2<sup>4</sup>). Instead, there are only 12 classes. This occurs because in regions of zero strain rate, the structural element is necessarily passive. The activity criterion, in other words, creates only two new classes, for a cumulative total of six. Use of the fully independent migration-rate criterion then doubles this number, for the total of 12 classes shown in Fig. 4.

What distinguishes processes responsible for the local accumulation of strain (‘deformation mechanisms’) from other processes that can go on in deforming materials but which do not directly contribute to the accumulation of strain? Figure 4 represents my suggestion that the necessary and sufficient condition for a deformation process to be a deformation mechanism is *activity* of the structural element. The idea is simply that for strain to occur across a structural element, there must be shearing parallel to the element, dilation or closing across it, or some combination of these kinds of discontinuity in the velocity field.

		Deformation Processes					
		$\dot{\epsilon} \neq 0$			$\dot{\epsilon} = 0$		
		Active		Passive			
		$\dot{M} \neq 0$	$\dot{M} = 0$	$\dot{M} \neq 0$	$\dot{M} = 0$	$\dot{M} \neq 0$	$\dot{M} = 0$
Conservative	1	dislocat'n glide twin bound. migration	3	dy. grain b'dy. migration segregat'n veining	7	9	11
				5	passive g.b. deformation folding with A.P. convect'n	st. grain b'dy migration exsolution with lam. gr'th	st. polyg'n ch. exch. at fixed b'dy
Non-conservative	2	dislocat'n climb dis'n at styl. margin	4	6	8	10	12
				dy. ph. b'dy migration	dy. ch. exch. w. g.b. fluid	st. repl. veining st. porph. growth	ch. exch w. g.b. fluid
		Deformation Mechanisms					

Fig. 4. A classification of deformation processes.  $\dot{\epsilon}$  and  $\dot{M}$  represent local strain rate and migration rate, respectively. The structural element is active in classes 1–4. In all other classes there is no velocity discontinuity across the element, and it is accordingly passive. Cited examples of each class are as follows. (1) Dislocation glide, twin boundary migration. (2) Dislocation climb, dissolution at stylolite margin. (3) Grain-boundary sliding, fault slip. (4) Cracking, dike injection. (5) Dynamic grain-boundary migration, segregation veining. (6) Dynamic phase-boundary migration. (7) Passive grain boundary deformation, folding with axial plane convection. (8) Dynamic chemical exchange with grain-boundary fluid. (9) Static grain-boundary migration, exsolution with lamella growth. (10) Static replacement veining, static porphyroblast growth. (11) Static polygonization, chemical exchange at fixed boundary. (12) Non-conservative chemical exchange with grain-boundary fluid.

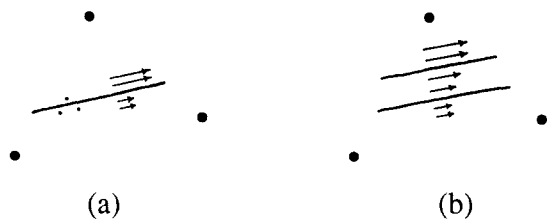


Fig. 5. (a) A velocity discontinuity at a sliding grain boundary. Arrows are velocity vectors. The small triangle is enlarged in (b), which shows no velocity discontinuity on the smaller scale.

This way of defining deformation mechanisms brings with it a potentially troublesome scale-dependence. Suppose, for example, that one is dealing with a grain boundary across which there is a shearing type of discontinuity in the velocity field (Fig. 5a). On this scale, the shearing phenomenon clearly involves an active structural element and the process going on there is a deformation mechanism producing distortion of the particle triangle. But now, reduce the size of the triangle, so that the shearing grain boundary is seen enlarged (Fig. 5b). Suppose that at this scale there is no fault-like feature, only a grain-boundary shear zone with no velocity discontinuity inside it or at its margins. By our previous definition, the grain boundary is no longer an active structural element, and the concentrated deformation rate there is no longer a deformation mechanism. Perhaps the best way out of this trouble is to relax the definition of activity somewhat, to say that an active element is an element corresponding to a velocity discontinuity *or* to a relatively steep, bounded gradient in the velocity field. This relaxed definition of activity is used in what follows.

An interesting feature of the class 1 deformation mechanisms is that there is a geometrically necessary proportionality between the rate of strain or displacement and the rate of migration of a structural element. An example is shown in Fig. 6, where the structural element is the vertical dashed line that migrates to the solid position. This line can be thought of as the trace of a migrating kink axial plane, in a situation where the fold interlimb angle is fixed, as in Paterson & Weiss (1966) kinking. Or the migrating line can be thought of as the trace of a mechanical twin boundary. A similar diagram could be drawn for any martensitic phase transformation. What all these processes have in common is that the difference in particle velocity  $t$  across the boundary is linearly related to the velocity  $m$  at which the boundary migrates laterally. Furthermore, all such class 1 processes produce kinematically reversible structural changes. If the particle displacements are reversed, the structural change is reversed. In Fig. 6 for example, if  $t$  is reversed,  $m$  is reversed, assuming coherence is maintained at the boundary. We see an application of the idea of kinematic reversibility later.

Class 2 processes like dislocation climb and partial dissolution at stylolite margins (Fig. 7) again involve a linear relation between migration velocities and strain or displacement rates, but these non-conservative pro-

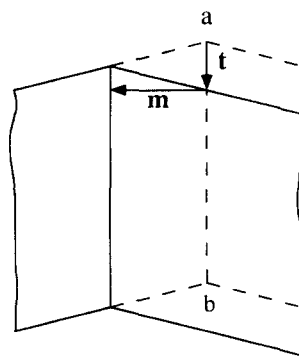


Fig. 6. Example of a class 1 process, in which there is a geometrically necessary, linear relationship between the difference in particle velocity  $t$  across a structural element (a-b) and the rate  $m$  at which the element migrates. This example could represent lateral migration of a (Paterson & Weiss 1966) kink boundary or a twin boundary.

cesses are not expected to be kinematically reversible in practice, although they are in principle. The reason they are not effectively reversible is that diffusive gain or loss of material is involved, and it would be highly unlikely for any diffusion process to reverse itself exactly. For example if a stylolite is forming by accumulation of insoluble clays and diffusive removal of calcite grains, it would be unlikely that, upon reversal of the straining, each of the dissolved calcite grains would begin to grow again. A more likely development would be formation of a calcite vein along the site of the former stylolite.

Processes of classes 9–12 are processes taking place in subregions where the current strain rate is zero. Examples are exsolution with migration of an exsolution lamella boundary, classical polygonization and chemical exchange between a grain and adjacent grain boundary fluid. A reasonable question to ask here is whether there really can be any regions of 'zero' strain rate in stressed materials? Shouldn't there be *some* strain rate around the site of any deformation process? Does this potentially make *any* deformation process a deformation mechanism? My suggestion here is that the answer to the last two questions is probably yes, but that we can still usefully think of classes with 'zero' strain rate, if we read this to mean negligible strain rate.

## COMPOSITE PROCESSES

Another difficulty with a classification like the one in Fig. 4, is that it is based on four criteria that sound simple enough at first, but which turn out to be difficult to apply to many recognized processes. Consider the example of stylolite formation. When seen at low magnification, so that both sides of a stylolitic seam are included, the structural element is the stylolitic seam as a whole, and stylolite formation is clearly a class 4 process, the inverse of cracking (Fletcher & Pollard 1981). but now change the magnification so that the structural element is just one of the boundaries between the clay seam and the host rock (Fig. 7). Is this structural element migrating or not? The structural element *is* migrating with respect to the clay mineral grains, in the sense that these grains

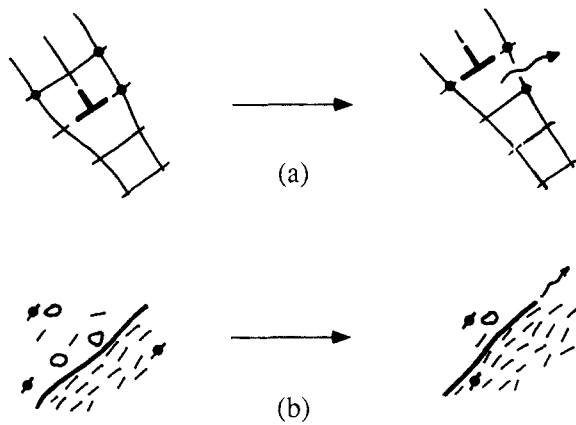


Fig. 7. Examples of class 2 processes. (a) Dislocation climb. (b) Dissolution at stylolite margin. Structural elements are dislocation core and edge of stylolitic seam, respectively.

cross the migrating interface. But the structural element is *not* migrating with respect to typical calcite grains. These grains move toward the clay seam boundary but vanish just as they get there. So the clay seam boundary is migrating through one set of grains but not migrating with respect to another set. It is not clear then whether stylolite formation should be placed in class 2 or in class 4.

I suggest that a way out of this dilemma (that may have some use besides salvaging the classification) is to say that dissolution at a clay seam boundary is a combination of two, *interpenetrating* processes. There is a type 2 process and a type 4 process acting in the same sub-region, at the same time. Seen this way, the process at a stylolite margin is what might be called a *composite* process, with components, on a given scale, in more than one compartment of Fig. 4. Another example of a composite process is general dislocation motion, with a conservative component (dislocation glide, class 1) and a non-conservative component (dislocation climb, class 2).

We can extend this line of thought to strain rate. Figure 4 is based partly on whether the local strain rate is zero or non-zero. But there may be situations where there are two, interpenetrating local strain rates, because the overall process is really a composite of two processes, each with its own strain rate. Consider a case of volume-diffusion flow in a grain of some silicate mineral where atoms of two kinds, say A and B, diffuse at different rates because they are held more or less tightly in their normal structural sites. The strain rate defined by three atoms of type A will tend to be different from the strain rate defined by the three atoms of type B. A predicted geometrical consequence of this kind of *composite diffusion process* is that non-stoichiometry must develop at grain margins, promoting dissolution or reaction processes there (Fig. 8). The theory of such stress-induced 'kinetic decomposition' has already been developed, by Dimos *et al.* (1988).

Can we extend the same kind of speculation to structural element *activity*? Can a structural element be

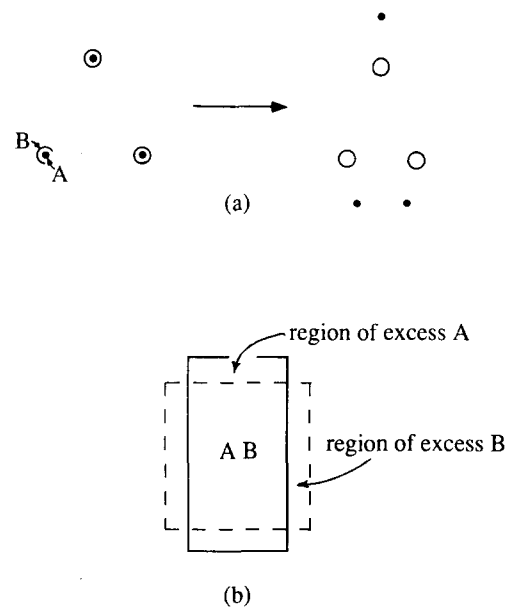


Fig. 8. (a) Composite volume-diffusion flow process, with a higher strain rate for atoms of species A than for atoms of species B, resulting in non-stoichiometry at grain margins (b).

active with respect to some kinds of particles and passive with respect to others? Perhaps there are examples among phase-transformations, where atoms A shear martensitically into their new structure but atoms of another species B diffuse in unco-ordinated fashion across the phase boundary and find correct sites in the new structure but not necessarily in their original unit cells. This is the interpretation of Furnish & Bassett (1983) of their observation of changing X-ray diffraction spectra obtained during the olivine-to-spinel transition in  $\text{Fe}_2\text{SiO}_4$ . The migrating phase boundary in a case like this is active with respect to atoms A but passive with respect to atoms B. The phase-change is a deformation mechanism with respect to atoms A but not a deformation mechanism with respect to atoms B. Species-specific strain rates and straining mechanisms are unfamiliar in the geological literature but are perhaps important in environments where diffusion processes dominate.

## REVERSIBILITY AND MEMORY STRUCTURES

It was pointed out earlier that class 1 processes give rise to structural changes that are in principle reversible if the particle displacements are reversed. This behavior follows from the strict geometrical relationship between particle velocities and migration velocities in all class 1 processes. An even simpler kind of reversibility is exhibited by processes of classes 3 and 7, where there is no migration of the structural element, so the element necessarily moves *with* the particles and reverses this motion if the particle motion reverses. Three examples are distortion of passive grain boundaries, folding with non-migrating axial planes, and fault slip. What all three reversible processes have in common is inability to record details of deformation history. For example in

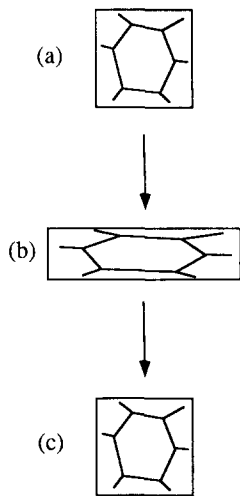


Fig. 9. A reversible structural change (passive grain-boundary deformation, class 7). The structure in (c) has no memory of state (b).

Fig. 9, the grain shape in (c) is no different from the grain shape in (a), so in a sense the structure in (c) has forgotten the history between (a) and (c). Of course no real structural change is ever perfectly reversible, because no real deformation of particles can ever be precisely reversed, but there may be sufficiently close approximations to structural reversibility to make it a practical concept.

Conservative processes of class 5 and all the non-conservative processes are likely to have at least some memory of deformation history. These are processes to which one can look for structures that remember whether deformation accumulation was, for example, close to coaxial or far from coaxial. This is not to say that non-coaxial indicators cannot arise where processes of classes 1, 3 and 7 are involved. In fact some familiar asymmetric structures, like asymmetric tails on porphyroblasts (Simpson & Schmid 1983) can arise under the influence of these processes too. For example, asymmetric tails may arise by passive distortion of a porphyroblast boundary (class 7) where it is dragged into a porphyroblast-bounding micro-shear zone. But here the asymmetry arises not because of an inherent property of the class 7 process, but because the porphyroblast spans an asymmetric pattern of deformation-rate distribution. The theory of memory structures needs to be better understood, because, for one thing, it is fundamental to improving and extending our use of sense-of-shear and sense-of-slip structures.

### STRUCTURAL SIGNATURES

A question of practical importance is whether different deformation processes necessarily leave distinct structural imprints on rocks, as individuals have distinct personal signatures. Structural 'signatures', if they exist, would be valuable indicators of previously active processes. Of course to some degree we *know* structural signatures exist. Broken-looking grain fragments with matching boundaries point to a fracture process; intra-

crystalline kink bands point to crystal plasticity. But can we assume that each distinct deformation process (characterized most fundamentally by a distinct pattern of particle motions) will produce a distinct structural record? I think the answer in principle must be yes, but that for practical purposes the answer will often be no, for two reasons. Later or concurrent interpenetrating processes can overprint and obscure the structural signature of a previously dominant process. This is a serious problem, particularly for identifying deformation mechanisms in formerly deep rocks, like high-grade metamorphic rocks. In such rocks, the structural signatures of deformation mechanisms may commonly have been written in 'fading ink'. The most familiar example of an overprinting process is probably grain-boundary migration, proceeding either during deformation or afterwards, that changes grain shapes back toward equiaxed shapes and eradicates the substructures which were the structural signature of the deformation mechanism. The other practical reason why different deformation mechanisms may not leave different signatures arises from the fact that we often look at deformed rocks just on one scale—especially the thin section scale. As Tullis & Yund (1987) have pointed out for feldspar, the optical microstructure produced by different processes (in their case suboptical distributed fracturing and dislocation motion) can be identical. Means & Ree (1988) have shown that subgrain boundaries produced by processes as different as polygonization and grain-boundary migration can be optically indistinguishable.

The above notwithstanding as a matter of principle, each distinct microscopic process should be accompanied, at least on the atomic scale, by a distinct structural evolution and, at any given moment, by a distinctive array of structural elements. Similarly on the regional scale. Each distinct process should generate a distinctive set of structural elements, which can in principle be told from similar but different sets by looking at critical indicators a size or two smaller than the main structures in question.

### CONCLUDING NOTE

The above ideas are intended to stimulate thinking about elementary aspects of rock transformations. They are not intended to introduce terminology or concepts of lasting value. In particular, the classification of Fig. 4 probably has no lasting value.

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### REFERENCES

- Cobbold, P. R., Means, W. D. & Bayly, M. B. 1984. Jumps in deformation gradients and particle velocities across propagating coherent boundaries. *Tectonophysics* **108**, 283–298.

- Dahlstrom, C. D. A. 1969. Balanced cross sections. *Can. J. Earth Sci.* **6**, 743–757.
- Dimos, D., Wolfenstine, J. & Kohlstedt, D. L. 1988. Kinetic demixing and decomposition of multicomponent oxides due to nonhydrostatic stress. *Acta metall.* 1543–1552.
- Fletcher, R. C. & Pollard, D. D. 1981. Anticrack model for solution surfaces in pressure solution. *Geology* **9**, 419–424.
- Furnish, M. D. & Bassett, W. A. 1983. Investigation of the mechanism of the olivine–spinel transition in fayalite by synchrotron radiation. *J. geophys. Res.* **88**, 10,333–10,341.
- Means, W. D. & Jessell, M. W. 1986. Accommodation migration of grain boundaries. *Tectonophysics* **127**, 67–86.
- Means, W. D. & Rec, J. H. 1988. Seven types of subgrain boundaries in octachloropropane. *J. Struct. Geol.* **10**, 765–770.
- Paterson, M. S. & Weiss, L. E. 1961. Symmetry concepts in the structural analysis of deformed rocks. *Bull. geol. Soc. Am.* **72**, 841–882.
- Paterson, M. S. & Weiss, L. E. 1966. Experimental deformation and folding in phyllite. *Bull. geol. Soc. Am.* **77**, 343–373.
- Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.
- Ramsay, J. G. and Huber, M. I. 1983. *The Techniques of Modern Structural Geology, Volume 1: Strain Analysis*. Academic Press, New York.
- Sander, B. 1970. *An Introduction to the Study of Fabrics of Geological Bodies* (English edition, translated by Phillips, F. C. & Windsor, G.). Pergamon Press, Oxford.
- Simpson, C. & Schmid, S. M. 1983. An evaluation of criteria to deduce the sense of movement in sheared rocks. *Bull. geol. Soc. Am.* **94**, 1281–1288.
- Tullis, J. & Yund, R. A. 1987. Transition from cataclastic flow to dislocation creep of feldspar: mechanisms and microstructures. *Geology* **15**, 607–609.
- Turner, F. J. and Weiss, L. E. 1963. *Structural Analysis of Metamorphic Tectonites*. McGraw-Hill, New York.
- Woodward, N. B., Boyer, S. E. & Suppe, J. 1985. An outline of balanced cross-sections. *Studies in Geology* 11. Department of Geological Sciences, University of Tennessee.