PERGAMON

Journal of Structural Geology 21 (1999) 959–967

JOURNAL OF STRUCTURAL GEOLOGY

www.elsevier.nl/locate/jstrugeo

Displacement control of geologic structures

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Received 4 March 1998; accepted 2 February 1999

Abstract

Structural geologists routinely undertake geological analyses, particularly studies of faulting, by assuming that applied stresses are the controlling parameters. An alternative view is the assumption that material velocities, incremental displacements, or total displacements are imposed on the system, with stresses then part of the material response to these imposed boundary conditions. In our view, taking velocities and displacements as independent variables in deformation and stresses as dependent variables requires fewer assumptions and is more consistent with the observed geology.

1. Introduction

There are two distinct ways of viewing the development of geological structures. In the first, one envisions applied stresses as *independent* parameters and material velocities, incremental displacements, or total displacements as a *dependent* material response. In the other, material velocities, incremental displacements, or total displacements are independent parameters, and stresses are one part of the dependent material response. These approaches tend to be relegated to certain sub-fields in structural geology. Stress-based approaches are typical of rock mechanics studies of geological features. In a typical rock-mechanics approach one might attempt to predict, using the Mohr–Coulomb criterion or a modification of it (Jaeger, 1969; Jaeger and Cook, 1976; Paterson, 1977; Reches, 1978; Reches, 1983; Reches and Dieterich, 1983; Suppe, 1985), the geometry of the faults or fractures that will form in rocks that experience only small, often recoverable, strains. Thus, rock mechanics approaches to the origin or propagation of fractures and faults are predicated upon the idea that fault or fracture arrays are a dependent material response. This approach has been particularly successful in analyzing which far-field applied stresses will lead to the propagation of fractures through an isotropic medium (see Atkinson, 1987; Ingraffea, 1987). This approach is, in our view, less successful in attempting to infer the stress system responsible for the formation of the faults. Tectonics, particularly analyses of ancient orogens, relies primarily on the displacement approach. The success of analyses of Mesozoic Cordillera tectonics has been due in large part to the assumption that plate motions can be inferred reliably (e.g. Engebretson et al., 1985, for western North America) and that they have a direct relation to the observed geology.

Most structural geologists adopt aspects of both approaches, considering both the strains (or displacement field) 'accommodated by' a given structure and the stresses 'responsible for' generating the structure. This language indicates that while we consider both sets of parameters we implicitly lean towards an assumption that stresses are independent parameters. Returning to the example of faults and faulting, we determine, on one hand, the geometric character of individual faults, their displacements or offsets (normal, oblique-slip, etc.), and consider the character of the overall deformation accommodated by faulting (crustal

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0191-8141/99/$ - see front matter
PII: S0191-8141(99)00045-0

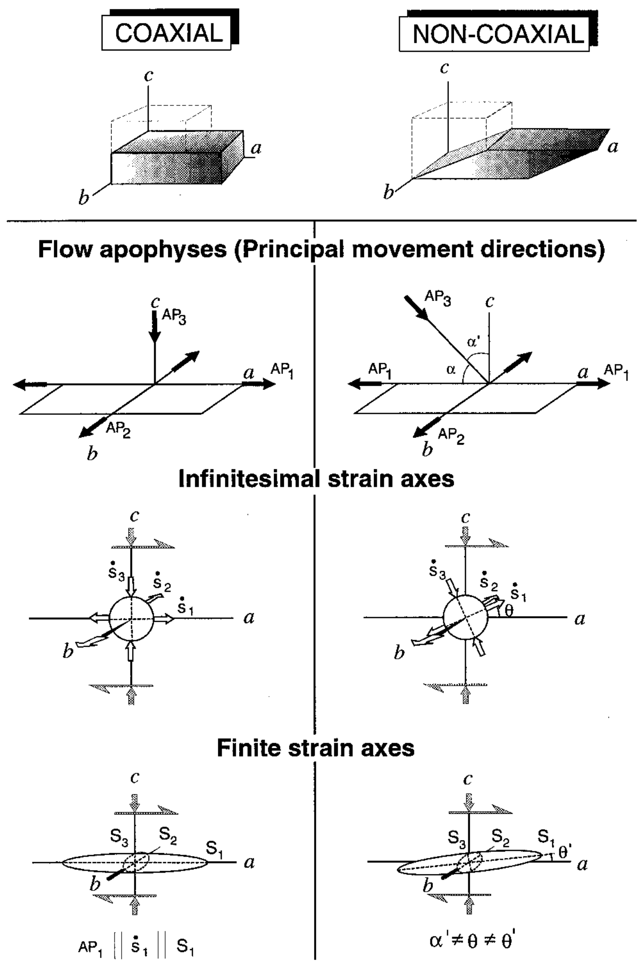

Fig. 1. The distinction between coaxial and non-coaxial deformations. For coaxial deformations, the flow apophyses (AP); infinitesimal strain axes ($\dot{s}$), and finite strain axes (S) are parallel throughout deformation. For non-coaxial deformation, these parameters all generally have different orientations.

extension, crustal shortening, etc.). On the other hand, we examine the origin of faults or arrays of faults in relation to stress distributions. We view nearly all other structural elements, folds, foliations, lineations, and even igneous intrusions, in similar ways. This tendency to examine the displacements 'accommodated by' structures and to consider the stresses 'responsible for' them probably results from the success of experimental analyses of rock deformation. In most experimental studies (e.g. Tullis and Tullis, 1986), stress is taken as the independent variable, with strains or strain rates taken as the dependent material response. We note, however, that even in these cases the movement of a piston or torsion of a cylinder imposes a displacement that we conceive to be a stress applied to the sample.

The constitutive relationships that govern deformation relate stresses to strains or stresses to strain rates, and a complete solution to any deformation problem requires an understanding of both stress and velocity fields (see Pollard and Segall, 1987). Our point here is to encourage critical examination of the character of the boundary conditions in natural deformations. One can envision comparable boundary value problems in which imposed stresses or forces, or imposed displacements or velocities, constitute the boundary conditions, or boundary conditions may consist of a mix of imposed tractions and velocities (Ford and Alexander, 1963, p. 499). In natural settings, as in boundary value problems, one of these situations will pertain, i.e. either stresses, velocities and displacements, or some combination of the two will be independent variables. The question we ask is: are stresses or are velocities, incremental displacements, and total displacements the more appropriate choice as an independent parameter in analyzing the development of structural features? In attempting to provide a context in which one may answer this question, we examine whether: (1) we can obtain useful information about stresses from deformed rocks; and (2) analyses of stresses yield information that we cannot obtain more reliably from other analytical approaches.

2. Coaxial vs non-coaxial deformation

In strain analyses, we commonly distinguish between coaxial and non-coaxial deformation (Fig. 1). This distinction has significant implications for whether one infers that stresses or displacements control deformation. If the infinitesimal strain (or stretching) axes are parallel to the finite strain (or stretching) axes, the deformation is coaxial (e.g. Means, 1976; Lister and Williams, 1983) (we do not consider here the possibility of an external rotation of reference frame or spin). For coaxial deformations, the principal movement directions (flow apophyses or eigenvectors of the velocity gradient tensor) are parallel to both the infinitesimal and finite strain axes and there is no internal (or shear-induced) vorticity to the system. Qualitatively, coaxial systems are relatively independent of the boundary conditions, since specifying either the directions and magnitudes of the compression and tension axes or the directions and magnitudes of the shortening and elongation axes necessarily gives the other.

Non-coaxiality refers to a lack of parallelism between the principal infinitesimal strain (or strain rate) axes and finite strain axes (Means et al., 1980); non-coaxiality includes but is not limited to simple shearing. For non-coaxial deformations, the principal movement directions (flow apophyses) are parallel to neither the infinitesimal nor finite strain axes, and there is a shear-induced vorticity (Fig. 1). Non-coaxial deformations are a natural consequence of interactions

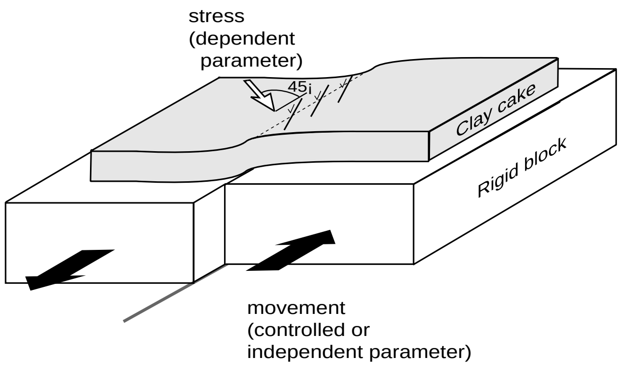

Fig. 2. A cartoon of the clay cake experiments of Riedel (1929). Contrary to the typical interpretation, displacement was the controlled (independent) variable in the experiment and the stresses were the dependent variable.

between rock anisotropy and minor perturbations in regular three-dimensional flows, and are the rule in natural deformation settings. Non-coaxial flow problems require more thorough specification of boundary conditions, because of the rotational component of deformation, if one wishes to understand either the kinematics or dynamics (e.g. Tikoff and Fossen, 1993).

Therefore, we believe that the distinction between control by stresses or control by displacements will ultimately be resolved in the context of examining non-coaxial deformations. One cannot distinguish between the nature of the boundary conditions in coaxial deformations because all the principal flow parameters are parallel. Note incidentally, that stress-based approaches have tended to be more successful in examining deformations subjected to coaxial boundary conditions (e.g. Mode 1 fractures, anti-cracks), although one can examine these structural elements in a context of displacement or velocity fields (see Pollard and Segall, 1987). Non-coaxial deformations are not well defined by specifying principal stress directions, probably because stress analysis does not account for the requisite finite rotation. This effect of finite rotation is that principal stress axes and principal finite strain axes are not parallel, and may even be orthogonal, for non-coaxial deformations (Tikoff and Teyssier, 1994). Additionally, non-coaxiality is often associated with material anisotropy (shear zone boundaries) and strong local displacement or velocity gradients, both of which require variations in stress orientation and magnitude. The common occurrence of non-coaxial deformation zones, such as ductile shear zones, suggests that imposed velocity boundary conditions control the deformation whereas stresses adopt orientations and magnitudes that conform with these conditions.

3. Displacement control of experimental deformation

Despite the presumption that stresses are independent parameters, it is generally the displacement or the velocity of a plate, bounding surface, piston, or cylinder that is imposed, i.e. that is the independent parameter, in the majority of deformation experiments. Normally, we infer stresses from the boundary displacements through an understanding of the material properties of the test rig and the sample. For example, the experiments of Cloos (1928) and Riedel (1929) created the first well-documented sets of en échelon fracture systems (Fig. 2). In both cases, the experimental setup consisted of a clay cake placed across two adjoining boards, with one board made to slide slowly past a stationary adjacent board. Obviously, in these experiments, the movement of the board was imposed. Stresses necessarily arise from this movement through a material contact between the board and the clay cake, and were therefore the dependent variable. Cloos and Riedel both inferred that the stress principal axes were oriented at 45° to the boundary between the two boards, and evaluated the deformation in terms of calculated stress magnitudes. Their intuition might be correct, but the imposed displacements were the independent parameters in these experiments. Other critical experiments on the origin of faults conducted by Oertel (1965) were coaxial (see below), so no clear distinction existed between principal stress axes and principal directions of the displacement field.

Many argue that investigators control stresses in high pressure and high temperature deformation experiments (e.g. Tullis and Tullis, 1986). We suggest, however, that because these deformations are generally coaxial, with principal movement directions (eigenvectors of the velocity field) parallel to bulk principal stress axes, the relative importance of the two cannot be evaluated (Fig. 3a). In our view, Zhang and Karato (1995) conducted a critical experiment in which a bounding surface was cut at a 45° angle to the direction of piston movement. In their experiments, shown in cartoon form in Fig. 3(b and c), deformation is non-coaxial. If the rock deformations were stress-controlled, imposing σ_1 at 45° to the rigid boundary should lead to simple shear kinematics. Their data indicate, however, that the material sheared and flattened, indicating that deformation was at least partially velocity controlled. Thus, in general specifying the principal stress directions does not completely constrain the deformation.

4. Can we obtain the orientations and magnitudes of stresses?

A critical question about any parameter that we

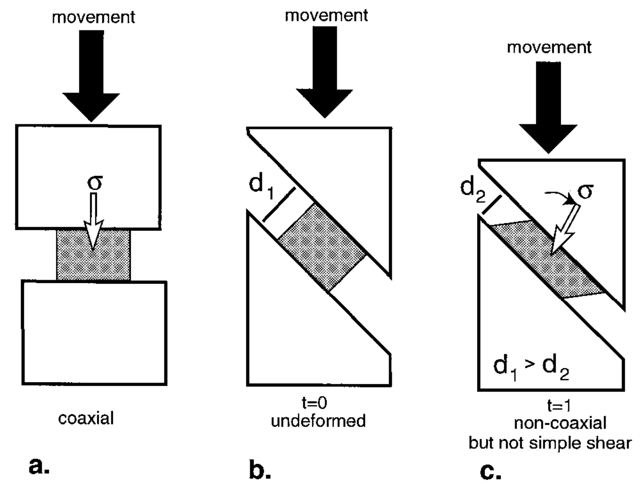

Fig. 3. A cartoon of the experimental deformations. (a) In coaxial deformation experiments, the direction of movement of the driving piston is parallel to the principal stress direction, which leads to the inference that stress is the dependent variable. (b) and (c) The non-coaxial experiments of Zhang and Karato (1995) indicate a component of displacement control. If the stresses were completely controlling deformation, the sample (in gray) would experience simple shear deformation in response to σ_1 fixed at 45° to boundary. The observed shortening perpendicular to the boundary (d) requires that the deformation was not simple shear but included shortening across the shear plane.

choose to study is whether we can measure it. Strain is clearly a measurable parameter, in both ancient orogens using the techniques of structural geology and in active tectonic settings using geodetic or satellite data and other empirical techniques. By combining either set of observations with different temporal measures, we can sometimes estimate displacement rates.

Measuring stress in rocks is significantly more difficult. All techniques by which structural geologists determine the orientation and/or magnitudes of components of stress tensors that operated during deformation depend upon the recognition of a distinctive incremental strain or finite strain features in rock, i.e. deformation twins in deformed carbonate mineral grains, deformation lamellae in deformed quartz grains, kinked mica grains, subgrains in plastically deformed minerals, or lineations on faults (e.g. Carter and Raleigh, 1969; Carey and Brunier, 1974; Jamison and Spang, 1976; Angelier, 1979; Carey, 1979; Etchecopar et al., 1981; cf. Arthaud, 1969; Newman, 1994). All of the techniques assume that the deformation adhered to the Lévy–Mises equations (Ford and Alexander, 1969, p. 410). Stated in a different way, one must assume that the principal directions and magnitudes of the incremental strain tensor are, respectively, parallel and proportional to the principal directions and magnitudes of the stress tensor (Wojtal and Pershing, 1991; Twiss and Moores, 1992, p. 410). This assumption is reasonable and justified in many geological settings, but it does require that the deforming material is isotropic. Furthermore, most piezometric techniques require that strains result from a single, coaxial deformation event. The effects of anisotropic materials (e.g. Lan and Hudleston, 1997), material anisotropy due to differences in grain orientation or grain size (e.g. Newman, 1994) or to differences in material behavior (e.g. Pollard et al., 1993), or non-coaxiality (Wojtal and Pershing, 1991) may invalidate the assumption that the deformation adhered to the Lévy–Mises equations and lead to unreliable results from the application of these techniques.

Another practical problem arises in applying some paleostress measurement techniques. In order to collect a sufficient number of readings of fault and slip lineation orientations to yield statistically reliable results from any one of several fault slip 'paleostress' techniques, one may need to compile data from regions in which stresses are not homogeneous. This problem may plague any paleo-piezometric technique, even recrystallized grain size methods (cf. Handy, 1994). Further, stresses may vary significantly during the development of an apparently simple coaxial deformation (see Oertel, 1965), even as the incremental strains appear to remain uniform.

Neotectonic measurements of stress include earthquake moments and borehole breakouts (e.g. Zoback et al., 1989). In the case of earthquake moments, the principal stress directions are inferred from an instantaneous displacement. Borehole breakouts also require observation of the elongation of the borehole, an incremental strain. One can *interpret* these data in terms of stresses, but it is strains that are measured (e.g. Twiss and Moores, 1992; Tikoff and Fossen, 1995). Moreover, as documented in several studies, there are large spatial variations in the orientations of stress principal axes, which raises the questions of the validity of any 'regional' stress (Rebai et al., 1992). Yet, when interpreted in terms of infinitesimal strains or small finite strains (i.e. incremental strains), the approach remains useful (e.g. Avé Lallemant and Gordon, in press) because the interpretation (strain) is based directly on the observations (displacement).

Similar effects occur at smaller scales in deforming rocks. In faulted rocks, larger faults often have numerous smaller splays with synthetic or antithetic geometries, and the smaller faults often have even smaller synthetic or antithetic splays (e.g. Wojtal, 1986). If one starts from a premise that faults form with a specific orientation with respect to the stress principal directions, one finds that the orientation of the stress principal directions one infers depends upon the scale of observation. The situation here is akin to the 'Riedel in Riedel' pattern described by Arboleya and Engelder (1995). Both of these patterns lead to confusion in defining the stresses in these rocks, for as one shrinks one's field of view, the magnitudes of traction vectors apparently do not approach limiting values. We believe, therefore, that it is more sensible to consider that the kinematics of the deformation, not the orientations of stresses, exert primary control on the geometry of faults in these settings, for one can choose a scale of observation at which the displacement or velocity fields are smooth and quasi-continuous (e.g. Wojtal, 1989).

Finally, wide variations in inferred stress magnitudes and principal directions occur because of inhomogeneities within regions subjected to quasi-continuous deformation (e.g. Handy, 1994; Newman, 1994). Variations in the magnitudes and orientations of principal stresses in small regions suggest to us that: (1) stresses will attain whatever value is required to deform rocks at the imposed strain rates, that is that stresses are dependent variables; and (2) it is very difficult to determine the stresses, since stresses do not readily approach homogeneous, limiting values as the field of view is reduced.

5. Do 'bulk' or 'regional' stresses exist?

If stress does control the resultant deformation, it is critical to understand how the stresses are oriented at different points in space. The inference that stresses control deformation leads naturally to the making of stress maps (e.g. Zoback et al., 1989). One difficulty with such generalized stress maps is that stress distributions are likely to be modified by existing structures, leading to local stresses that diverge significantly from regional patterns (e.g. Rebai et al., 1992). A striking example is an inferred component of *sinistral* tangential stress across the *dextral* San Andreas fault system, in the vicinity of the Big Bend in the Transverse Ranges (Zoback and Healy, 1992; Scholz and Saucier, 1993). In light of this inference, it is difficult to imagine that stresses are the controlling parameter in the local deformation.

The world stress map of Zoback (1992) shows that stress orientations are constant over large areas of continents and often parallel to plate motions. This suggests that the movements of the tectonic plates control the deformation and that stresses attain whatever value is needed to accommodate the displacements. In broad, relatively isotropic regions where deformation is coaxial, such as continental interiors, we expect a high degree of conformity between the incremental displacement directions (or the material movement) and stresses (Zoback et al., 1989). The effect of non-coaxial deformation, perhaps due to inherent anisotropies that are in turn due either to pre-existing features or to features generated early in a protracted deformation, is to locally modify this very straightforward result. In non-coaxial deformation settings, we believe that material movements are the independent parameters and that the stresses are the dependent response. For example, the observation of sinistral resolved stresses across the San Andreas fault system make some intuitive sense—stresses respond to the local heterogeneities (Scholz and Saucier, 1993) as the displacements control the deformation of the system.

6. Lack of principal stress planes

Treagus and Lisle (1997) demonstrated that the planes of principal planes of stress and strain do not exist for all geological deformations. In deformations such as the torsion of a circular cylinder, one cannot define principal planes of stress. Yet, for each of these deformations, one can define continuous displacement or velocity fields. For the case of the torsion of a cylinder, for instance, the displacement vectors are segments of circular arcs about the axis of the cylinder. Similar conclusions apply in a variety of realistic tectonic scenarios that lack well-defined principal planes (Treagus and Lisle, 1997).

If continuous principal surfaces of stress are not defined but the displacement field is continuous, we suggest reevaluating fractures that formed in these settings. Principal stresses can be defined at every point, and thus contribute to the formation of *individual* fractures. The resulting *array* of fractures cannot, however, be considered as evidence of a 'regional' stress field because such a regional stress field does not exist. We anticipate the occurrence of a relationship between the geometry of the arrays and the character of the three-dimensional displacement field.

7. 'Just' kinematics

Stress is a critical parameter in the development of structural elements. Available geological data strongly support, however, the notion that stresses vary radically from place to place as a result of the imposed displacements. In experimental deformations on which many tectonic interpretations are based, displacements and velocities are demonstrably the independent variables and stresses the dependent variables (Riedel, 1929). Treated in this way, the large stress variations in Oertel's experiments (Oertel, 1965) or sinistrally resolved shear stresses adjacent to the San Andreas fault (Zoback and Healy, 1992) make sense—stresses adjust to accommodate the imposed displacements.

Lithospheric strength profiles may also support the notion of displacements as the independent variable in upper-crustal, tectonic deformation. If one accepts that the upper mantle is the strongest layer of the lithosphere, it presumably controls the bulk response of the lithosphere. In this scenario, within the lithospheric mantle, the relation between displacement and stress is complex and no distinction between independent and dependent variables can be made. However, this is not true of the overlying lithospheric layers, including the upper crust. In analogy with the Reidel experiments, the weaker overlying crustal layers (i.e. the clay) are controlled by a velocity field imposed from below by the lithospheric mantle (i.e. the boards). Consequently, for tectonic deformations, the crust may simply respond to the imposed velocities from below (Molnar, 1992; Teyssier and Tikoff, 1998).

If structural geologists adopt this outlook of displacements as the independent parameters, we are left with investigating 'just' kinematics. There is an inherent bias against kinematic analyses (see, for example, Johnson and Fletcher, 1994, p. 71), perhaps based on the misconception that these analyses are necessarily more qualitative. An appropriate analogy to counteract this bias is finite element models. In these numerical models, one chooses between stress-based boundary conditions and velocity-based boundary conditions. The choice depends on the structural or tectonic problem that is modeled, and that choice often dictates the outcome. Neither approach is more quantitative nor more correct; as critical observers it is up to us to choose which parameter is a better descriptor of the controlling parameter on the scale of the model.

One distinct advantage of studying kinematics is its relative simplicity. Kinematics is an accurate description of a physical system, with many fewer assumptions than an interpretation derived from an inferred stress distribution. We do not need to assume forces or rheology (constitutive relations), both of which are required by dynamic models and are poorly known for naturally deformed rock. In contrast, if we can evaluate, objectively and thoroughly, the kinematics of a deforming zone, we have a chance to constrain the rather simplistic geological assumptions that are presently found in dynamic models. The kinematics of deformation will ultimately provide constraints on realistic tectonic models, provided that we understand and have documented three-dimensional kinematics. At present, earth scientists have barely enough data to evaluate the three-dimensional kinematics, even in well-studied areas.

8. The generality of displacement/finite strain analysis

It is routinely assumed that if the orientations of the principal stress axes are known, the deformation of the material is defined. This assumption is incorrect for three-dimensional deformations, even for very simplistic models of rock rheology (Fig. 4; Tikoff and Fossen, 1995). Consider, for example, a material that follows a simple constitutive relationship, namely that of homogeneous, linearly viscous (Newtonian) flow. The orientations of the infinitesimal strain principal axes are directly correlated to the orientations of the stress principal axes. If deformation is steady-state, the infinitesimal strain principal axes and the stress axes remain parallel throughout deformation.

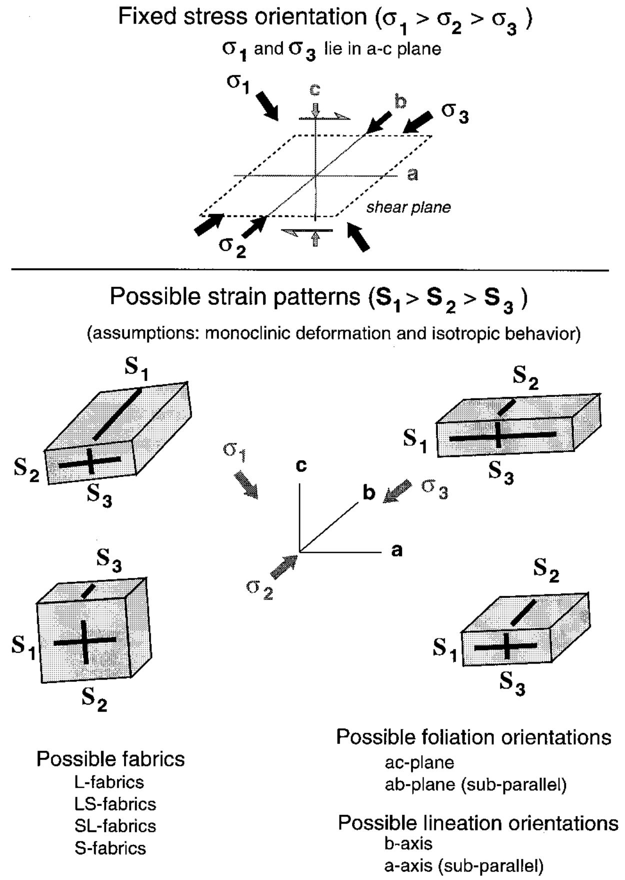

Fig. 4. Possible finite strain orientations resulting from fixed orientations of the stress principal directions and a shear plane. σ_1 makes a 50° angle with the kinematic *a*-axis (in *ac*-plane), σ_2 is parallel to the kinematic *b*-axis, and σ_3 makes a 40° angle with the kinematic *a*-axis (in *ac*-plane). Three distinct orientations of finite strain axes, two distinct foliation orientations, two distinct lineation orientations, and any finite strain ellipsoid geometry (from pure flattening to pure constriction) are possible.

Even in a volume constant, plane-strain (two-dimensional) deformation, the orientations of the stress principal axes do not uniquely determine the nature of non-coaxial deformations without specification of a shear plane. For a given orientation of the principal stress axes in two-dimensions, deformation can proceed by either pure shear, simple shear, or some combination. Thus, in order for the stress axes to uniquely specify a deformation, one must specify the boundary conditions (i.e. shear plane orientation) (Tikoff and Fossen, 1993, 1995).

Tikoff and Fossen (1995) show that even for simple three-dimensional deformations, defining the shear plane and the orientations of the three principal stress axes does not uniquely define the character of the deformation. Figure 4 gives an arbitrary orientation for stress axes and a shear plane, with σ_1 inclined 50° to the kinematic *a*-axis and lying in the *ac*-plane, σ_2 parallel to the *b*-axis, and σ_3 inclined 40° to the *a*-axis and lying in the *ac*-plane. There are an infinite number of deformations that can result from the stress state, including sub-simple shearing (plane strain), transpression, and simultaneous flattening and simple shearing. The orientation of finite strain axes and the displacement field are unknown. Knowledge of the stress principal axis orientation is not sufficient to distinguish between the three-dimensional deformations and the resultant rock fabric (e.g. foliation and lineation). Different deformation cases are distinguishable only if the orientation and absolute magnitude of the infinitesimal strain axes are also known. In terms of stress, this requires an exact knowledge of orientation and *magnitude* of principal stress axes, in addition to a well-defined rheology. This type of information is practically impossible to retrieve from naturally deformed rock.

Consider the reverse case, where the orientation and magnitudes of finite strain axes are known. Using either: (1) a shear plane orientation, a finite strain measurement, and assuming steady-state deformation; (2) gradients in finite strain (Elliott, 1972); or one can calculate the strain history and infinitesimal strain quantities of the deformation (Tikoff and Fossen, 1995). Assuming that the stress principal axes are parallel to the infinitesimal strain principal axes, the stress axes orientations are retrievable from finite strain, but not vice versa.

The same is true of any other infinitesimal strain quantities: knowledge of the infinitesimal strain axes, flow apophyses, or kinematic vorticity does not uniquely define a three-dimensional deformation, even for steady-state deformation (Tikoff and Fossen, 1995). Contrarily, if either deformation path (progressive finite strain) or boundary conditions are known, infinitesimal strain quantities (or stresses, by assuming a particular constitutive relationship) can be determined from the finite strain if steady-state deformation is assumed. Thus, in three-dimensions, the finite strain and displacement vectors are a better descriptor of rock deformation and more useful than information on the orientations of the principal axes of the stress tensor.

8.1. Rock anisotropy

The statements made in the above analysis presume that the deforming material is isotropic. The principal axes of the infinitesimal strain tensor and the principal axes of the stress tensor are parallel and proportional in magnitude in isotropic materials (e.g. Carter and Raleigh, 1969; Elliott, 1972). However, rocks are seldom isotropic and their material behavior is strongly dependent on their anisotropy (e.g. Donath, 1961). In the case of deformed anisotropic rock, we do not have, at present, general techniques for determining stresses. Stress analyses conducted in anisotropic rock do not yield consistent results. Variations in the orientations of infinitesimal strain principal axes in these cases may indicate real variations in stress magnitudes and orientations in deforming materials, but we have no reliable way to separate significant variations from measurement errors. On the other hand, analyses of displacement fields or strain variations in the same examples yield very useful and predictive information (Dennis and Secor, 1990; Pray et al., 1997). We therefore believe that the displacement field approach is an appropriate way to address the mechanical response of these systems.

9. Conclusions

Our intention with this contribution is to make three points. First, we believe that the structural geology community needs to distinguish between observations and inferences in the matter of displacement and stresses, and to use that distinction to identify objectively which parameters are independent and which are dependent. In many cases, such as the physical experiments of Riedel (1929), displacements are clearly the independent variables. We draw the same conclusion in analyzing most in-situ stress measurements, deformation experiments, and paleostress analyses. There exist, in our view, compelling data suggesting that material velocities are independent parameters in some natural deformations. Of course, geologists must examine each example critically but with an open mind. Second, we hope to counteract the emphasis put on stresses and dynamic analyses, with little attention paid to either real structures or the geologically obtainable displacement field. This is particularly true of regional-scale approaches to tectonic problems. Earth scientists may not understand what actually controls plate tectonics for some time, but we can certainly constrain the debate by accurately describing the kinematics. Third, structural geologists are now at a stage where we need to address the three-dimensional character of deformation, both in our kinematic analyses and in our dynamic models. Many of the shortcomings of our present understanding are directly due to a lack of thinking about deformation in three dimensions. Analyzing deformation in three dimensions often leads to surprising and counter-intuitive results (e.g. Fossen and Tikoff, 1995; Treagus and Lisle, 1997). We need, as a community, to document how rocks actually deform, rather than analyzing how we think rocks might deform.

Acknowledgements

We thank Michelle Markley for a very helpful review of an early draft of this manuscript. Richard Lisle's careful review and thought-provoking comments led to significant improvements, and we thank him. Likewise, Jim Evans' comments led to substantial improvements for which we are grateful. Steven F. Wojtal thanks Wendy Kozol for her support and Ronnie Kusznir for her assistance in the final preparation of the manuscript. Basil Tikoff was supported by NSF EAR-9607018, and thanks H. Avé Lallemant and H. Fossen for helpful conversations.

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