DEFORMATION AND PALEOMAGNETISM

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Abstract. We may use tectonic structures to confirm the primary age of a paleomagnetic remanence component but only if we know how to undo the natural strain history. It is normally insufficient to untilt fold limbs, as in the original version of Graham's Fold Test. One may need to remove also the bulk or local strain and account for strain heterogeneities, achieved by grain-strain and the more elusive intergranular flow. Most important, one must know the sequence of strains and tilts that occurred through geological history because the order of these noncommutative events critically affects the final orientation of the remanence component.

In many non-metamorphic rocks, strain-rotation of a remanence component approximates a simple formula, although the actual rotation mechanism is complex. This simple, *passive line* approximation is confirmed experimentally for strains up to 45% oblate shortening. The passive line hypothesis has permitted successful paleomagnetic restorations in several natural case studies.

Experimental deformation of samples with multicomponent remanences shows that differential stresses above a threshold value near 25 MPa selectively remove components with coercivities <25mT, due to domain wall rearrangements in large multidomain magnetite grains. Higher coercivity components are less reduced so that the net remanence vector spins always toward the high-coercivity component, at rates and along paths not predicted by any structural geological formula. Experimentally deformed samples with very fine hematite in the matrix showed their net remanence spinning away from the high coercivity component. This is due to easier mechanical disorientation of the very fine hematite grains, scattering their magnetic moments more and reducing their contribution to the overall remanence. Thus, muticomponent remanences have their components selected for survival based on rock-magnetic *and* microstructural criteria. Such stress-rotation by coercivity selection does not depend on the orientations of the principal stresses or strains, a concept that is counterintuitive to conventional structural geology.

Syn-tectonic remagnetization is common in deformed sedimentary sequences and laboratory experiments reveal that a only moderate differential stress remagnetization is required to add components parallel to the ambient field, without significant strain. Alternating field demagnetization isolates components smeared along the great circle between the initial remanence direction and the remagnetizing field direction. In this case, the principal directions of the stress and finite strain tensors are irrelevant; remagnetization is triggered by a threshold differential stress. The final remanence direction is controlled by the ambient field direction and the remagnetization path lies along a great circle between the ambient field and the initial remanence direction.

Key words: Paleomagnetism, Fold test, strain-rotation, stress, coercivity-selection, stress remagnetization.

1. Introduction

The value of combined structural and paleomagnetic studies was first realized by Graham (1949) who devised a "fold test" to discriminate between pretectonic remanences dispersed by folding and remanences that overprint folded rocks. In fact, we can adapt any tectonic structure formed by continuum heterogeneous strain to the purpose of Graham's "fold" test. (Heterogeneous strain occurs during deformation

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in which straight lines do not remain straight and parallel lines do not remain parallel.) Therefore, many other structures provide suitable tests for the primary age of remanence. Shear zones, strain-shadows around augen, boudins or pretectonic plutons (e.g., Henry, 1992); growth faults, terminal curvature near faults, differential compaction features (e.g., supratenuous folds), kinks, diapir emplacement and fold or thrust nappes (e.g., Vetter et al., 1989) provide suitable noncoaxial, heterogeneous, continuum strain environments for a general "tectonic" test of the primary character of remanence. In this paper, I will discuss general implications of the tectonic test for primary remanence. These include the importance of untilting, destraining, noncommutative deformation sequences, and the different effects of bulk strain and grain-strain. The first part of the paper deals with rotation of remanence dictated primarily by finite strain. The second part concerns *stress*-rotation of remanence and syn-stress remagnetization.

2. Deformation Terminology

We loosely refer to all of the tectonic processes acting on a rock, at any scale, as deformation. However, confusion occurs where this umbrella term is used synonymously with strain. Essentially, deformation encompasses the following terms.

(1) **Translation**: motion without rotation or distortion from one coordinate to another.
(2) **Rigid Body Rotation**: turning of body without change in shape or translation.
(3) **Volume change**.
(4) **Strain**: change in shape, distortion. It is homogeneous where straight lines remain straight and parallel lines remain parallel, or otherwise heterogeneous. The same body may be heterogeneously strained at the grain-scale, homogeneously strained at the hand-sample scale, and heterogeneously strained at the folded outcrop scale. Slates commonly show this range of strain behavior.

Strain is a second rank tensor commonly represented by an ellipsoid with axes $X \geq Y \geq Z$. It is common practice to normalize the magnitudes of finite strain by dividing each by $(XYZ)^{1/3}$. The orientations of the X, Y and Z axes are normally symmetrically arranged with respect to petrofabric elements where the strain history is coaxial. Thus, for a single episode of tectonic deformation the XY plane may be parallel to schistosity and X parallel to a mineral lineation. Natural regions of markedly noncoaxial strain history have X, Y, Z directions that change both temporally and spatially through the rock-volume giving heterogeneous strain. The vicinities of fold hinges, shear zones, faults, boudin necklines and strain shadows near plutons are examples of regions that show such complex strain histories and trajectories. Ellipsoid-symmetry is represented by T where $1 > T > 0$ for disk shapes and $0 > T > -1$ for rod-shapes (Jelinek, 1981). $T = 0$ corresponds to neutral ellipsoids with $X/Y = Y/Z$, the *plane strain* [sic] ellipsoid

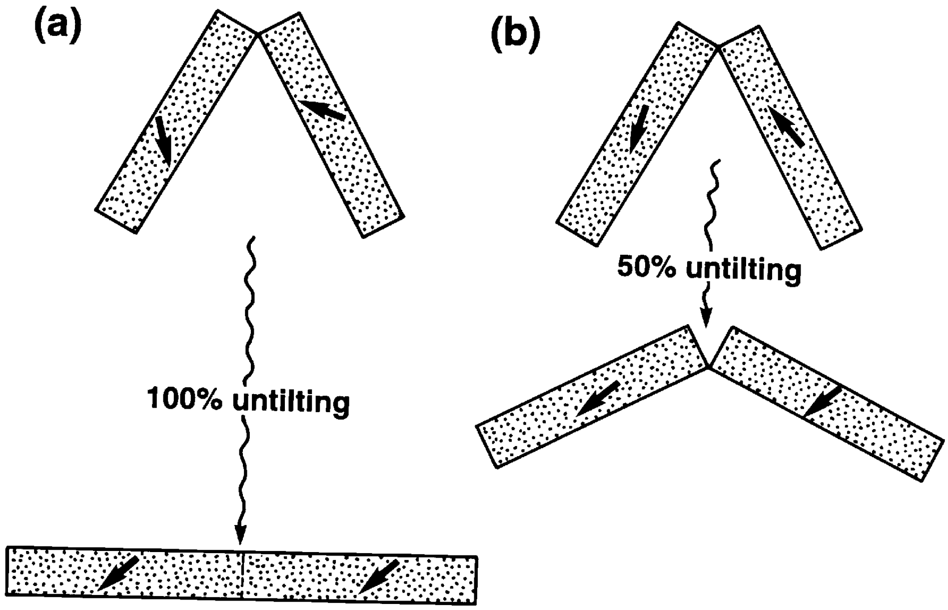

Figure 1. In Graham's fold test, remanences dispersed by limb rotation will be restored to a common direction when the layer is straightened (a). This is not "unfolding" but merely untilting of tilted layers. In real folds the limbs are always strained internally.

of structural geology. This convention illustrates finite strain better than the Flinn diagram because the two axes separately represent symmetry and intensity of strain. Moreover, the symmetry or shape representation is symmetrical from -1 to $+1$ rather than from zero to infinity for the Flinn diagram. Jelinek's scheme is used to illustrate both the anisotropy of magnetic properties and finite strain states.

Of these four elements of deformation, strain is most commonly measured by structural geologists. On the other hand, paleomagnetists measure translations (from paleolatitude estimates) and rotations (from changes in paleodeclination or paleoinclination). Despite the similar approach of evaluating deformation events these two types of study are rarely integrated.

At first, let us note the elegant simplicity of Graham's fold test (Figure 1). The remanence is primary if the vectors in both limbs become parallel when we turn the limbs into horizontal parallelism. This robust test has served us well and will continue to do so. However, it assumes the limbs of the fold are completely unstrained. In fact, no fold can be created without some strain of the limbs. Fortunately, it may be small enough to ignore for the purposes of paleomagnetic fold tests in feebly folds, preferably of chevron style. Note that the restoration involves untilting (a rigid-body rotation) not unfolding (Figure 1a) which would require the removal of heterogeneous strain. In most real folds, remanence-restoration requires destraining (Facer, 1983; van der Pluijm, 1983; Kodama, 1988).

As paleomagnetic work in deformed terrains became more commonplace, it was found that untilting rocks (usually still misnamed “unfolding”) sometimes produced a worse dispersion of the remanences. However, partial untilting often restored the remanences to parallelism (Figure 1b) (Bachtadse et al., 1987; Schwartz and van der Voo, 1984; Scotese et al., 1982). This lead to the common assertion that the remanences were synfolding, established at a certain point during the folding episode (e.g., McLelland-Brown, 1982; Miller and Kent, 1986; Stead and Kodama, 1984; Kodama, 1988; Stamatakos and Kodama, 1991a). Occasionally, the reasons for syn-folding remanences are well documented as chemical events triggered during folding (e.g., Benthien and Elmore, 1987; McCabe et al., 1983). However, we will see later that there may be good microstructural reasons for establishing syntectonic remanence, part way through a folding episode. However, one notes that the most common restorations still use only simple untilting, not unfolding. This will always introduce errors in the restoration process if finite strains are neglected, no matter how sophisticated the statistical treatment of the untilted remanence directions (e.g., McFadden, 1990; McFadden and Jones, 1981; Bazhenov and Shipunov, 1991).

A simple demonstration that untilting may be inadequate is shown by kink folds (Figure 2), in which the limbs show limited internal deformation of their flanks. Such folds are normally restricted to shallow depths (<5 km) because their formation requires some dilation at hinge zones. Various mechanisms are known for the formation of kinks (e.g., Ramsay, 1967; Ramsay and Huber, 1983; 1987), but a common type penetratively shears the kinked portion as shown in Figure 2. Thus, a modestly steep remanence outside the kink will always be shallower with respect to bedding inside the kink zone because of strain. The amount of shallowing varies with the passivity or rheology of the marker that carries the remanence. Simply untilting, until the layer is completely horizontal, produces “two” remanence directions, at least one of which could be a tectonic artefact as in Figure 2. This extreme example illustrates potential pitfalls of an oversimplified restoration.

How can we improve the restoration process? One should first realize strains are not commutative. The outcome depends on the order in which nature combines successive strains, or strains interspersed with rotations (Lowrie et al., 1986; Borradaile and Mothersill, 1989). Figure 3 shows a steeply tilted, strained bed. In (a) the geology is first destrained, then untilted. The restored remanence appears perpendicular to bedding. In (b) untilting precedes destraining and the restored remanence is inclined near 45° to bedding. Clearly, we cannot restore remanences arbitrarily, without some knowledge of the sequence of events in nature. Each case history should be evaluated separately. Chevron folds and compressed buckles provide two common examples that we can tackle. Figure 4a illustrates the self-explanatory restoration of a chevron fold, only requiring untilting. Figure 4b shows a common situation in which a layer buckled slightly. Subsequently bulk strain amplified its amplitude, the fold limbs behaving almost as passive markers.

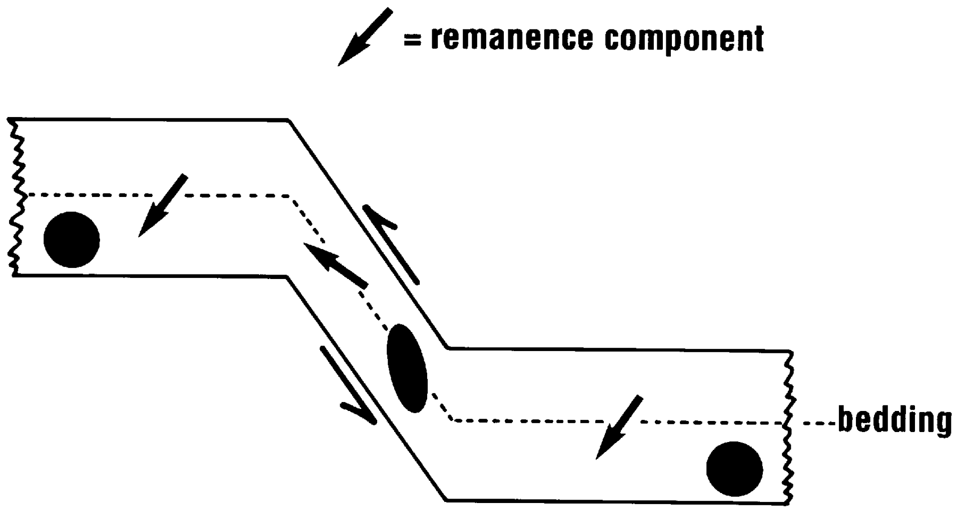

Figure 2. A kink-zone or kink fold shows the complications arising from strain combined with tilting. The tilted, kink-zone has a sinistral shear and the resulting strain (ellipse in black) reduces the remanence-inclination with respect to the bedding. Here, untilting will never restore the pretectonic remanence to a uniform direction.

(Passive markers have no rheological contrast with their matrix.) This sequence could be reversed, yielding a successful restoration of remanence.

The following examples are oversimplified; deeper treatment of specific examples is found in the literature (e.g., Facer, 1983; Kodama, 1988; van der Pluijm, 1987). Nevertheless, the simple examples below show some commonly observed structural features that may complicate our attempts at paleomagnetic restoration. Perhaps the reader will forgive the simplicity of the two-dimensional sketches and a single pretectonic remanence inclined near 45° to bedding.

A very common situation in deformed rocks is that early, low amplitude buckles are subsequently strained passively (Figure 5). In the initial buckling episode, the internal strain of the limbs is negligible. Thus, the remanences are steepened on one limb and shallowed on the other limb, in geographical coordinates. Simply untilting will suffice to restore the original remanence directions at this stage (Figure 5a). The next step in the structural sequence involves the shortening of the buckle that causes significant strains in the flanks and hinge (Figure 5b). The strain on the flanks further rotates the remanence in bedding coordinates. If these two mechanisms did not overlap, remanence restoration could be achieved by reversing the sequence of operations. Removing the finite strain from the fold should recreate the buckle situation, restoring the remanences in bedding coordinates (Figure 5b), after which untilting would restore the paleomagnetic vectors to their original orientation in geographical coordinates. The main obstacles to this simple restoration procedure are, first, the episodes of buckling and fold amplification may overlap. Second, the fold amplification may not be due to passive behavior, i.e., the layers may not all

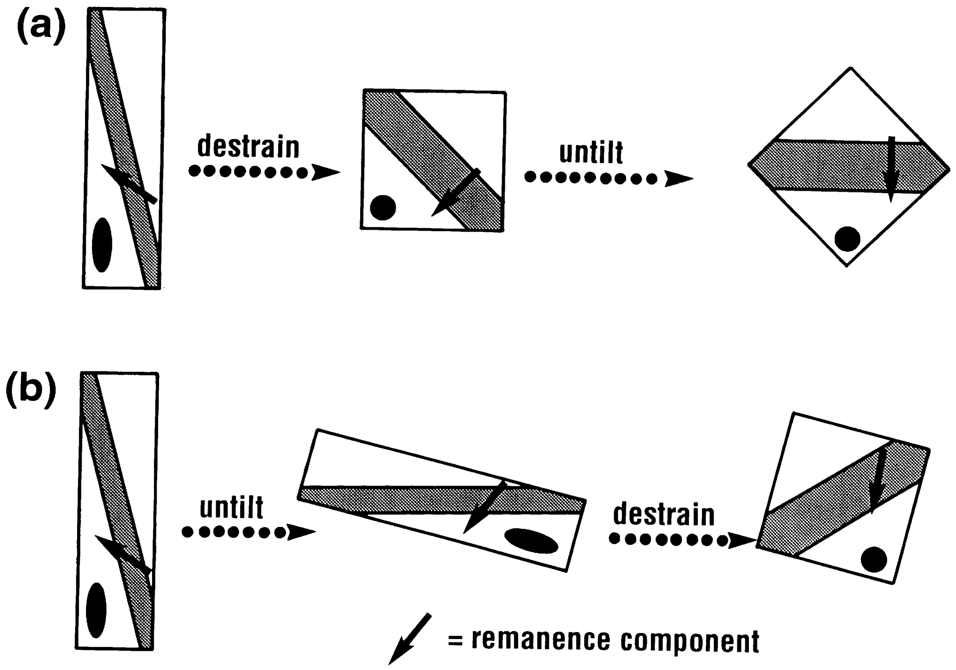

Figure 3. Successive strains and strains interspersed with tilts combine noncommutatively so that the result depends on the sequence of the deformation components. (a) and (b) show different possible restorations depending on whether the geology is first destrained and then untilted, or vice-versa. In nature, straining and tilting overlap, making the restoration even more difficult.

have the same viscosity (Ramsay 1967). There, a simple geometrical removal of some estimated bulk strain would yield an invalid paleomagnetic restoration.

Chevron folds tempt us to use a simple untilting restoration (Figure 6). However, their flanks are normally sheared, in all lithologies, to varying degrees. The tops of the layers shear toward the hinge-line of an antiform, steepening the remanence on one limb and shallowing it on the other, in bedding coordinates. Simple untilting would give the false impression of syntectonic magnetization (Figure 6b,c). Most folds in which the limbs are differently strained yield similar problems (Figure 7). For some folds, untilting would be so unsatisfactory that untilting would lead to “negative unfolding” before the remanences were restored to parallelism. The examples of Figure 6 and Figure 7 show the importance of destraining and untilting the fold’s profile. However, we may be uncertain of the temporal overlap of these stages, and of their sequence (e.g., Lowrie et al., 1986 and Kodama, 1991a).

Further problems arise where fold flanks show a continuous strain-variation. One cannot simply remove a bulk strain from the whole fold profile (e.g., Figure 5b). Instead we must determine and remove the strain at each position where we measured remanence. This is not as tedious as it appears. Structural geologists do this routinely (Ramsay, 1967; Ramsay and Huber, 1983) and it has been also done

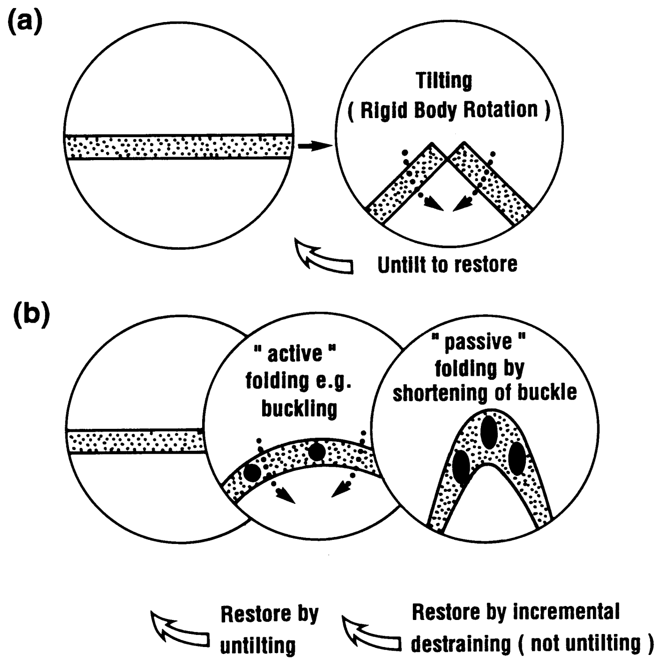

Figure 4. (a) An idealized Graham's fold test. No limb-strain is involved, thus rigid-body rotations may detect a positive fold test, or perhaps a syntectonic remanence under appropriate circumstances. (b) Tilting was an integral part of the straining process so that unrolling the limbs (untilting) is an invalid method of restoring remanences.

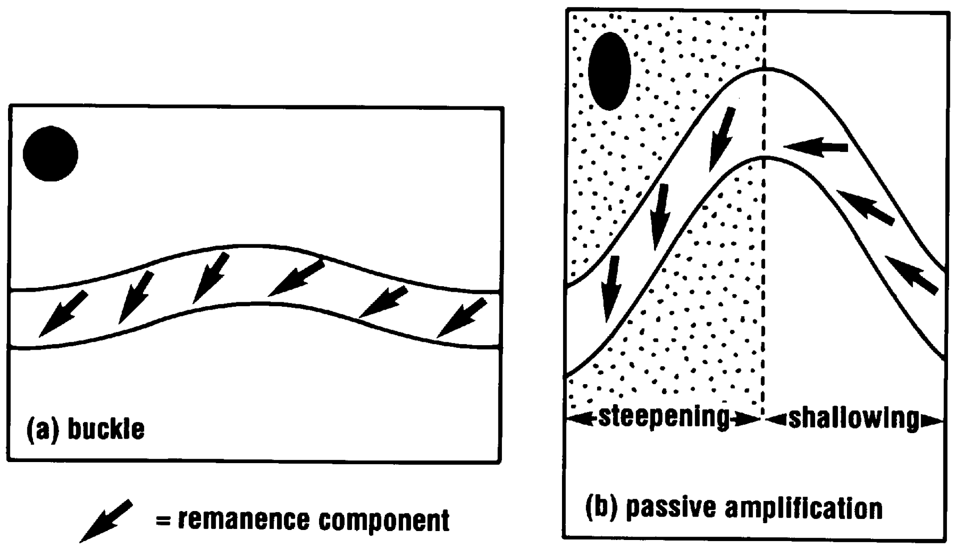

Figure 5. (a) An incipient buckle disperses a remanence component in geographic coordinates. Untillting at this stage would produce a positive fold test. (b) Bulk strain amplifies the buckle passively dispersing the remanences further in bedding coordinates. Removing the passive homogeneous strain (black strain ellipse) would restore the situation to (a).

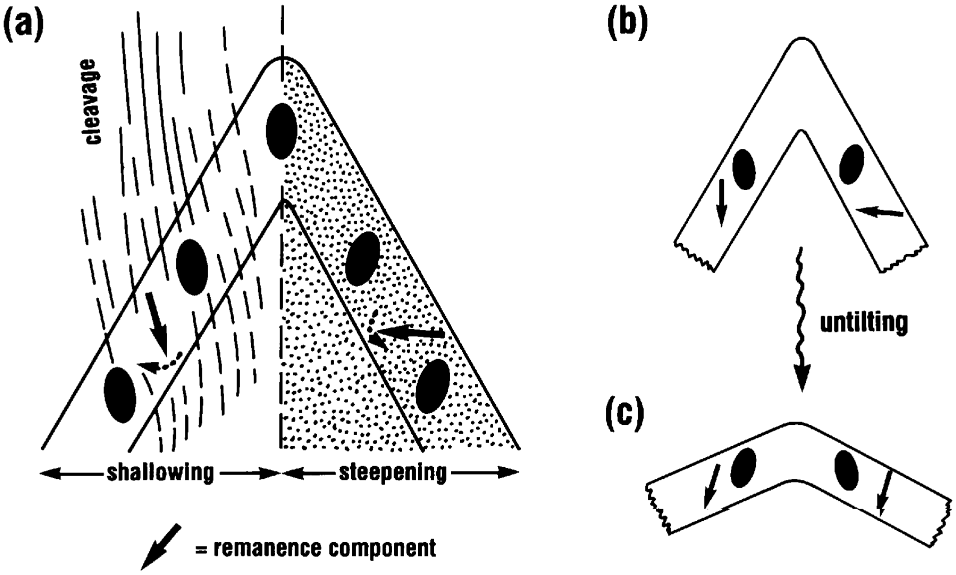

Figure 6. A chevron fold steepening remanence on one limb and shallow it on the other. Such remanences might be restored by successive small strain decrements, alternating with small untilts. Knowing relative tilt-rates and strain-rates may permit a unique restoration.

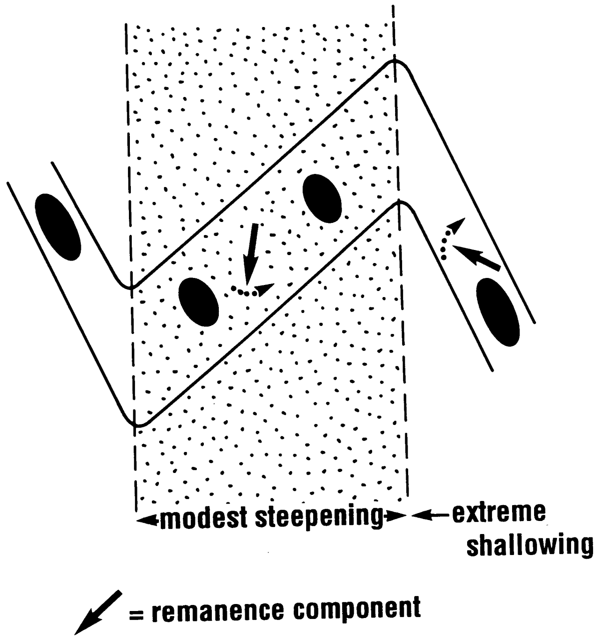

Figure 7. A chevron fold has one limb more strained. Careful selection of synchronous destraining and untiltng increments may reinstate the pretectonic geometry.

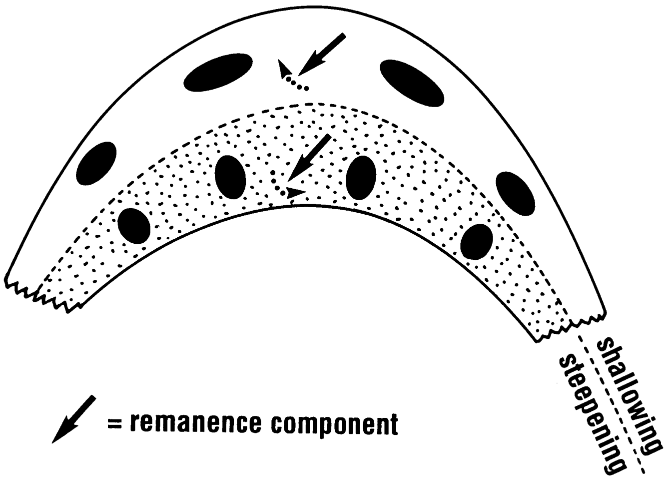

Figure 8. Disharmonic folds reveal amplitude variation along the axial surface. Strain distributions are different on the inner and outer arcs of the fold. The outer arc shows shallowing, the inner arc shows steepening. A correct restoration requires that each site be destrained individually.

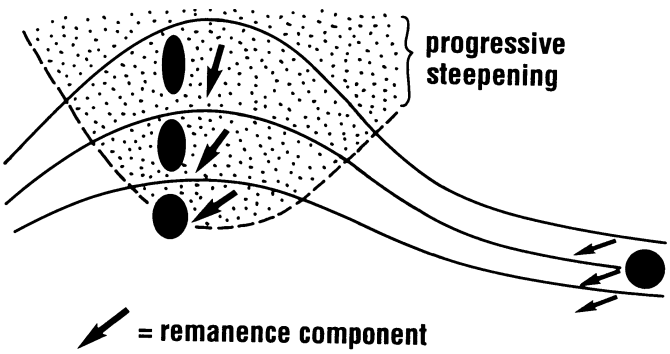

Figure 9. This disharmonic fold has neither limbs nor layers of similar strain. Destraining may restore the limbs to horizontal. This restoration may be paleomagnetically valid if the remanences are then parallel throughout.

in paleomagnetic contexts (e.g., Stamatakos and Kodama, 1991b). Figures 8 and 9 represent disharmonic folds that must be treated in this way. One notes that the zones of steepening or shallowing are no longer confined to separate limbs but are controlled by the strain distribution (see also Kodama, 1988). These illustrate complications produced by heterogeneous strain distributions that are symmetrical with respect to the axial plane. In Figure 7, the effects of different strain magnitude on opposed flanks were noted. In nature, the strain varies also perpendicular to the profile of the fold, along the plunge direction. Heterogeneous strain is invariably three-dimensional.

Where the symmetry of strain varies, so will the trajectory taken by a linear marker. Assuming the remanence component is a passive strain marker and the

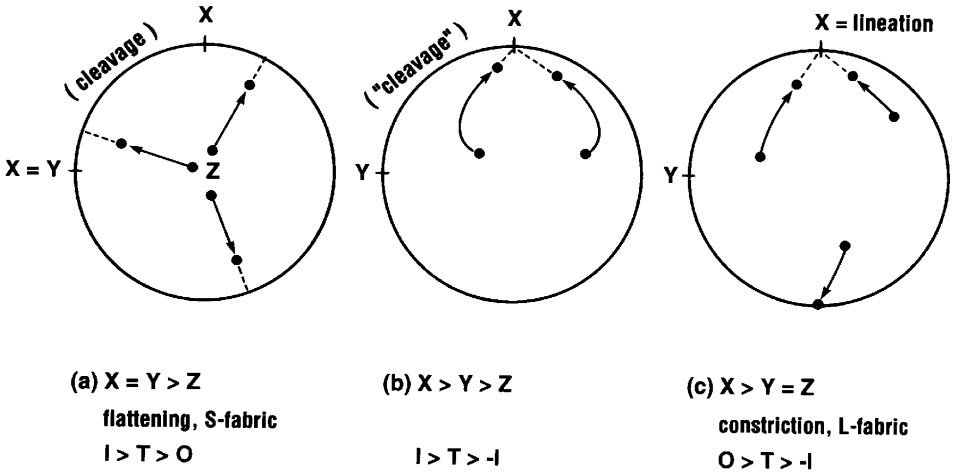

Figure 10. Trajectories of passive linear markers (e.g., remanence under some conditions) during the accumulation of finite strain depends on the shape (symmetry) of the strain ellipsoid. (a) In perfect flattening, the passive vectors move along great circles directly away from Z. (b) In the general case, ($X > Y > Z$) the route will not be a great circle, but veer toward Y on route for X. (c) In constriction ($X > Y = Z$), passive vectors spin directly to X along great circle trajectories. The distance that a vector moves to its destination depends on the strain ratios, X/Y and Y/Z.

principal axes of bulk strain remain fixed with respect to the material (coaxial strain), the remanence vector moves as shown (Figure 10). We most commonly encounter flattening fabrics in mountain belts, producing cleavages if $T > 0$, perhaps with some degree of mineral lineation especially if $T \approx 1$. Constricted fabrics produce well-defined lineations or L-fabrics if $T < 0$, especially if $T \rightarrow 1$. Constriction is common around diapirs and in some mylonite zones, between fault blocks and thrust sheets, and in Archean terrains.

Figure 11 shows a hypothetical paleomagnetic application that involves an antiparallel pair of normal and reversed remanences. On the west flank of the fold, constriction steepens the reversed magnetization direction with respect to bedding. Flattening on the east flank causes shallowing of the normal version of the remanence component. Without knowing the differences in the strain-ellipsoid shape, it would not have been suspected that the dispersed, posttectonic remanences were an originally antiparallel, normal-reversed pair.

In practice, we can avoid some potential problems, by making a few strain observations, without an extensive survey of finite strain. Merely knowing the strain ellipsoid shape (disk, neutral or rod-shaped) and the orientation of its principal axes, at strategic localities, constrains possible restoration sequences and alerts us to problems such as shown in Figure 11. The greatest problem is deciding on the sequence of deformation processes and their degree of overlap (Figure 3).

Figure 12 shows profiles of major folds in western Scotland (Borradaile 1979a,b). Full three-dimensional strain analyses, sometimes using hundreds of strain-markers

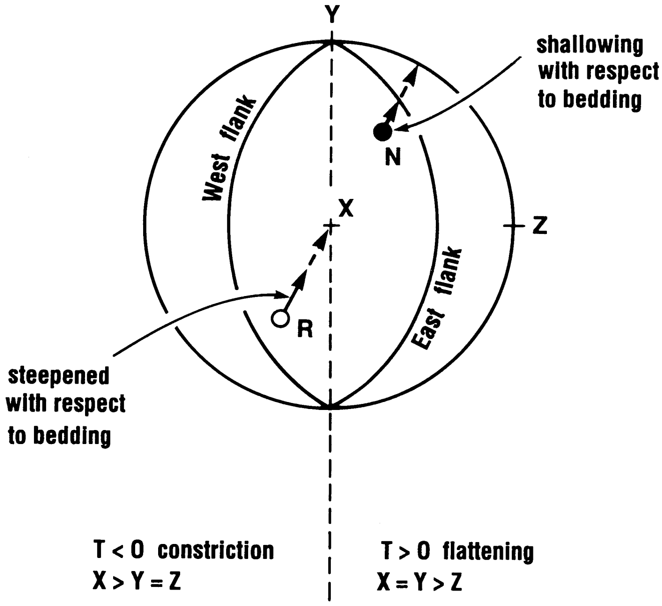

Figure 11. Lower hemisphere stereogram shows a simple fold with a normal remanence component on the east flank and an originally antiparallel reversed component on the west flank. The west flank is constricted, causing the reversed remanence to steepen. On the east flank, flattening makes the normal component shallower. Restoring the antiparallel remanences without first recognizing differences in strain-ellipsoid shape throughout the fold would be extremely difficult.

were made at one hundred and three sites and permitted incremental destraining of the profiles. The strain analyses were derived from the shapes of distorted grains of feldspar and of quartz using Rf/phi and runs methods as well as the analysis of ooids, and strained sedimentary structures. In each case, reversing the strain reduced the amplitude of the folds. In (a) the profile is perpendicular to the XY plane (cleavage). In (b) the profile is perpendicular to YZ, cleavage is very poorly developed and the maximum extension is into the page. This accounts for the apparent inflation of the profile.

Viewing Figure 12, one might assume that since strain has been removed, we may freely untilt the fold limbs to complete a restoration. One might be excused into accepting the destrained fold in (b) as a buckle, but the destrained folds in (b) are clearly not embryonic folds. Destraining is unsuccessful because we have only removed *grain-strain* in most outcrops. Much strain in (a) and most in (b), is taken up by invisible, intergranular motion called particulate flow (Borradaile 1981). Although unrecorded by grain shape, particulate flow permits the rock mass to change shape very effectively. Some paleomagnetic restorations have recognized specific advantages of grain-scale observations, differentiating between grain-scale strain effects on remanence rotation and the effects of rotation on a larger scale (Housen et al., 1993; van der Pluijm, 1987b; and Kodama, 1991).

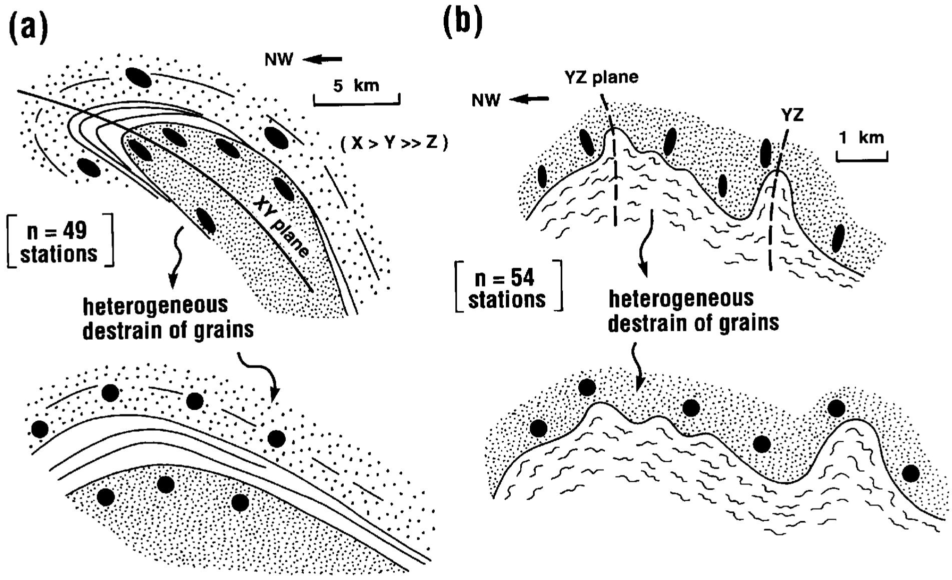

Figure 12. Examples of regional-scale folds that have been destrained mostly using grain-scale markers (Borradaile 1979a,b). (a) shows the Islay nappe after destraining, (b) shows a series of folds of a basement-cover unconformity after destraining. The residual folds indicate that the restorations are incomplete and paleomagnetic interpretations would be premature. Untilting of the remaining gentle folds might improve the situation.

In essence, the preceding discussion recognizes the extra effort needed to apply Graham's fold test to folds with notably strained limbs. However, the arguments are applicable to any tectonic structure used to test the primary character of remanence. We must pay attention to the heterogeneity of strain and to the noncommutative nature of strains and tilts. Further concerns are raised because we do not normally *unfold* rocks properly. We merely untilt their flanks; the overlapping nature of strains and tilting makes it difficult to reverse the natural deformation path. Specific idealized models of fold generation have been used to forward-model the expected behavior of remanence during deformation. However, to assume these models for natural restorations may be no better than to assume untilting alone, as in Graham's original fold test. In reality, there are infinite possible combinations for the strain paths of all parts of a fold. We must use strain determinations from the field in order to destrain and unfold remanence directions successfully. Strain is ubiquitous in folded rocks and can rarely be neglected. Without a successful fold test for primary remanence, we should always be on guard for the possibility of synfolding complications. A further great concern is the implicit assumption that a remanence component is strained passively at all scales from the magnetic domain to the hand-specimen. This simplification is difficult to justify in most cases.

3. The Rotation of a Remanence Component Due to Large Finite Strain

Experimental investigations of experimental deformation on remanence-carrying materials have a long history (among others, Kern, 1961a, b; Lanham and Fuller, 1988; Martin et al., 1978, 1980, 1988; Nagata and Carleton, 1968; Nagata and Kinoshita, 1964; Revol et al., 1977a,b; Pozzi, 1975; Pozzi and Aifa., 1989; Zlotnicki et al., 1981). These studies were directed largely toward using temporal or spatial remanence-variations to predict seismic or volcanic activity or to improve interpretation of geomagnetic anomalies. They did not need to simulate the conditions of ductile tectonic deformation required in this study.

The author's experimental program has addressed these needs to give paleomagnetists and tectonicians relevant remanence-behaviour during *macroscopic ductile* deformation. This is achieved by high confining pressures ($P3 \geq 200$ MPa), low differential macroscopic stresses ($P1 - P3 < 100$ MPa) and low strain-rates (10^{-5} to 10^{-6} s^{-1}) that suppress fracturing of both grains and samples. Because forces are transmitted though a grain-aggregate across very small grain-contacts, the differential stresses imposed on grains may be much higher than the macroscopic differential stress ($P1 - P3$).

Over the last decade, the author has employed computer-controlled, constant strain-rate tests (10^{-5} or 10^{-6} s^{-1}) and constant differential-stress-rate tests to simulate ductile flow at room temperature in experiments lasting from a few hours to a few days (e.g., Borradaile and Mothersill, 1989, 1991; Borradaile, 1991, 1992a,b; 1993a,b; 1994; Borradaile and Jackson, 1993; Jackson et al., 1993). Maximum strains of 45% have been achieved, in some experiments a controlled pore fluid pressure was applied by an external apparatus. However, any similarity between the experimental samples and naturally deformed samples is due to the specially selected materials. Calcite bonded with Portland Cement, limestones, and weathered arkosic sandstones have provided excellent media. The experimental textures are similar to those of natural, weakly deformed rock because of the special, room-temperature, rheological properties. Reproducing the experiments in rocks of greater interest to paleomagnetists, such as basalt, gabbro or granite, is normally impossible without high-temperatures that would destroy the initial remanence due to thermal energy.

In pioneering studies of tectonized rocks, paleomagnetists applied the simplest possible model for the rotation of a remanence component (e.g., Lowrie et al., 1986; Cogné and Perroud, 1985; 1987). This assumed that the remanence component behaved as a passive line with no rheological contrast with the rock. In retrospect, this assumption seems restrictive. The remanence component is a vector sum of spin-moments borne by different grains or subgrains, dispersed through an aggregate with grain-scale principal stresses that may be quite differently oriented from the macroscopic principal stresses ($P1 \geq P2 \geq P3$). Nevertheless, remarkably successful restorations were obtained from field studies of rocks that were strained homogeneously at the outcrop scale.

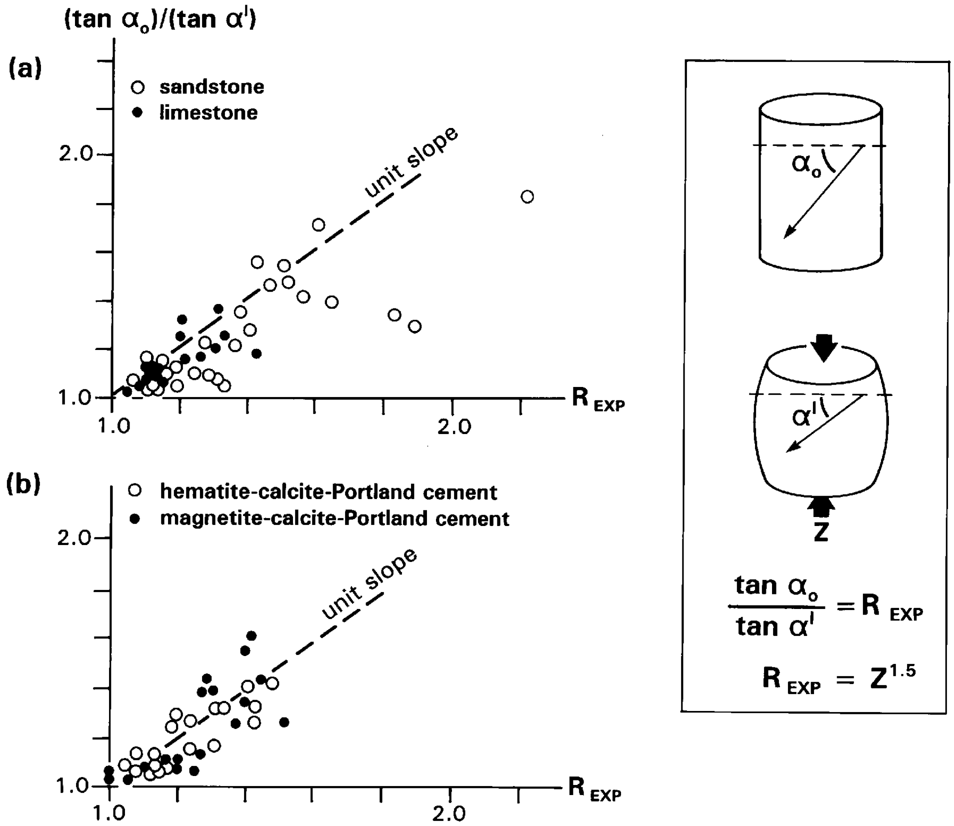

Figure 13. Experimental deformation of (a) natural samples (b) and synthetic samples with an initial single ARM or IRM remanence-component. Initial magnetizations were imposed at varying initial inclinations (α_0). The samples were shortened by amounts varying from 2% to 45% expressed as a strain-ratio R_{EXP}. Changes in inclination agree with those predicted by the equation of passive line rotation.

Now follow some experimental results that confirm this interpretation of strain-rotation of remanence (Borradaile 1993a,b). Figure 13 shows data from experimentally deformed, natural samples (a) and synthetic samples (b). I imposed a near-saturation, single-component, ARM or IRM in some direction and then deformed the sample. The resulting remanences always rotated toward the plane of flattening. The change in inclination from α_0 to α' depends on the strain ratio X/Z where $X = Y > Z$ for flattening deformation. In two dimensions the relationship is given by Ramsay (1967, Equation (3-34)).

$$\frac{\tan\alpha'}{\tan\alpha_0} = \frac{Z}{X}.$$

However, for three dimensions we must carefully evaluate the appropriate strain ratio and rotation path. For example, the experiments shown here all possessed a simple flattening rotation path (Figure 10a) and using the normalization that $X.Y.Z = 1$ the restoration formula would be

$$\frac{\tan \alpha'}{\tan \alpha_0} = Z^{1.5},$$

derived for the special case of uniaxial flattening.

In each field situation, the appropriate strain ratio must be selected to restore α' to α_0. In experiments, at ≈30% shortening a reasonable cleavage appears, but the high strains of most slates (>60% shortening) are unobtainable due to experimental limitations on strain homogeneity. Of course, all samples were incrementally demagnetized to investigate the postexperiment vectorial components of the remanence that are normally significantly changed by deformation.

The equation for rotation of a passive line predicts the expected inclination and this was compared with the postexperiment inclination. The results for both natural samples (Figure 13a) and synthetic ones (Figure 13b) agree with the passive-line model of remanence rotation for strains varying from 2% to 45% shortening and for a wide range of α_0. When high pore fluid pressures are applied to the samples, the microscopic textures may appear more ductile, even at elevated strain rates, due to the enhanced particulate flow and restricted grain-strain. Nevertheless, the passive line model, given by the above equation, is still a reasonable approximation of the experimental changes in inclination (Figure 14).

It is encouraging that such a simple mathematical model successfully restores paleomagnetic characteristic vectors, in homogeneously deformed portions of tectonized rock. However, the passive-line model is clearly not an accurate mechanical description of the complex process by which a remanence component spins away from the direction of shortening. In reality, remanence is carried in grains that spin in a microscopic-scale, heterogenous strain field within the sample. In the case of magnetite, many grains are multidomainal so that deformation can turn the remanence without spinning the grain, by moving domain walls. Even in the most homogeneously strained hand-sample of slate, microscopic observations show marked heterogeneity at the grain-scale. Thus, strain at the grain-scale cannot be neglected in paleomagnetic restorations (Housen et al., 1993; van der Pluijm, 1987b; and Kodama, 1991a). Moreover, as noted earlier, large amounts of strain may be concealed as intergranular displacements, called particulate flow.

Where free grains of magnetite or hematite may carry the remanence, as with clasts of sedimentary or low grade metasedimentary rocks a direct relationship may be expected between their finite strain and the deflection of remanence. However, in high grade metamorphic rocks and some igneous rocks, silicates or other rock-forming grains may include the ferromagnetic carriers (Borradaile 1994a). Clearly, these matrix-grains did not change shape passively. Instead, they rotated as semi-rigid markers and will show poorer agreement with the passive line response model.

Structural geologists are aware that high pore fluid pressures are common due to metamorphic dewatering of underlying rocks and local aquathermal pressuring

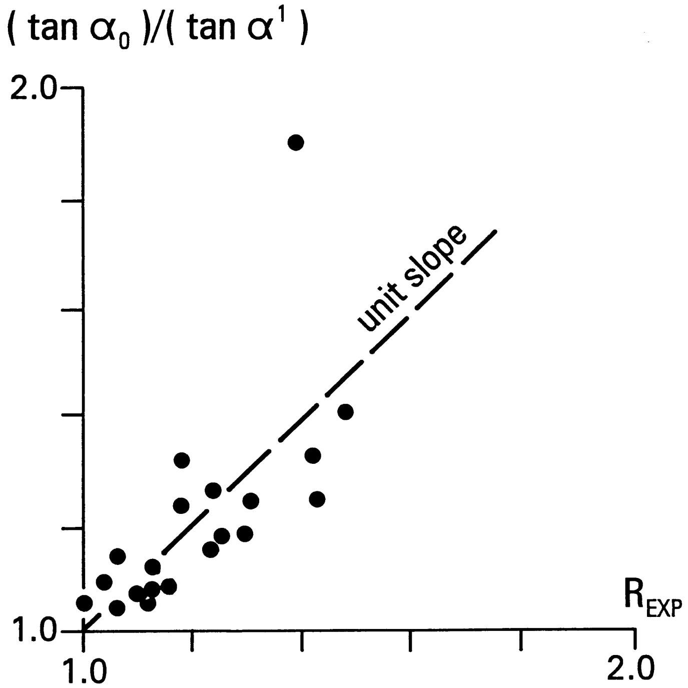

Figure 14. Synthetic samples with a single component remanence, strained experimentally at 10^{-5} s^{-1} with confining pressure $P3$ = 200 MPa and a pore fluid at a pressure $>0.6 \times P3$. This enhanced flow and produced greater bulk strains more easily. However, the final inclinations of remanence (α'_{EXP}) still approximate those predicted for the rotation of a passive line ($\alpha'_{\text{passive line}}$), as in the dry tests of Figure 13.

(Norris and Henley, 1974; Fyfe et al., 1978). This causes ephemeral dissaggregation and particulate flow of rocks (Borradaile, 1981). Commonly observed ductile folds in non-metamorphosed, sedimentary sequences show that this is a common phenomenon. Similarly, local ductile shear zones in otherwise undeformed granites indicate the same process. The aspect ratio of these grains is the chief factor controlling their rate of rotation and the rotation of their remanences (Borradaile 1993a, b). Clastic magnetite grains normally have low aspect ratios, e.g., 1.2, and we would not expect the remanences they carry to conform to the passive-line model (Figure 15). However, most remanence-bearing grains are inclusions in silicates that have high aspect ratios, e.g., 2 to 5. The rotation of such rigid grains approximates the rate of rotation of the passive-line. In part, this may explain the success of the passive line model for strain-rotation of remanence in case studies (e.g., Cogné 1987b, 1988, 1991; Cogné and Canot-Laurent, 1992; Cogné and Gapais, 1986; Cogné and Perroud, 1985, 1987; Cogné et al., 1986; Hirt et al., 1986; Kligfield et al., 1983; Lowrie et al., 1986). Noncoaxial strain histories present complications that are surmountable in theory, or with sufficient field data. However, if significant metamorphic recrystallization accompanies strain, the remanence dispersal processes

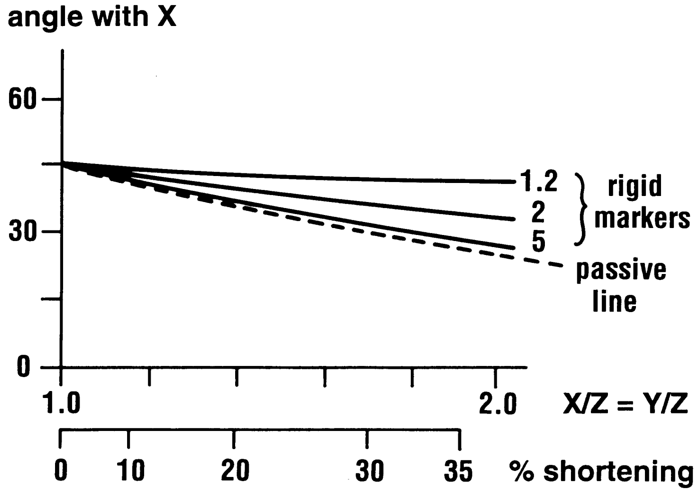

Figure 15. Finite strain rotates a linear marker toward the flattening direction (for $X = Y > Z$) according to the properties of the marker. Passive lines spin more quickly than rigid markers. However, rigid markers of high aspect ratio (e.g., 5 : 1) rotate almost as quickly as passive lines.

are quite different, perhaps related to a momentary, syncrystallization stress field and unrelated to finite strain directions (Werner and Borradaile, 1996). In this case, neither a fold test, nor a deformation test based on another heterogeneously strained structure, is meaningful.

4. Effects of Finite Strain on Hysteresis Properties

Triaxial experiments ($P1 - P3 < 100$ MPa, $P3$ = 200 MPa) and hydrostatic experiments (P1 = P3 = 200MPa) consistently reveal that even small differential stress (>25 Mpa) for short periods (2 hours) changes the fundamental magnetic properties of magnetite and hematite (Borradaile 1991, 1992a,b; 1994a,b; 1996). Even without macroscopic differential stress, as in hydrostatic tests, large differential stresses exist at the grain-scale due to the material's noncontinuum nature. Differential stress on magnetic grains enlarge the area of the hysteresis loop, increasing both coercivity (H_C) and the capacity to carry a zero-field remanence (M_R). Thus, strain improves the sample's ability to preserve a paleomagnetic record as shown by the magnetite-bearing sample in Figure 16. Coercivity spectra reveal the change in distribution of coercivities due to penetrative strain (Figure 17). Grain-structures of higher coercivity replace those of lower coercivity. For magnetite-bearing samples, the threshold is close to 20 mT; coercivities below 20 mT largely disappear at the expense of higher-coercivity material (Figure 17). It is suggested that crystal damage accumulates due to differential, grain-scale stress and this impedes subsequent domain wall motions, thus causing "magnetic hardening" (Jackson et al., 1993). Theoretical and experimental studies show that the concentration of crystal

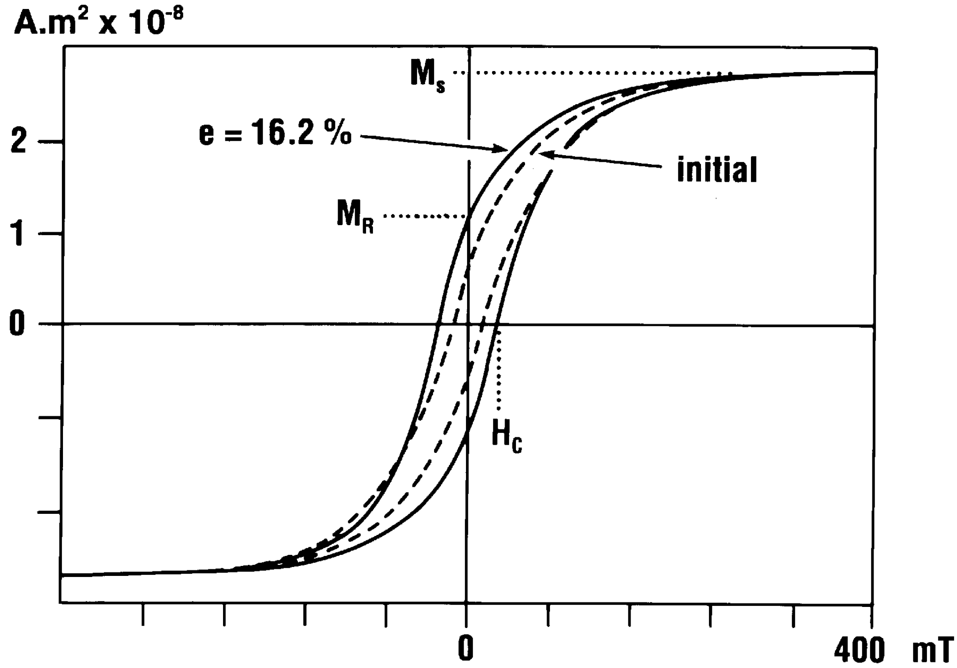

Figure 16. The hysteresis loops show the remanence properties of a magnetite-bearing sample before and after 16.2% shortening in coaxial flattening at 10^{-5} s^{-1}. Strain causes magnetic hardening rendering hysteresis properties comparable to that of smaller grains. However, grain-size is not reduced. It is inferred that intracrystalline damage causes the changes in coercivity (H_C) and coercivity of remanence (H_{CR}). M_S is the saturation remanence.

dislocations and intracrystalline stress affect fundamental remanence characteristic properties such as coercivity and coercivity of remanence (e.g., Carmichael, 1968a,b; Borradaile and Jackson, 1993; Graham et al., 1957; Hodych, 1990; Shive and Butler, 1969; Soffel, 1966; Xu and Merrill, 1989; 1992; Yun and Merrill, 1995). Preliminary electron microscope observations seem to show that this is not due to a mechanical reduction in magnetite grain-size. Grain-size remains constant in ductile limestone or cement samples but their hysteresis properties change toward those expected of smaller magnetite grains during experimental deformation (Borradaile, 1991; Jackson et al, 1993). Those workers refer to this phenomenon as *magnetic hardening*. Only in less ductile materials, such as sandstone with coarse magnetite grains, do we see a reduction in grain size accompanying magnetic hardening (Borradaile and Mothersill, 1991).

The history of crystal damage that grains have inherited, strongly affects their response to subsequent deformation events. Two different types of magnetite were added to calcite aggregates. One magnetite sample had been crushed in the laboratory, with uniaxial stresses of at least 100 MPa. The other sample of magnetite was chemically precipitated and is essentially unstressed by comparison. Deformation

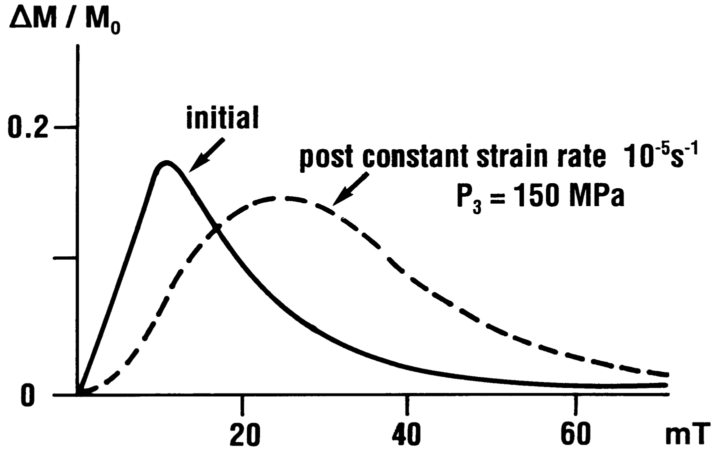

Figure 17. Experimental deformation of magnetite changes its coercivity-distribution. The peak or mode of the postdeformation distribution shifts to higher coercivities when higher differential stresses $(P1 - P3)$ or higher strain-rates are applied. Significant coercivity-redistribution requires a threshold value of $(P1 - P3) \geq 25$ MPa. Naturally, the grain-scale stresses must be much higher than the macroscopic stresses $(P1, P2 = P3)$ applied to the sample.

experiments applied macroscopic hydrostatic pressure of 200 MPa to the samples. However, the heterogeneous grain structure and the amplification of stress at small grain contacts ensure the presence of large *differential* stresses at the grain-scale.

The coercivity spectrum of the prestressed magnetite was almost unchanged by compaction (Figure 18a), because the microscopic stresses of macroscopic hydrostatic compaction could not reset the dislocation patterns set in the magnetite by its initial crushing-preparation. Its ability to carry or preserve remanence is unaltered by experimental deformation. In contrast, the chemically precipitated magnetite was susceptible to crystal damage during experimental deformation. Macroscopic hydrostatic compaction shifted its coercivity spectrum, improving its ability to act as a magnetic recorder (Figure 18b).

Consider the precipitated magnetite to be analogous to diagenetic magnetite in nature, and the crushed magnetite analogous to stressed metamorphic magnetite. Thus, one might infer that the qualities of diagenetic magnetite as a paleomagnetic recorder might improve if subjected to some compaction or deformation. Conversely, strained metamorphic magnetite, work-hardened with high dislocation densities, carries remanences that subsequent deformation does not modify or erase easily (Borradaile and Jackson, 1993).

5. Coercivity Selection Rotates Multicomponent Remanence

The effects of *finite* strain on a single component of remanence were discussed above. This may be the primary component needed for paleogeographic recon-

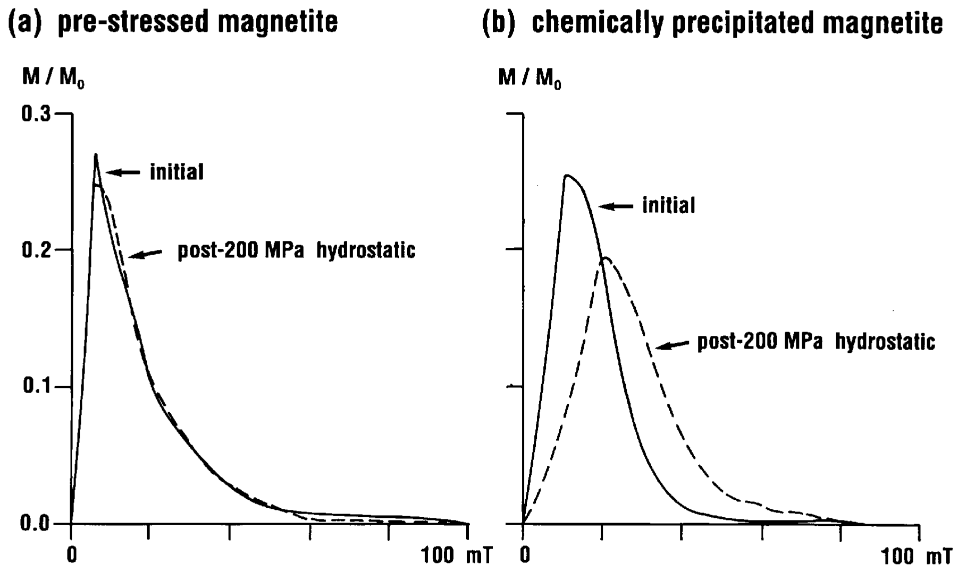

Figure 18. Experimental hydrostatic compaction ($P1 = P2 = P3 = 200$ MPa) of samples containing (a) *prestressed* magnetite, and (b) low-stress, chemically precipitated magnetite. Nevertheless, at the grain scale, significant differential microscopic stresses are recognized due to the noncontinuum nature of the aggregate. (a) Compaction of prestressed magnetite produces negligible redistribution of coercivities because the experiment cannot overcome or change the existing dislocation tangles that fix domain walls. (b) Initially stress-free magnetite accumulates much intracrystalline damage under the same conditions changing its coercivity-distribution.

struction. Some secondary component defined as a separate vector by bounding coercivities or unblocking temperatures may be equally valuable in mobile paleogeography. Below, we consider the effects of experimental deformation on components of different coercivity that we may compare with NRM components of different age in routine paleomagnetic studies.

In all paleomagnetic studies we isolate the individual vector-components that comprise the NRM, using alternating field or thermal demagnetization techniques. The effects of stress or finite strain on each individual vector component could be considered, as in the above text. Nevertheless, let us consider the effects of experimental deformation on the *vector sum*, which is analogous to the initial NRM that would be measured by a paleomagnetist before any laboratory demagnetization. We will see that vector components of different coercivity respond differently to strain, causing significant changes in the direction of the vector sum. In this examination, we consider only *small strains* that produce $<15\%$ shortening in perfect flattening experiments ($X = Y > Z$). Thus, the rotations of individual remanence components are negligible, in contrast to the previous section on large finite strain.

In early experiments, sandstone core samples, bearing an NRM, were shortened perpendicular to bedding in triaxial experiments. Replicate tests confirmed that

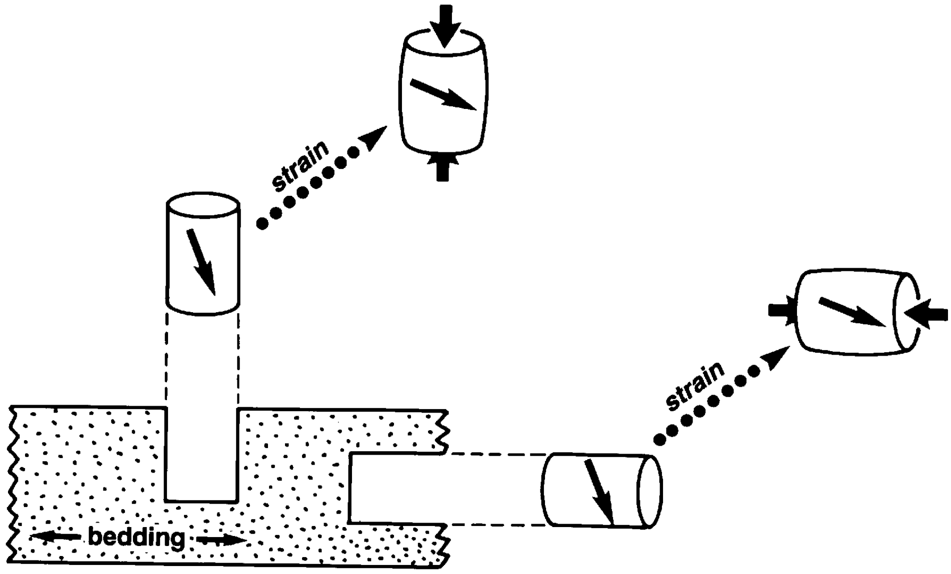

Figure 19. Deformation experiment of multicomponent remanence clarifies the relative behavior of different remanence components. Ductile compression of a sandstone perpendicular to bedding, spins the net remanence away from the shortening direction, as with a single remanence component (e.g., Figure 13). However, when the rock samples are shortened parallel to bedding the vector sum remanence turns toward the shortening direction. This is due to deformation selectively removing certain coercivity components that lie perpendicular to bedding, *regardless of the direction of shortening.*

remanence rotates away from the shortening axis. However, in this sandstone, for cores parallel to bedding, the net NRM remanence rotates in the wrong sense (Figure 19), incompatible with the rotation expected from finite strain. This demonstrated that the composition of remanence components overwhelmed the effects of strain in determining the final remanences (Borradaile and Mothersill, 1989; 1991). Of course, the reader realizes that strain was selectively removing a soft component; the orientation with respect to the strain ellipsoid was of no consequence (Borradaile 1991, 1992a,b). On first inspection this may seem counterintuitive to structural geologists where the angular relations between the strained element and the strain axes are paramount.

Experimental campaigns on magnetite-bearing limestones (Borradaile, 1992b) and calcite aggregates containing synthetic hematite (Borradaile 1992a) reveal how coercivity-distribution plays a more significant role than strain directions in rotation of the net remanence. To each sample, two orthogonal remanence components were applied as an ARM or IRM. A large, near-saturation remanence was applied in one direction. The sample was then AF cleaned along three axes to 20 mT, for example. A 20 mT magnetization was then applied perpendicular to the first. This produced two clearly distinguished remanence components from 0–20 mT, and from 20 mT to at least 200 mT. Naturally, the orthogonal components were not always parallel and perpendicular to the core axis but only these cases are illustrated, for simplicity (Figure 20). Where the initial artificial remanence was an IRM, an induction of 1.0 T was applied. For an ARM a 0.1 mT bias direct field applied the remanence over an AF demagnetization window from 200 mT down to zero. Figure 20 shows experiments in which samples were shortened parallel or perpendicular to the hard component of remanence. Of course, other angular arrangements were also investigated.

The small experimental straining of magnetite-bearing samples shows a distinct reduction in the magnitudes of both vector components, but especially the soft one ($<$20 mT). Consequently, the apparent rotation (ω) of net remanence is always toward the original direction of the harder component, *despite the direction of shortening* (Figure 20a,b). This is because the softer components of remanence are carried by larger magnetite grains whose magnetic moments are more easily dispersed by small strains or internal domain wall rearrangements.

Where small experimental strains are applied to calcite-cement aggregates with fine-grained hematite, the *net* magnetic vector rotates in the opposite sense, toward the soft component. The hard component of remanence is reduced in intensity more than the soft component. Thus, the net remanence spins toward the shortening axis (Figure 20c,d). The selective reduction of the hard component of remanence is due to the very fine-grained hematite being easily disturbed in the Portland cement matrix. Thus, their moments are scattered, their vector sum is significantly reduced and the magnetic vector for the sample turns toward the shortening axis (Borradaile 1992b), contrary to the expectations of elementary structural geology. However, I do not suggest that $+\omega$ rotations characterize strained magnetite and $-\omega$ rotations characterize strained hematite in general. Instead, the sense of rotation of the net remanence depends on the orientation of the vector components of different coercivity (Figure 20), the grain-size distribution and the microstructural behavior of the magnetic grains during deformation of the particular material.

Thus, strain selects which components may survive and which may be eliminated, based on the coercivity and microstructural behavior of the magnetic grains and their host or matrix. Although based on laboratory experiments, this knowledge may be useful when interpreting the multicomponent magnetizations in tectonically deformed terrains. Local, heterogeneous strain, for example on specific fold limbs, could selectively eliminate soft secondary components, or alternatively, hard components. The survival of the characteristic or primary component depends on how much strain occurred and the strain response mechanism for the magnetite-bearing grains. Recall that in deformation of remanence, the strain response model includes both rock-magnetic and microstructural criteria.

The examples shown produce dramatic changes in intensities with $<$15% shortening. Incipient slaty cleavage requires 30% shortening and most slates or schists have $>$60% shortening. The reader may at this point consider these comments merely academic, or unnecessarily alarming. However, we have yet to consider the combination of the destructive effects of small strains with the constructive effects

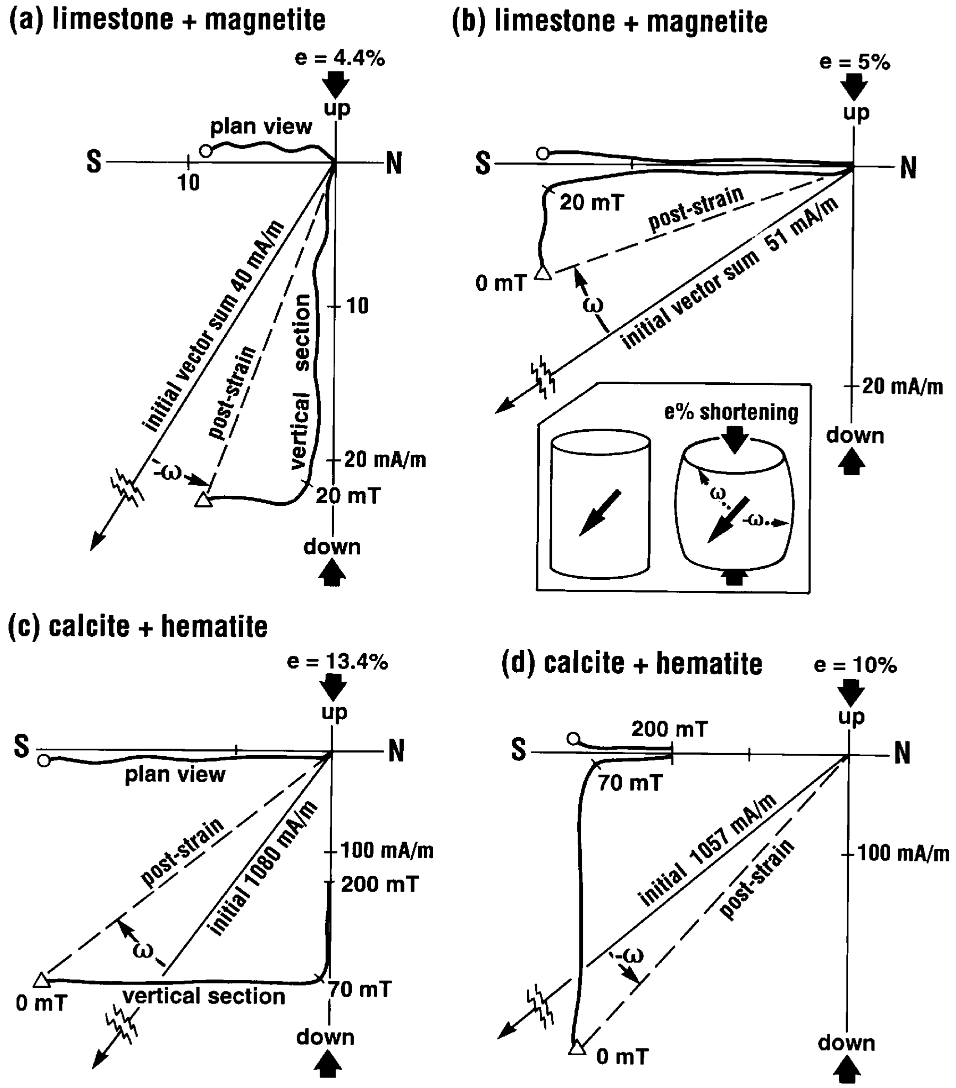

Figure 20. (a, b) Magnetite-bearing samples experimentally deformed at $P3 = 150$ or 200 MPa, $10^{-5}\ s^{-1}$ and small strains ($<15\%$ shortening) show selective removal of low coercivity components, probably by domain wall movement. (c, d) samples with fine hematite in the cement matrix, high coercivity components are selectively removed because the very fine "magnetically hard grains" are more easily disoriented by deformation.

of stress accompanying magnetization, namely tectonic remagnetization, discussed in the following section.

6. Stress-Remagnetization with Small Finite Strains

Initially, from careful applications of Graham's fold test, paleomagnetists recognized remanence components that could only be syntectonic. Normally, the field evidence is convincing. However, the following summary of an experimental program shows the geometric simplicity of a process that can cause syn-stress remagnetization in homogeneously strained samples bearing pseudo-single domain magnetite (Borradaile, 1994; 1996). This may explain why syntectonic remagnetization, such as that documented by Hudson et al. (1989), could be common in nature. Of course, most remagnetization is not directly attributed to tectonic stress but to fluid, chemical, or viscous effects (e.g., Dunlop, 1989; Elmore and McCabe, 1991). Moreover, where chemical or fluids play a role, the phenomenon is not strictly "remagnetization" because new minerals grow whose chemical remanent magnetization swamps that of magnetically softer, earlier grains.

One of the triaxial rigs in the author's laboratory has an internal field of 30 μT that has been constant in intensity and direction for a decade. Through trial-and-error, a synthetic calcite magnetite medium was created that could be remagnetized in the pressure vessel when the sample was subject to hydrostatic compaction ($P1 = P2 = P3$) or triaxial differential stress ($P1 > P2 = P3$). Figure 21a shows samples, all shortened in the Z-direction by 15%, with a differential stress of 150 MPa for 2 hours. The samples each had an initial, single-component ARM, applied in a different direction in each sample. After strain, AF demagnetization reveals that soft components were added during deformation parallel to the field in the pressure vessel. Thus, the progressive demagnetization components lie along a great-circle path from the initial direction toward the pressure-vessel field. Note that the orientations of the strain axes (X, Y, Z) are irrelevant. Only the initial remanence and the remagnetizing field determine the remagnetization path. This seems paradoxical after the early discussion of the complications of strain and the fold test. However, we must remember that the sample-interior does not experience the uniformly oriented macroscopic stresses, $P1$, $P2$ and $P3$, applied outside the specimen. Grain-scale, stress-heterogeneity provides *any* required local stress-orientation to be available to act on some ferromagnetic grain, permitting its domain-configuration to shift and trap the ambient, "remagnetizing" field.

A Zijderveld (1967) graph reveals the demagnetization of one sample that had an initial remanence antiparallel to the remagnetizing field (Figure 21b). Stress-induced remagnetization clearly flips all components with coercivities $<$20 mT into the direction of the syn-strain, almost antiparallel magnetic field. Detailed studies show that the *larger* the differential stress ($P1 - P3$) the *greater* the range of coercivities that switch to the remagnetizing field direction. Moreover, *sample-scale*, differential stress must exceed a 20 MPa threshold to produce any remagnetization. Remarkably, remagnetization occurs in two hours with macroscopic differential stresses that are comparable to those in nature and negligible strains.

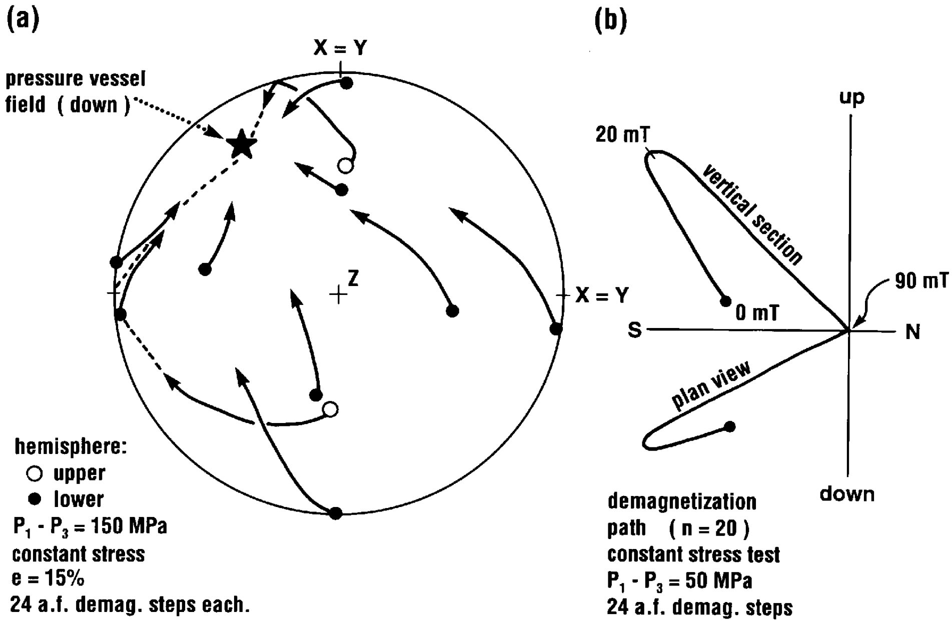

Figure 21. (a) Experimental remagnetization of magnetite-bearing samples with differently directed, univectorial ARM, during an experimental shortening of the Z-direction by 15%, under constant macroscopic differential stress (≈ "creep test"). After deformation, AF demagnetization reveals that successive vector components now lie on a great circle from the initial remanence toward the direction of the weak field in the pressure vessel. Only the initial remanence direction and the remagnetizing field dictate the remagnetization geometry. (b) A Zijderveld vector plot illustrates the effects of constant differential stress on an initial single component remanence that was almost antiparallel to the remagnetizing field in the pressure vessel. The initial remanence extends from the origin outward. Soft components below 20 mT have been erased and replaced with antiparallel components in the direction of the pressure vessel field. At this differential stress, 20 mT represents the upper coercivity limit of remagnetization. This limit increases with higher differential stress.

At this point we may take comfort that experimental stress-remagnetization affects only soft remanence components, which are normally secondary in nature and therefore of less value to paleomagnetists. The primary components, normally recognized by their high coercivity in field studies, should be undisturbed. However, some experiments showed remagnetization of hard components (>60 mT) also, that could be confused with primary magnetizations (Figure 22), in a natural study. Of course, I took care that spurious ARM and GRM were not produced at the high AF demagnetization in this investigation (Stephenson, 1980; Stephenson and Molyneux, 1987).

7. Conclusions

Any tectonic structure formed by bulk, heterogeneous strain can test the primary nature of remanence if we know how to restore the structure to its initial state.

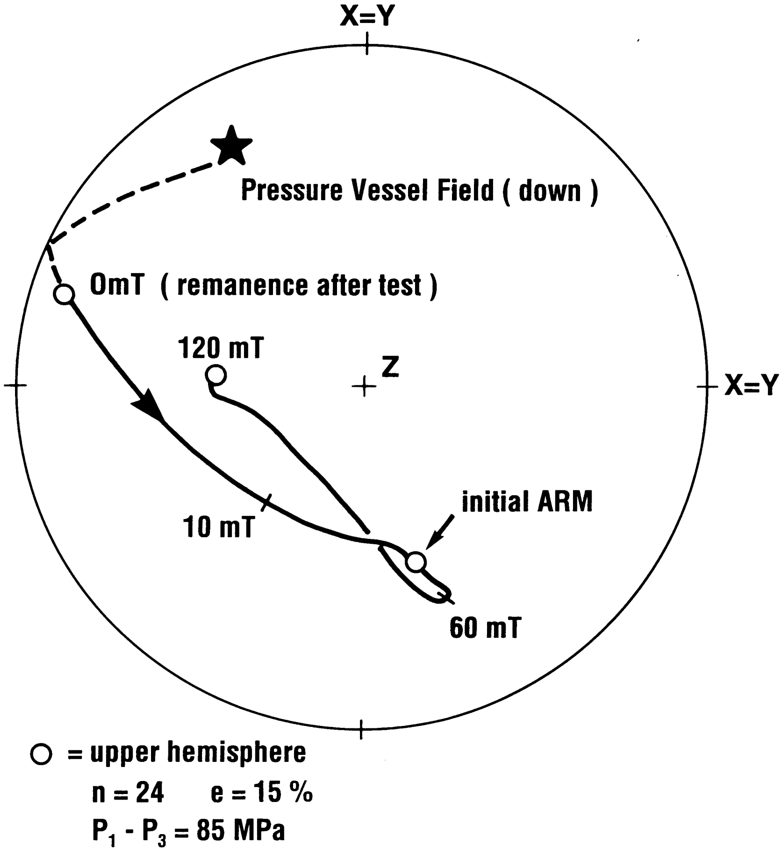

Figure 22. Remagnetization during experimental deformation may add both high and low coercivity components. Here, an initial upwards ARM is modified by stress over two hours. AF demagnetization reveals soft remanence vector components smeared along the great circle toward the pressure vessel field direction. However, some high coercivity components (60–120 mT) are also added by remagnetization.

Ideally, we must know the finite strains through the structure, the sequence of successive tilts and strains and their spatial and temporal overlap. Underestimating the complexity required to restore a heterogeneous structure and its remanence could lead to inaccurate claims of syntectonic magnetization or of multiple, early, significant remanence components.

The author's experimental program of the differing effects of stress, and of strain on remanence can only be extended to natural situations with caution. Although materials were chosen to simulate natural ductility under room temperature experimental conditions, the computer-controlled strain rates were about one million times faster than those in nature. The finite strains were at the lower part of the range found in tectonically deformed sediments and slates, and the differential stresses were similar to the long term values found in nature.

Where finite strain effects dominate, a single component of hard remanence rotates quite simply, due to homogeneous strain. The actual mechanisms of rotation of remanence are complex, due to heterogeneous grain-scale rotation and the nonpassive behavior of the mineral grains. Nevertheless, these experiments and other field studies reveal that the natural macroscopic response closely approximates the rotation of a passive line. This permits a simple restoration of a remanence component from the field, knowing the state of homogeneous strain. Such reconstructions are normally invalid if metamorphic recrystallization accompanied or succeeded strain.

During the accumulation of finite strain the hysteresis properties of the magnetic carriers change. Because of intracrystalline damage, dislocation densities increase, impeding domain-wall mobility and therefore improving the rock's ability to remember a paleofield. Thus, for similar magnetic grain sizes, diagenetically precipitated magnetite may be a poor magnetic recorder whereas tectonically stressed magnetite of metamorphic origin may be superior.

Even with small finite strains, short pulses of differential stress at high pressure ($P3$ = 200 MPa) can remagnetize pseudo-single domain magnetite in laboratory experiments. It occurs in a few hours if a threshold macroscopic differential stress is exceeded ($P1 - P3 > 20$ MPa). The new remanence directions are close to the remagnetizing field direction and accumulate at the expense of initial low coercivity components ($<$20 mT). However, in some cases, high coercivity components are added also parallel to the ambient field. AF demagnetization shows that vector components smear along a great circle between the initial remanence direction and the direction of the remagnetizing field. The actual directions of stress or strain are irrelevant in these stress-remagnetization experiments because the microscopically heterogeneous stress fields in the sample can present any required stress orientation to any grain, triggering its remagnetization.

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