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A comparison of quartz *c*-axis preferred orientations in experimentally deformed aplites and quartzites

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Abstract—The effect of feldspar on the development of quartz *c*-axis preferred orientations was investigated by experimentally deforming aplite (30% quartz) and pure quartzite under equivalent conditions and comparing the quartz *c*-axis patterns. Axial compression experiments were performed at temperatures of 800 and 900°C, strain rates of 10^{-6} and $10^{-5}/s^{-1}$, and a confining pressure of 1500 MPa. Under these conditions, both quartz and feldspar deform by dislocation creep with only minor syntectonic recrystallization. The quartz *c*-axis preferred orientations produced in the two rock types at equivalent deformation conditions and quartz grain strains were found to be essentially identical in pattern and strength. These results indicate that the presence of feldspar has little or no effect on the operative slip systems in adjacent quartz grains in quartzo-feldspathic rocks, and that quartz *c*-axis preferred orientation analyses can be performed on a wide range of crustal rock types.

INTRODUCTION

QUARTZ *c*-axis preferred orientations in naturally deformed rocks are commonly used as an indicator of the orientation and relative magnitude of finite strain. A combination of theoretical work (e.g. Etchecopar 1977, Lister *et al.* 1978), experimental work (e.g. Green *et al.* 1970, Tullis *et al.* 1973, Tullis 1977), and detailed observations of naturally deformed rocks (e.g. Schmid & Casey, in press) has shown that the resulting *c*-axis preferred orientation depends on the active slip systems, the type and magnitude of the finite strain, and the strain path. Our present understanding of *c*-axis orientations in quartz is largely based on studies of pure quartzites (e.g. Green *et al.* 1970, Tullis *et al.* 1973, Tullis 1977). An important remaining question is whether the presence of other mineral phases can influence the development of a quartz *c*-axis preferred orientation and cause a change in either the pattern or the intensity of the preferred orientation. Other phases might exert an influence by: (1) causing a change in the active slip systems in the quartz due to different grain boundary constraints, (2) causing at least local changes in the type and magnitude of finite strain resulting from slip in the quartz, due to the second phase being much stronger or weaker than the quartz or (3) favoring grain boundary sliding which would rotate the quartz grains randomly.

Although there appears to be a general expectation that the presence of other phases will inhibit the development of a quartz preferred orientation (e.g. Hobbs *et al.* 1976, Malavieille & Etchecopar 1981), few studies have sought to test this hypothesis and the accumulated evidence appears somewhat contradictory. In their study of naturally deformed rocks, Starkey & Cutforth (1978) measured quartz *c*-axes in 18 different samples of varying lithology from two different localities. They found a linear correlation between the strength of the quartz *c*-axis preferred orientation and the modal abundance of quartz in the rock. They concluded the most probable explanation for this relation to be that increased amounts of matrix allow a switch in deformation mechanism to favor grain boundary sliding. However, they do not report any information on the amount or type of finite strain in the rocks, the quartz grains, or the quartz regions, nor do they have any evidence that grain boundary sliding has in fact occurred. Thus the significance of the reported correlation remains unclear.

Two studies have been done on the effect of mica on quartz *c*-axis preferred orientations, but they appear to give conflicting results. In a naturally deformed quartz pebble conglomerate, Kronenberg (1981) found a strong *c*-axis preferred orientation in the pure, coarsely recrystallized quartz cobbles but a random distribution of *c*-axes in the fine, recrystallized quartz–mica matrix. He postulated that the mica (which itself has a strong preferred orientation) allows for quartz grain rotations, thereby randomizing the orientation of the *c*-axes. In contrast, White *et al.* (1982) found no difference in the *c*-axis orientations of recrystallized quartz grains between pure quartz and quartz–mica layers of a quartz-rich rock associated with the Moine thrust. Unfortunately, the mechanism of grain boundary sliding in these rocks cannot be proven or disproven, and the intragranular strain accomplished by slip within the quartz grains is also unknown because they are completely recrystallized. Thus the reason for this apparent contradiction remains unresolved.

Many workers have found strong preferred orientations of quartz *c*-axes in naturally deformed quartzo-feldspathic rocks (e.g. Simpson 1980, White *et al.* 1982, Jenson 1984, O'Hara & Gromet 1985), but only two previous studies have investigated the effect of feldspar on the development of quartz *c*-axis preferred orientations. In a quartzo-feldspathic mylonite, Lister & Price (1978) demonstrated that rigid feldspar porphyroclasts cause inhomogeneous deformation in adjacent quartz grains, producing variations in deformation paths of the quartz grains and corresponding variations in their preferred orientations. In a similar rock, Price (1978) demonstrated that the (recrystallized) quartz grains in zones of higher feldspar content have a different texture and also a somewhat different pattern of preferred orientation than do those in pure quartz regions. The reason for this difference was not determined.

In order to address the question of how other phases affect the development of quartz *c*-axis preferred orientations, we have done comparison deformation experiments on a pure quartzite and an aplite containing 30% quartz and 67% feldspar. Previous work has shown that experimentally and naturally deformed quartzites with the same finite strain and optical microstructures (flattened original grains with little or no recrystallization) have similar preferred orientations, indicating that the same slip systems have been operative (e.g. Tullis 1977). In the present study, experiments were performed at conditions where both phases deform by dislocation creep, developing flattened original grains with only minor syntectonic recrystallization. We have measured the quartz *c*-axis preferred orientations produced under equivalent deformation conditions in the quartzite and aplite samples for the case of axial compression, and find that they are essentially identical.

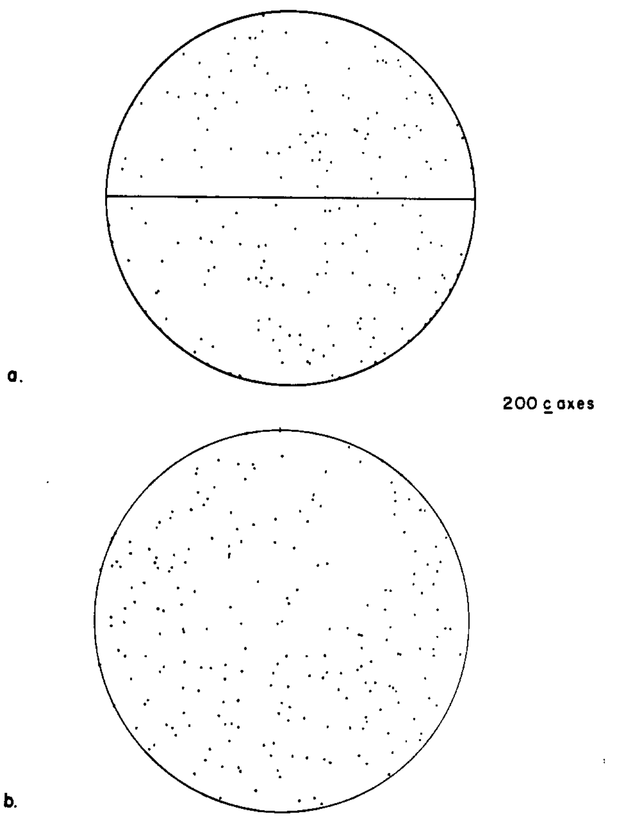

Fig. 2. Equal area, lower hemisphere stereographic projections of 200 quartz *c*-axes for undeformed Enfield aplite (a) and Heavitree quartzite (b). The horizontal line in (a) marks the foliation plane defined by biotite grains.

EXPERIMENTAL PROCEDURES

Starting materials

The aplite used in these experiments was collected from dikes in the core granite of the Mascoma dome at the abandoned Moose Mountain quarry near the town of Enfield, New Hampshire (Stop No. 3, Naylor 1971). We have called the material Enfield aplite. It contains approximately 30% quartz, 40% microcline (Or_{94}), 27% oligoclase ($Ab_{83}An_{11}Or_1$) and 3% biotite and magnetite. It has a granular texture with a grain size of 150 ± 50 μm and consists of a continuous matrix of feldspar with isolated grains of quartz and mica (Fig. 1a). There is a faint foliation due to alignment of the mica and the slightly inequant quartz and feldspar grains (aspect ratios of 0.9 and 0.8, respectively). All samples were cored perpendicular to the foliation. The quartz exhibits no initial preferred orientation (Fig. 2a).

Heavitree quartzite (from central Australia, kindly donated by M. Paterson) was used for the comparison experiments. It has relatively equant grains with an average diameter of 200 ± 50 μm (Kronenberg & Tullis 1984), and contains <1% impurities, which are mostly opaque phases and muscovite (Fig. 1b). The quartzite also exhibits no initial preferred orientation (Fig. 2b).

Experimental conditions

Sample cores (6.35 mm diameter and 14–15 mm long) were dried in an oven at 90–100°C to remove excess water acquired during preparation, and then were mechanically sealed in 0.10 mm thick jackets of Ag or Pt (inner layer) and Ni (outer layer). Axial compression, constant displacement rate experiments were performed in a Griggs-type apparatus using NaCl as the confining medium. The use of stepped furnaces and zirconia pistons has greatly decreased temperature gradients in the samples and the deformation is extremely uniform. Details of the sample assemblies can be found in Jaoul *et al.* (1984). Experiments were done at a confining pressure of 1500 MPa, constant strain rates of 10^{-5} and $10^{-6}\ s^{-1}$, and temperatures of 800 and 900°C. These conditions are within the alpha quartz field (Cohen & Klement 1967), and within the dislocation creep field for both quartz and feldspar (Tullis & Yund 1977, Dell'Angelo & Tullis 1982 and in preparation). Total sample strains varied from 20 to 60%. The experiments are summarized in Table 1.

Table 1. Summary of experimental data

Material	Confining pressure	$\dot{\varepsilon}$	T(°C)	Flow σ(MPa)	Total sample ε	Quartz grain ε	Feldspar grain ε	Run no.
Aplite	1500 MPa	$10^{-6}s^{-1}$	800	1000	20%	23.9	24.3–29.5	W-305
Aplite	1500 MPa	$10^{-6}s^{-1}$	800	950	58%	58.6	51.5–54.9	W-316
Aplite	1500 MPa	$10^{-6}s^{-1}$	900	380	22%	25.9	19.3–25	W-154
Aplite	1500 MPa	$10^{-6}s^{-1}$	900	450	60%	64.3	50.1–55.1	W-304
Quartzite	1500 MPa	$10^{-5}s^{-1}$	900	600	60%	—	—	QM-41
Quartzite	1500 MPa	$10^{-6}s^{-1}$	800	350	58%	—	—	W-370

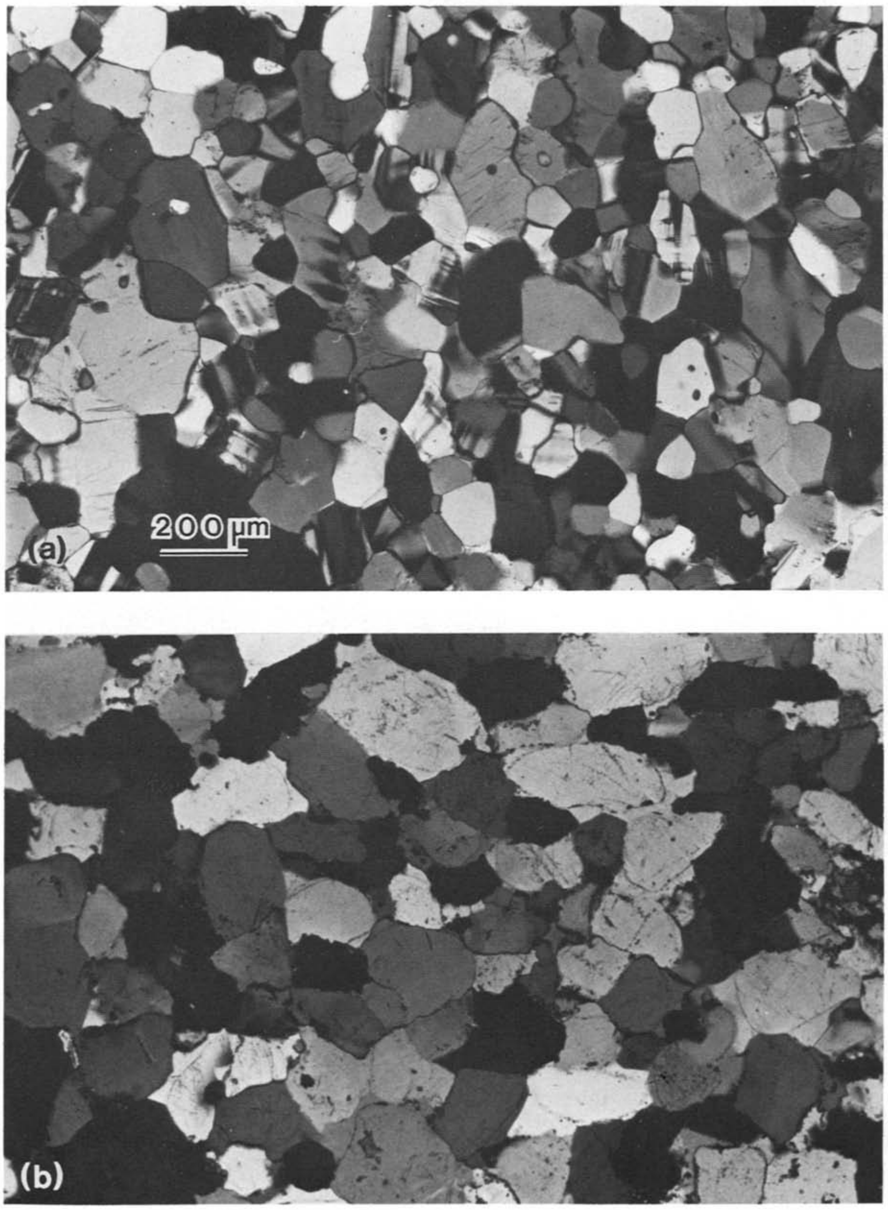

Fig. 1. Optical photomicrographs of undeformed Enfield aplite (a) and Heavitree quartzite (b) under crossed-nicols (and same scale). The aplite contains approximately 30% quartz, 40% microcline, 27% oligoclase, and 3% biotite and magnetite.

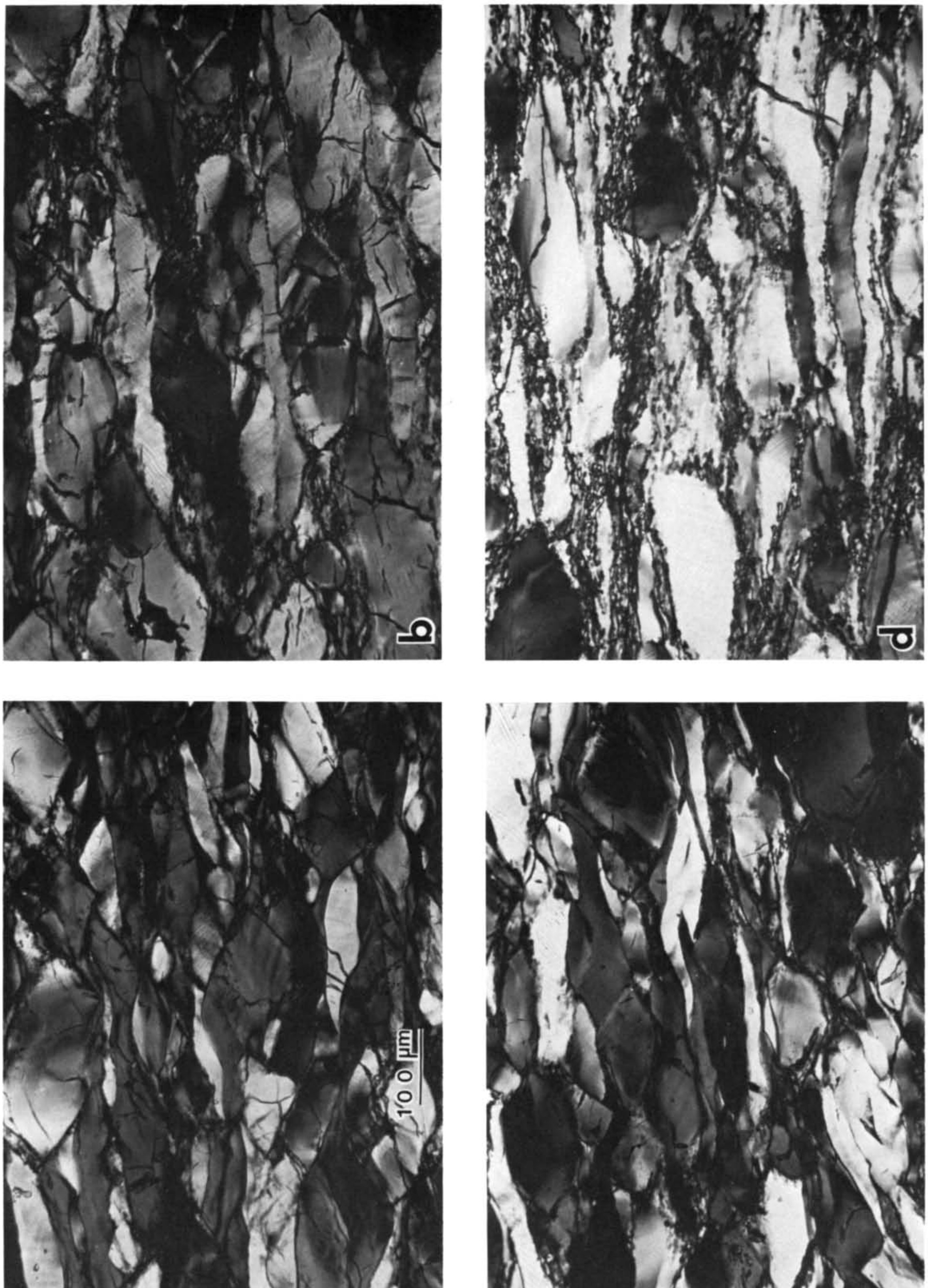

Fig. 7. Optical photomicrographs of Enfield aplite and Heavitree quartzite experimentally deformed at 1500 MPa. All are at the same scale as shown in (a). Compression is N–S. (a) Aplite strained 58% at 800°C, 10^{-6} s^{-1}. (b) Quartzite strained 60% at 900°C, 10^{-5} s^{-1}. (c) Aplite strained 60% at 900°C, 10^{-6} s^{-1}. (d) Quartzite strained 58% at 800°C, 10^{-6} s^{-1}. Note fairly homogeneously flattened quartz grains with only minor recrystallization.

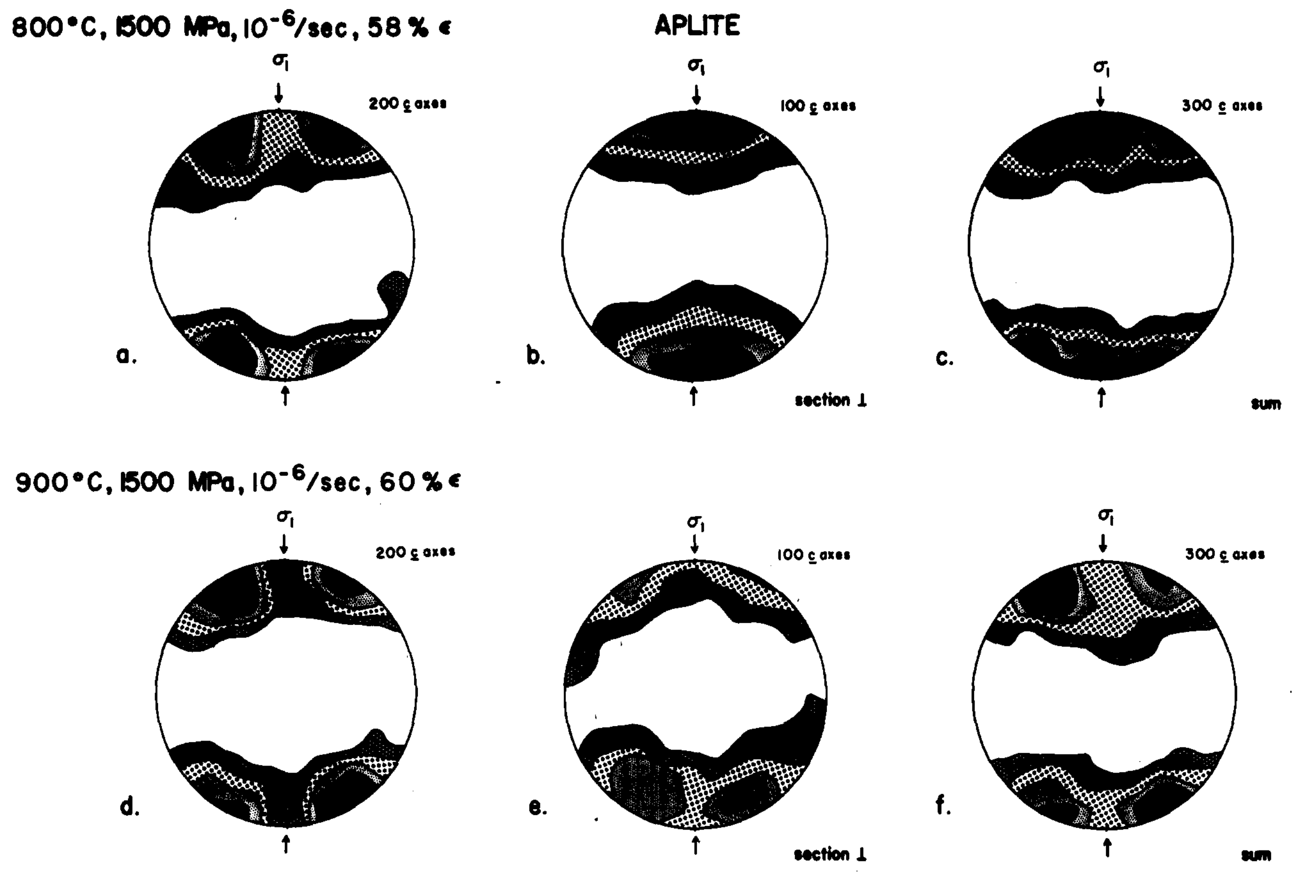

Fig. 3. Quartz c-axis pole figures (equal-area projection) for aplite samples deformed at 10^{-6} s^{-1}, 1500 MPa, to high strain. Kamb (1959) contours of E $- \sigma_1$, E $+ \sigma_1$, E $+ 3\sigma_1$, E $+ 5\sigma_1$. (a)–(c) are for a sample deformed at 800°C; (a) and (b) are from perpendicular thin sections and (c) represents their sum. (d)–(f) are for a sample deformed at 900°C; (d) and (e) are from perpendicular thin sections, and (f) represents their sum.

Measuring techniques

All quartz c-axis preferred orientations were measured in the central region of the sample where deformation is most homogeneous. Thin sections were cut parallel to σ_1 and the c-axis orientation of each of 200–300 grains from each sample were measured optically on a universal stage. The resulting pole figures were contoured using the procedure described by Kamb (1959).

Many samples showed an orthorhombic pattern (e.g. Figs. 3a & d), although one would expect the preferred orientations to exhibit axial symmetry due to the initially random c-axis pattern found in the starting materials and the axial symmetry of the deformation. Similar orthorhombic patterns have been measured in other axially deformed quartz aggregates (e.g. Green 1968, Tullis *et al.* 1973, Jaoul *et al.* 1984); they appear to be an artifact of the optical measurement. This hypothesis was tested by measuring 100 quartz grains in perpendicular (longitudinal) thin sections from two different aplite samples (Figs. 3b & e). The quartz c-axes in one showed a definite orthorhombic pattern (Fig. 3e), which when rotated appropriately and added to the c-axes measured from the original section produced a better axial distribution (Fig. 3f). However, the quartz c-axes in the perpendicular section for the second sample (Fig. 3b) and the sum of the two sections (Fig. 3c) showed axial patterns; the reason for this variation is not understood.

In order to compare the quartz c-axis preferred orientations of aplite and quartzite samples with identical amounts of quartz grain strain, aplite samples were deformed so that their average quartz grain strain was the same as the total sample strain of the corresponding quartzite samples. Presumably, sample strain equals average grain strain for a monomineralic aggregate when deformation is predominantly by dislocation creep with little or no recrystallization. The average grain strain was determined by measuring the aspect ratios of 200 grains parallel and perpendicular to σ_1. Longitudinal strain was calculated for each grain assuming constant volume and initially spherical grains. The values were then averaged.

RESULTS

Quartz c-axis orientations were measured in four aplite samples deformed at 10^{-6} s^{-1}, including low and high strain samples at 800 and 900°C (Figs. 4 and 5). For comparison, c-axes were also measured in two high strain quartzite samples deformed at equivalent conditions (Fig. 6). The flow stresses, sample strains and average grain strains for these experiments are listed in

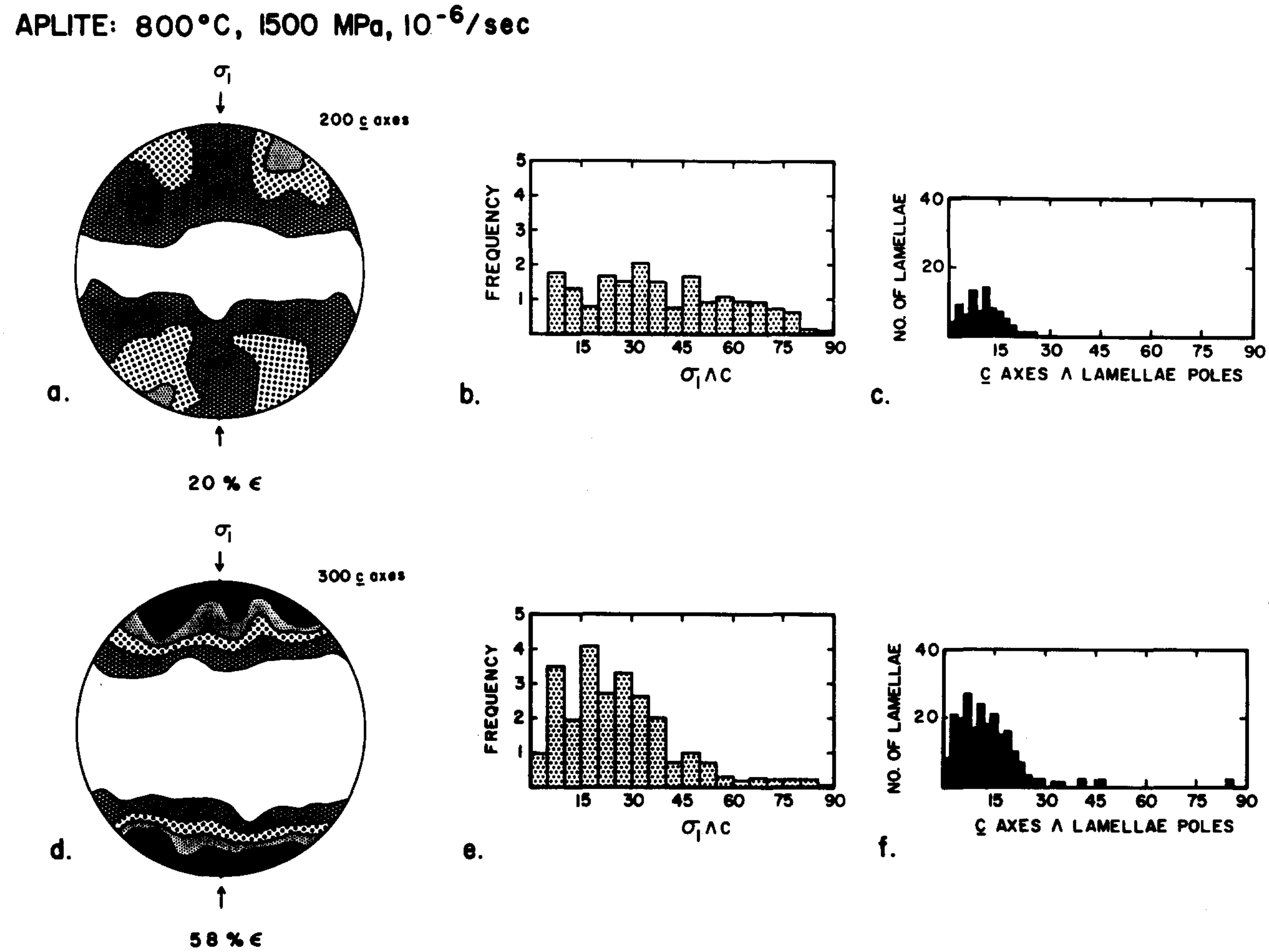

Fig. 4. Quartz preferred orientation data (equal-area projection) for aplite samples deformed at 800°C, 10^{-6} s^{-1}, and 1500 MPa. (a)–(c) are for a sample with 20% strain and (d)–(f) are for a sample with 58% strain. (a) and (d) are pole figures of 200 *c*-axes. (b) and (e) are frequency histograms showing *c*-axis profiles (uniform distribution = 1.0). (c) and (f) show orientations of poles to deformation lamellae.

Table 1 and the optical microstructures are illustrated in Fig. 7.

Aplite

The aplite samples deformed at both 800 and 900°C (and 10^{-6} s^{-1}) show progressive flattening of both the quartz and feldspar grains with increasing strain. At 800°C the grain strains of the quartz and feldspar are almost identical to each other and to the sample strain (Table 1). Both phases are homogeneously flattened, with little evidence of syntectonic recrystallization (Fig. 7a). At 900°C, the quartz grain strain is somewhat greater than the feldspar grain strain and the sample strain (Table 1). The quartz grains are homogeneously flattened, and in the high strain sample there is a small amount of fine-grained recrystallization along most quartz–quartz grain boundaries (Fig. 7c).

The quartz grains in aplite samples deformed at both 800 and 900°C develop a small circle girdle pattern of *c*-axes. At 800°C in the low strain sample (20% shortening), the girdle is broad and weak, with a maximum of 2× uniform distribution, centered at 30° to σ_1 (Figs. 4a & b). In the high strain sample (58% shortening) the girdle is much sharper and stronger, with a maximum of 4× uniform distribution, and an opening angle of 15–20° to σ_1 (Figs. 4d & e). Deformation lamellae were observed in 40 and 72% of the grains in these samples, respectively; they are almost entirely sub-basal in orientation (Figs. 4c & f).

At 900°C the results are similar in that for the low strain sample (22% shortening) the girdle is broad and weak, with a maximum of only 2× uniform distribution, centered about 30° to σ_1 (Figs. 5a & b). In the more highly strained sample (60% shortening) the girdle is sharper and stronger, with a maximum of 3.5× uniform distribution, and is centered at 25–30° from σ_1 (Figs. 5d & e). Deformation lamellae are observed in 40 and 68% of the grains of these two samples, respectively, and again are predominantly sub-basal in orientation (Figs. 5c & f).

In summary, for aplite samples deformed at both 800 and at 900°C and 10^{-6} s^{-1}, the *c*-axes of flattened original quartz grains exhibit a small circle girdle pattern whose strength increases with increasing quartz grain strain. The opening angle of the girdle increases with increasing temperature, from 15–20° at 800°C to 25–30° at 900°C. This is consistent with dominantly basal slip but increasing amounts of prismatic slip at higher temperatures (Tullis *et al.* 1973). The deformation lamellae orienta-

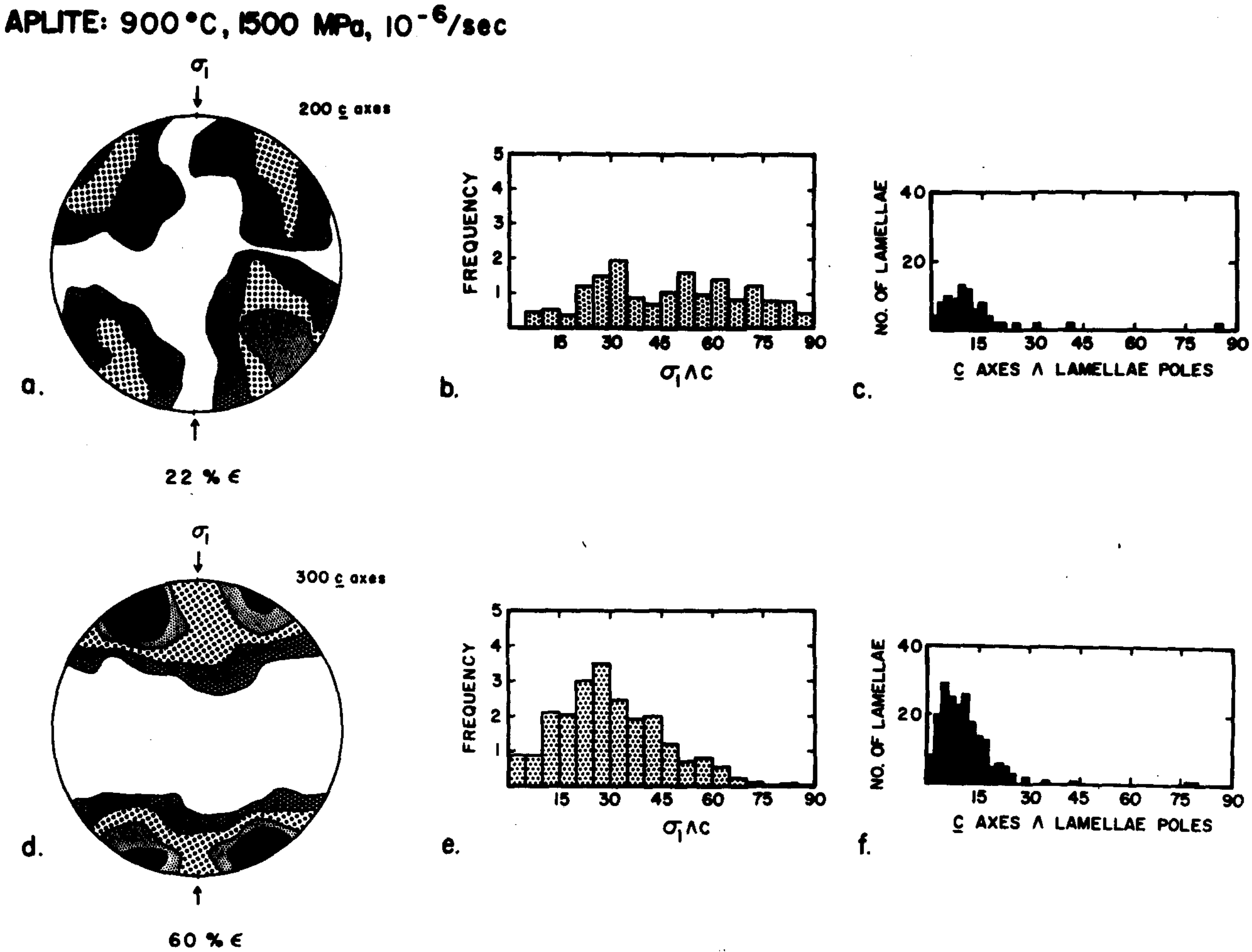

Fig. 5. Quartz preferred orientation data for aplite samples deformed at 900°C, 10^{-6} s^{-1}, and 1500 MPa. (a)–(c) are for a sample with 22% strain, and (d)–(f) are for a sample with 60% strain. The format is the same as for Fig. 4.

tions for the two temperatures appear to be identical, with very few prismatic lamellae; however, lamellae resulting from prismatic slip are usually very faint and hard to discern. This may be due to enhanced recovery at higher temperature, which would eliminate the density of dislocations along slip planes and decrease the optical effect.

Quartzite

Ideally, we would like to compare the *c*-axis preferred orientations of quartz grains in an aplite with those in a pure quartz aggregate deformed at identical experimental conditions. However, this is feasible only if the quartz grains in the two materials have the same type and amount of structurally incorporated 'water'. Within the dislocation creep regime even minor differences in water content can cause significant changes in the strength of quartz (e.g. Mainprice & Paterson 1984, Kronenberg & Tullis 1984). Differences in the microstructural behavior between the quartz in the Enfield aplite and the Heavitree quartzite deformed at the same experimental conditions suggest that the former is effectively drier. Therefore, in order to compare the quartz preferred orientations in these two rocks, we have chosen deformation conditions for the quartzite which produce the same microstructures as observed in regions of several adjacent quartz grains in the aplite samples described above.

Finding 'equivalent' deformation conditions for the aplite and quartzite involved various approximations and assumptions. Heavitree quartzite deformed at 800°C and 10^{-6} s^{-1} appears to be a good match for the quartz grains in the Enfield aplite deformed at 900°C and 10^{-6} s^{-1}. The microstructures are very similar (Figs. 7c & d), although the quartzite contains more recrystallization because it has more quartz–quartz grain boundaries, where recrystallization most frequently initiates. There is no evidence that the grain boundary recrystallization changes the quartz preferred orientation; Tullis *et al.* (1973) found that the preferred orientation of original grains in experimentally deformed quartzites showed a gradual change with increasing deformation temperature, but no sharp change associated with the onset of recrystallization. The flow strengths of the 800°C quartzite and the 900°C aplite are also similar (350 and 450 MPa, respectively). Although the aplite begins with a stress-supporting framework of feldspar, at these conditions the quartz shows higher grain strain and becomes increasingly connected above about 40% strain, suggesting that the quartz gradually comes to dominate the flow

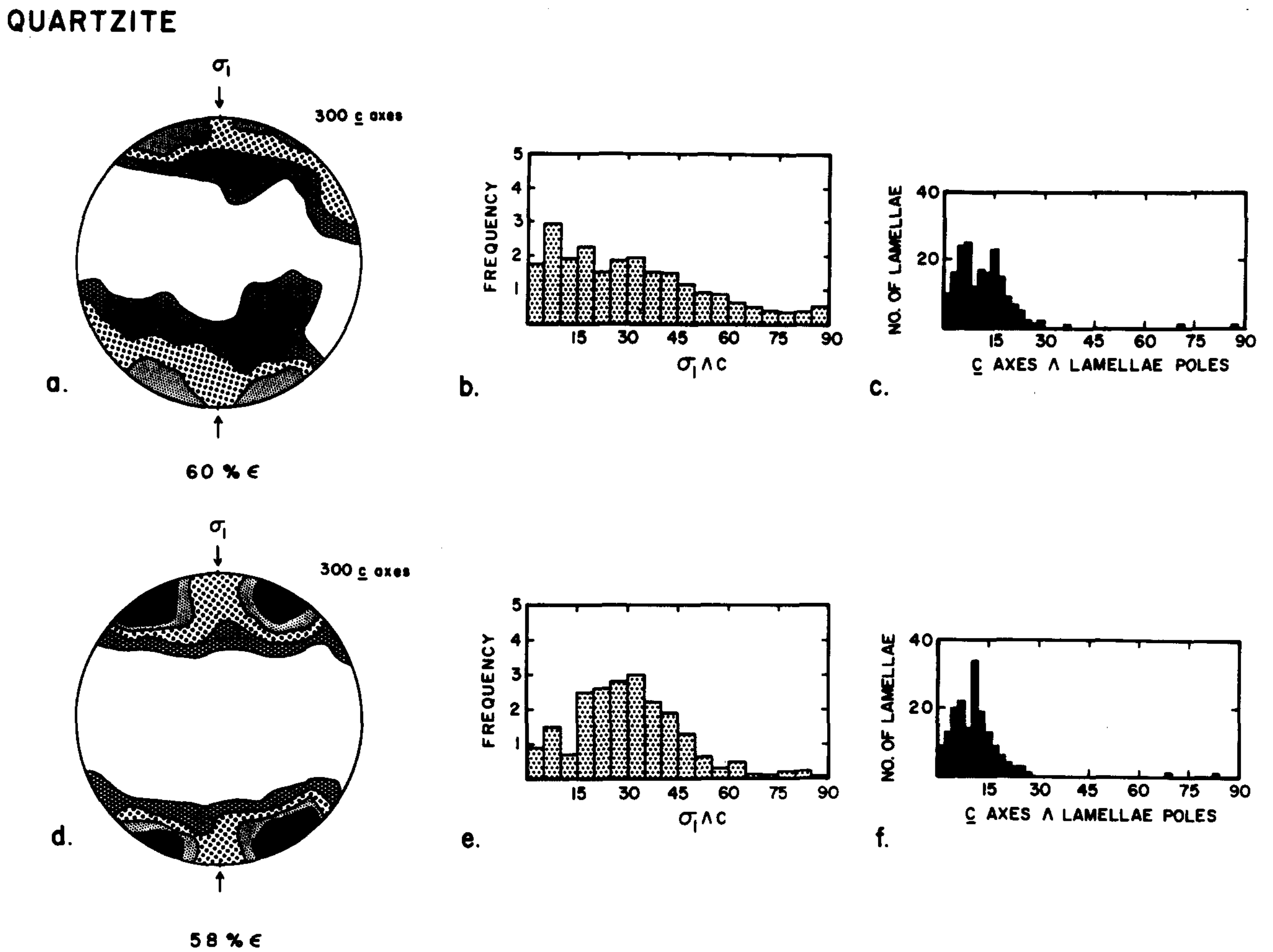

Fig. 6. Quartz *c*-axis preferred orientation data for quartzite samples deformed at conditions equivalent to the high strain aplite samples. The format of the data is the same as in Figs. 4 and 5. (a)–(c) are for quartzite strained 60% at 900°C, 1500 MPa, and 10^{-5} s^{-1}, and should be compared with aplite data in Figs. 4(d)–(f). (d)–(f) are for quartzite strained 58% at 800°C, 1500 MPa and 10^{-6} s^{-1}, and should be compared with aplite data in Figs. 5(d)–(f).

stress of the aggregate (Dell'Angelo & Tullis 1982, in preparation).

Heavitree quartzite deformed at 900°C and 10^{-5} s^{-1} is a reasonably good match for the quartz grains in Enfield aplite deformed at 800°C and 10^{-6} s^{-1}. The yield stress of the quartzite is lower than that of the aplite (600 vs 950 MPa, respectively), because that of the latter is dominantly determined by the stronger feldspar grains, which do not lose connectivity. The microstructures of the quartz grains in the two materials are very similar (Figs. 7a & b). The quartzite does contain a very minor amount of melt (approximately 2%) along some triple junctions, due to impurities. However, comparisons of the microstructures and preferred orientations of this sample with those of other quartzites without melt deformed at the same conditions indicate that the melt has not influenced the development of the quartz *c*-axis preferred orientation.

The quartzite deformed at 900°C and 10^{-5} s^{-1} has a *c*-axis preferred orientation pattern very similar to that of the quartz in the aplite deformed at 800°C and 10^{-6} s^{-1} (Figs. 4d and 6a). The *c*-axes form a small circle girdle about σ_1 with an opening angle of 5–10° and 3× maximum distribution (Figs. 6a & b). Approximately 62% of the grains have deformation lamellae and almost all of them are sub-basal in orientation (Fig. 6c).

The quartzite deformed at 800°C and 10^{-6} s^{-1} exhibits a small circle girdle pattern of *c*-axes with an opening angle centered at 30–35° to σ_1 and a maximum of 3× uniform distribution (Figs. 6d & e). Approximately 58% of the grains have measurable deformation lamellae, again primarily sub-basal in orientation (Fig. 6f). The pattern and strength of the preferred orientation in this sample are almost identical to those for the corresponding aplite, deformed at 900°C and 10^{-6} s^{-1} (Figs. 5d–f).

DISCUSSION AND CONCLUSIONS

Our experimental study has demonstrated that for conditions where quartz and feldspar deform by dislocation creep with little or no syntectonic recrystallization, quartz grains which are surrounded by feldspar grains develop the same pattern and strength of preferred orientation as those in a pure quartzite, if one compares samples in which the quartz grain strains are approximately equal. In both cases the strength of the pattern increases with increasing sample strain; this is consistent with results of Tullis *et al.* (1973) and Green (1968). Our results indicate that quartz–feldspar grain boundaries do not provide any detectably different constraints on slip within the quartz grains and that grain boundary sliding (at these conditions) is as unlikely to occur at quartz–feldspar grain boundaries as it is at quartz–quartz grain boundaries. Grain boundary sliding has not been observed in silicate rocks under experimental conditions even with very fine grain sizes (e.g. Kronenberg & Tullis, 1984).

There are several possible limitations in applying our results to naturally deformed quartzo-feldspathic rocks. At the conditions of our experiments, both the feldspar and the quartz deform by dislocation creep. Feldspar is somewhat stronger than quartz at these conditions, but it still develops approximately the same grain strain as the quartz because it initially forms the continuous, stress-supporting framework. This is not always the case in naturally deformed quartzo-feldspathic rocks. In rocks where relatively undeformed feldspar clasts are isolated in a matrix of highly deformed quartz, the feldspar may have two separate but competing effects on quartz preferred orientations. First, for two volumes of rock which undergo the same type and magnitude of strain, one consisting of just quartz and the other of half quartz and half (rigid) feldspar, the latter should show a stronger quartz preferred orientation because the quartz grains will have to undergo twice as much strain to accomplish the same total rock strain. As is demonstrated in our results, the strength of the preferred orientation is strongly dependent on the finite strain in the quartz. Second, the deformation of the quartz in a rock containing rigid feldspar augen may be more inhomogeneous, causing local variations in the preferred orientations as demonstrated by Lister & Price (1978); this may result in an overall pattern which is more diffuse and lower in strength than would be the case for a pure quartz rock.

There are two other possible limitations of the applicability of our experimental results to natural deformations. First, in most rocks naturally deformed at greenschist grade or higher, the quartz grains are completely recrystallized. It might be expected that the effect of rigid feldspar on the development of preferred orientations in such cases should be even less, because the mobility of the quartz boundaries would make them more independent of adjacent phases with a different deformation behavior. However, the results of Price (1978) indicate that rigid feldspars may have a local effect on the orientation of quartz *c*-axes. This remains to be tested experimentally. Second, the symmetry of our axial compression experiments is higher than that of most natural deformations. However, there is no reason to anticipate any major difference in the effect of feldspar on the quartz *c*-axis preferred orientations for more general deformations, because the important factor is how much grain strain (by slip) the quartz undergoes.

In summary, comparison of deformation experiments on aplite and quartzite demonstrates that the development of quartz *c*-axis preferred orientations is not affected by the presence of feldspar, for conditions where quartz deforms predominantly by slip with only minor syntectonic recrystallization. At such conditions, the pattern of the preferred orientation depends on the active slip systems and the strain geometry while the strength of the preferred orientation depends mostly on the amount of finite strain. If the presence of a second phase causes a decrease in the average quartz grain strain resulting from slip, a change in the operative slip system, and/or an increase in the amount of grain boundary sliding, then that phase would affect the development of a preferred orientation in quartz. The presence of feldspar, however, does not have such effects. We see no reason why the presence of feldspar should have any greater effect under the more common natural deformation conditions involving more general strains, isolated rather than interconnected feldspar grains, and completely recrystallized quartz, but this should be checked by careful field observations and additional experiments.

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