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Grain-size sensitive deformation of a stretched conglomerate from Plymouth, Vermont

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Abstract—The finite strain of clasts (maximum aspect ratio varying from 2 to 40) in a deformed conglomerate from Dry Hill, Plymouth, Vermont, correlates inversely with the average grain size (300–150 μm) in the clast, suggesting that the operative deformation mechanism was grain-size sensitive. In a general way, the average quartz grain size appeared to be smaller in those clasts with higher volume of minerals other than quartz. Dislocation densities varied by as much as a factor of 10 from grain to grain within a clast, but the average dislocation density was relatively constant from clast to clast. If grain-size sensitivity of strength is accepted as a working hypothesis, other elements of the microstructure, such as grain flattening, grain morphology, and dislocation structure can be reconciled as happening either through a late, low strain, high stress pulse—if the current palaeostress indicators are correct to within a factor of 10—or as happening concurrently with the grain-size sensitive mechanism if the current palaeostress estimates are in error. The evidence from this study agrees with several previously published suggestions that grain-size sensitive deformation occurs in the crust for quartzose rocks with grain size of 100 to 300 μm at temperatures of 350 to 420°C.

INTRODUCTION

It is now widely accepted that rocks may flow plastically under geologic conditions. However, many questions remain about the mechanisms of this remarkable phenomenon. One difficulty is that both the stresses active during the deformation and the plastic strain rates are usually unknown; were they known, intelligent guesses might be made about the plastic mechanism. Study of deformed objects of known initial shape appears fruitful (Ramsay 1967); here, at least, total plastic strains can be determined. Deformed conglomerates fall into this category. We have studied a deformed conglomerate which, in addition to strained clasts, had initially different grain sizes and mineralogy. The interplay of grain size and plastic strain in this rock provides unique insight into the plastic mechanisms.

The rocks we studied, from the Plymouth area in central Vermont, have historical significance as they were probably the first deformed conglomerates described in North America (Hitchcock *et al.* 1861). In fact, they were the basis of a major dispute among leading American geologists of the time, the issue being whether or not the pebbles in the conglomerate had been plastically deformed. Few American geologists in 1861 were ready to accept rock plasticity; Hitchcock *et al.* (1861) argued strongly in its favour, based on the shapes of the clasts and the progressive 'stretching' evident as one went from less to more strongly deformed parts of the formation. Hitchcock's position in this argument was remarkable, in that the mechanisms of solid plasticity would not become known even for metals until about 75 years later!

Thompson (1950) first described in detail the conglomerates near Ludlow and Plymouth, Vermont (Fig. 1), and reported several features which stimulated the present study. The conglomerate contains clasts of several different materials, including quartzite, gneiss, and pegmatitic quartz or feldspar. The clasts were deformed into roughly ellipsoidal shapes, the amount of deformation depending on composition and grain size. The order of increasing plasticity as deduced from relative aspect ratios was pegmatite, quartzite, and gneiss: the order is the same as that for boudinage (Thompson 1950). Thompson's key field observation was that increasing plastic strain of a clast was related to decreasing grain size within the clast,

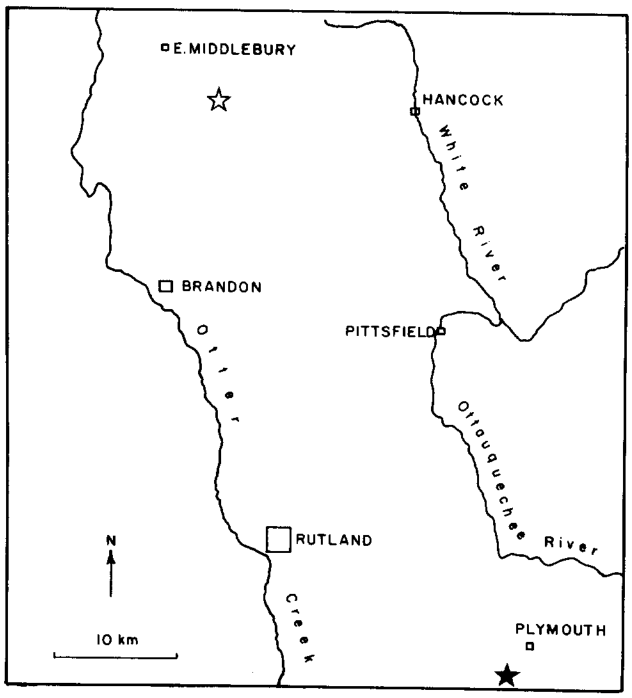


Fig. 1. Index map. Open star shows the location of the undeformed conglomerate, and the solid star, the Dry Hill area where highly deformed conglomerates occur.

and it is on the basis of this observation that we will suggest a mechanism of plastic flow.

The particular beds of conglomerate we chose to study, at Dry Hill, 1.2 km north of Tyson, Vermont, are highly deformed; the ratio of major to minor axes of the clasts ranges up to 40. Plastic strain of the clasts could therefore be determined approximately (Ramsay 1967) and compared as a function of grain size and mineralogy. Relatively undeformed conglomerates do not crop out in the Plymouth area; as described below, we collected some of this material elsewhere.

Geologic setting

The conglomerate at Dry Hill belongs to the Tyson formation, the basal member of a thick Cambro–Ordovician clastic sequence. The Tyson formation rests unconformably on gneiss, schist, quartzite, and marble of Precambrian age, and clasts in the Tyson conglomerate are derived from these rocks. Argillaceous rocks in the Tyson formation contain biotite, chlorite, and muscovite, and rocks of volcanic derivation, epidote, albite, ankerite, and calcite (Thompson 1950, Brace 1953).

The geologic structure in this part of Vermont is dominated by the so-called Green Mountain anticlinorium, which trends about north–south. The Dry Hill exposures of the Tyson formation belong to the east limb of this large fold (Hitchcock *et al.* 1861, Thompson 1950, Brace 1953), and bedding dips between 40 and 50 degrees eastward. The principal foliation on this limb is nearly everywhere parallel with bedding and the longest and intermediate pebble axes are parallel with the foliation. Several isoclinal minor folds are exposed on Dry Hill; their wave length is 2–5 m and they plunge 10–15° north. The vergence indicates relative upward movement on the east. Pebble long axes follow the bedding foliation around these small folds and are everywhere nearly perpendicular to the fold axes.

The conglomerates in the Tyson formation are lenticular units up to a few tens of metres thick, interbedded with quartz-rich metagreywackes and thin beds of dolomite and sericite–chlorite–albite schist. Rare graded bedding and scour-and-fill structure can be found in the Plymouth area. The clasts in the conglomerate range in length from 1 cm to 1 m with maximum–minimum axial ratios from 2 to 40. The conglomerate matrix makes up 20–80% of the volume of the rock and is composed of quartz, microcline, albite, sericite, chlorite, biotite, epidote, and magnetite–ilmenite. When sufficient mica is present in the matrix, or in interbedded schist, a weak crenulation cleavage is evident. Its intersection with bedding defines a lineation which is nearly parallel with the fold axes and, therefore, subnormal to maximum pebble elongation (Fig. 2).

Undeformed conglomerates are rather unusual in central Vermont, and we collected samples in one of the few areas where they are described (Osberg 1952, fig. 2). This is near East Middlebury, 1.1 km east of the summit of Mt. Mossalamoo (Fig. 1). Lenticular bodies of conglomerate are abundant near the base of this lowest Cambrian formation and consist of clasts 3 mm to 2 m in dimension in a schistose matrix of quartz, chlorite and muscovite. The clasts are well rounded with an average sphericity of 0.76 (Osberg 1952). We measured the ratio of maximum to minimum diameter of some fifty clasts and found that the ratio was typically 1.5; a few reached 2.0. The more elongate clasts had long axes parallel with the foliation in the schistose matrix.

Assuming that metamorphism and deformation of the conglomerates of central Vermont occurred at the same

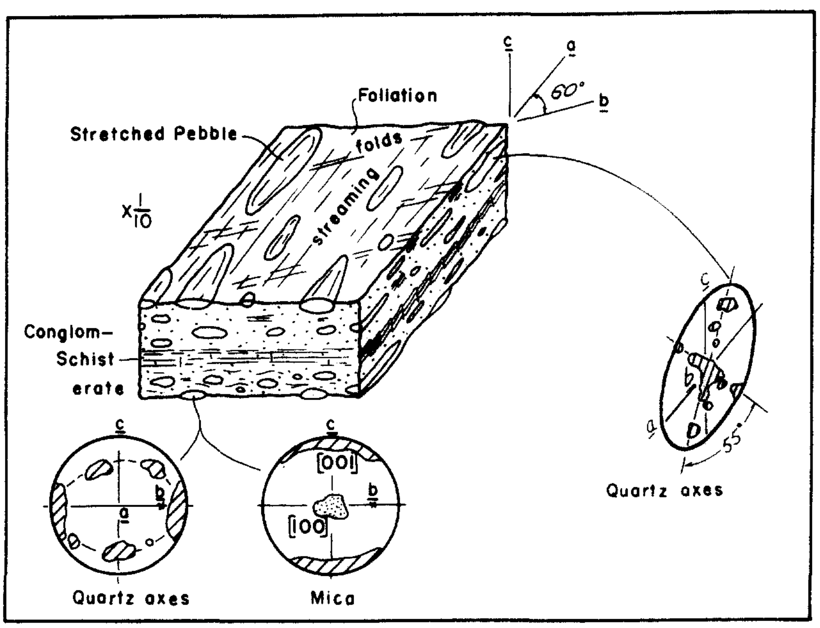


Fig. 2. Schematic relation of stretched pebbles to compositional banding and fold axes. The fabric of quartz [(0001) axes] and mica are also shown schematically in relation to pebble long axis, *a*, and fold axis, *b*. Axis *c* is pebble short axis (Brace 1953).

time, then limits can be placed on the pressure and temperature at the time of deformation. Oxygen isotope fractionation determinations between quartz and calcite suggest temperatures between 370 and 420°C for rocks similar to the Tyson formation, in this part of Vermont (Schwarz *et al.* 1970). Pressure is more difficult to bracket; Thompson (personal communication 1979) suggested a value slightly above the alumino-silicate triple point, namely 450 to 500 MPa (1 MPa = 10 bar).

Rocks used in this study

Dry Hill was chosen for sampling for several reasons. First, the Tyson conglomerate reaches its maximum thickness here and, second, deformation is sufficiently intense. Intense deformation seemed necessary if we were to use the measured strains in the conglomerate quantitatively; since axial ratios of 1.5 : 1 seemed to characterize the relatively undeformed material, then axial ratios in deformed material ought to be as large as possible.

Deformation varies on a scale of 10 m within the Tyson formation, due possibly to variation in stress or lithology. To minimize these effects we chose our deformed samples from a single large block which contained about 100 clasts. A section cut perpendicular to the intermediate pebble axis (Fig. 3) reveals the general aspect of this rock. Grains of feldspar, presumably fragments of pegmatite, appear white and are very nearly equant in cross section. Feldspar grains commonly indent quartzite clasts. Quartzite clasts appear dark grey in Fig. 3, the ends may be pointed or of different curvature (Thompson 1950). Where deformation is extreme the clasts appear as thin parallel layers. Fragments of gneiss appear whitish; even at this scale the influence of grain size is evident. The coarse-grained clasts at the top and bottom are only slightly ellipsoidal; the fine-grained gneiss fragments, in contrast, have flowed into thin, almost continuous bands.

The locality sampled is 0.50 km southeast of the summit of Dry Hill near the 1520 foot contour.

Characterization of the microstructure

Optical and transmission microscopy were used to measure several microstructural parameters in the deformed clasts, including clast aspect ratio, clast mineralogy, grain size, mean free path between mica particles, dislocation density, and dislocation subgrain size. A general review of quantitative stereology which includes most of the methods used in this study is given by Underwood (1970). Most of the measurements were taken from thin sections which were oriented so that the maximum (a) and minimum (c) strain axes were contained in the thin section (see Fig. 2); however, some measurements were made from thin sections oriented perpendicular to a. The measurements are summarized in Table 1 and a brief description of the methods used follows.

The clast dimensions were measured with a set of machinist's calipers directly from the hand specimen. The finite strain (ε) was computed from the aspect ratio.

$$\varepsilon = \frac{2}{3} \ln \frac{a}{c} \tag{1}$$

where a and c are the maximum and minimum axes of the clast respectively, and assuming no volume loss or addition. This assumption may be approximately correct since in most cases the pebbles were not pitted or indented. In addition to uncertainties in ε due to initial departure from sphericity, uncertainties in the clast dimensions arose from the gradual nature of the transition from clast to matrix.

We determined the grain size using a lineal intercept method. This method determines a mean intercept length ($\bar{L}_2$) from the reciprocal of the number of grain boundary intercepts per unit length of test line (N_L):

$$\bar{L}_2 = \frac{1}{N_L} \quad \text{(Underwood 1970, p. 81).} \tag{2}$$

$\bar{L}_2$ is a fundamental quantity in stereographic analysis and is related to the average distance across a grain in the

Table 1. Microstructural parameters of clasts from Dry Hill, Plymouth, Vermont, conglomerate

Aspect ratio (A/c)	Finite strain (ε)	Second phase impurity	Mean intercept length ($\bar{L}_2$)	$\sigma(\bar{L}_2)$	Mica mean free path (λ)	Dislocation density (ρ)	$\sigma(\rho)$	No. of plates	Subgrain boundary spacing
		vol. %	10^{-4}cm	MPa	mm	10^8/cm^2	MPa		10^{-4}cm*
3.1	.75	-	260 ± 90	7	-	2.5	100	10	5.2
5.6	1.15	.2	240 ± 50	8	1.7	2.7	110	25	1.0
10.9	1.59	.4	190 ± 50	9	.8	-	-	-	-
10.6	1.57	.3	200 ± 30	9	.9	-	-	-	-
12.5	1.68	.6	150 ± 20	11	.5	1.1	70	5	-
20.2	2.00	-	160 ± 20	9	-	3.4	120	10	4.0
21.1	2.03	.4	180 ± 50	9	.5	2.9	110	15	>5.0
24.2	2.12	-	170 ± 40	10	-	2.8	110	5	-
46.0	2.55	1.2	140 ± 50	11	.3	-	-	-	-

*Should be regarded as an estimate since no criterion for orientation of the subgrain boundary or the angle of lattice mismatch across the boundary was used.

plane of the thin section. In addition, for space-filling grains, $2N_L = S_V$, where S_V is the average surface area per unit volume of grain. Modal analyses were done on thin sections of clasts which were predominantly quartz ($<5\%$ of other minerals) to determine the purity of the quartz clasts. The 'impurities' or 'second phases' in these clasts were micas, feldspars, magnetite, and other opaques. 2000 points were counted for each modal analysis. The second phase mean free path (λ_{Π}), which is the edge to edge distance between second phase grains, was calculated from the formula

$$\lambda_{\Pi} = \frac{1-(V_V)}{(N_L)_{\Pi}} \quad \text{(Underwood 1970, p. 82)} \tag{3}$$

where $(V_V)_{\Pi}$ is the volume fraction of second phase and $(N_L)_{\Pi}$ is the average number of impurity grains intersecting a unit length of test line.

TEM specimens were prepared by the standard ion beam thinning technique and observed on a Phillips EM 300 microscope operating at 100 keV. From 5 to 20 plates were taken of each TEM specimen. Because the dislocation density was low enough to recognize individual dislocations, the dislocation density was determined by counting the number of intersections per unit area of the dislocation with both the upper and lower foil surfaces (P_A). Assuming that either the foil surfaces or the dislocations are randomly oriented, the dislocation density, ρ, is simply given by

$$\rho = P_A \quad \text{(Underwood 1970, p. 157).} \tag{4}$$

Some authors have used the relation, $\rho = P_A/2$, but it is important to notice that this equation is true only for the very special case of straight parallel dislocations perpendicular to the surface of the foil (Underwood 1970, p. 157, De Hoff & Rhines 1968, p. 87) and that dislocation densities for the randomly oriented case are given correctly by equation (4) above.

DISCUSSION

Brief comparison of the microstructure of stretched and unstretched clasts of the Tyson conglomerate

Assuming that the source rocks for the conglomerates near East Middlebury and Plymouth are similar (Osburg 1952), comparison of the microstructures of the deformed and undeformed conglomerates provides information on the effect of the deformational episode which stretched the clasts. Photomicrographs of three stretched clasts (Figs. 4a–c) from Dry Hill and three undeformed clasts (Figs. 4d–f) from East Middlebury are shown. The minerals in the unstretched clasts show ample evidence of prior deformation of the conglomerate source rocks. The clasts frequently exhibit bimodal grain size distributions, the smaller grain size of which is roughly similar to the optically detectable subgrains of the larger grains. In hand sample, the clasts show gneissic structures of alternating phyllosilicate-rich and -poor layers. In thin section, the phyllosilicates are often observed to be clumped in polycrystalline aggregates (Figs. 4e & f) rather than being widely dispersed, as in the stretched clasts (Figs. 4a–c). The larger grains are commonly very elongate and frequently show patchy and undulatory extinction, indicating abundant dislocation flow. Grain size varies quite widely because of the bimodality of the structure: the largest grains can often be several millimeters in size, while the smaller grain size fraction is typically several tenths of a millimetre in dimension, as are the subgrains of the larger grains. Grain boundaries are typically highly sutured and, in some cases, are very tortuous. Micaceous material may either be clumped, as pointed out earlier, or disturbed along a linear transgranular zone which may indicate the growth of secondarily formed minerals along a crack or fissure.

The structures contrast quite sharply with the grain sizes and shapes found in the stretched conglomerates. In no case was a stretched clast found to have an obviously bimodal grain structure. Phyllosilicates in the Dry Hill rocks were dispersed to a far greater degree than those from East Middlebury. Grain sizes in the stretched clasts were much finer (150–300 μm) than most of those which were not deformed. We believe that these differences provide good reason to suspect that grain boundaries were mobile during the deformational episode which stretched the conglomerates and that abundant recrystallization has occurred. If this is true, it is certain that the dislocation microstructure of the clasts was reset during the stretching of the conglomerate. It also appears that the phyllosilicates may have been 'kneaded' into the clasts, forming a much more diffuse mixture.

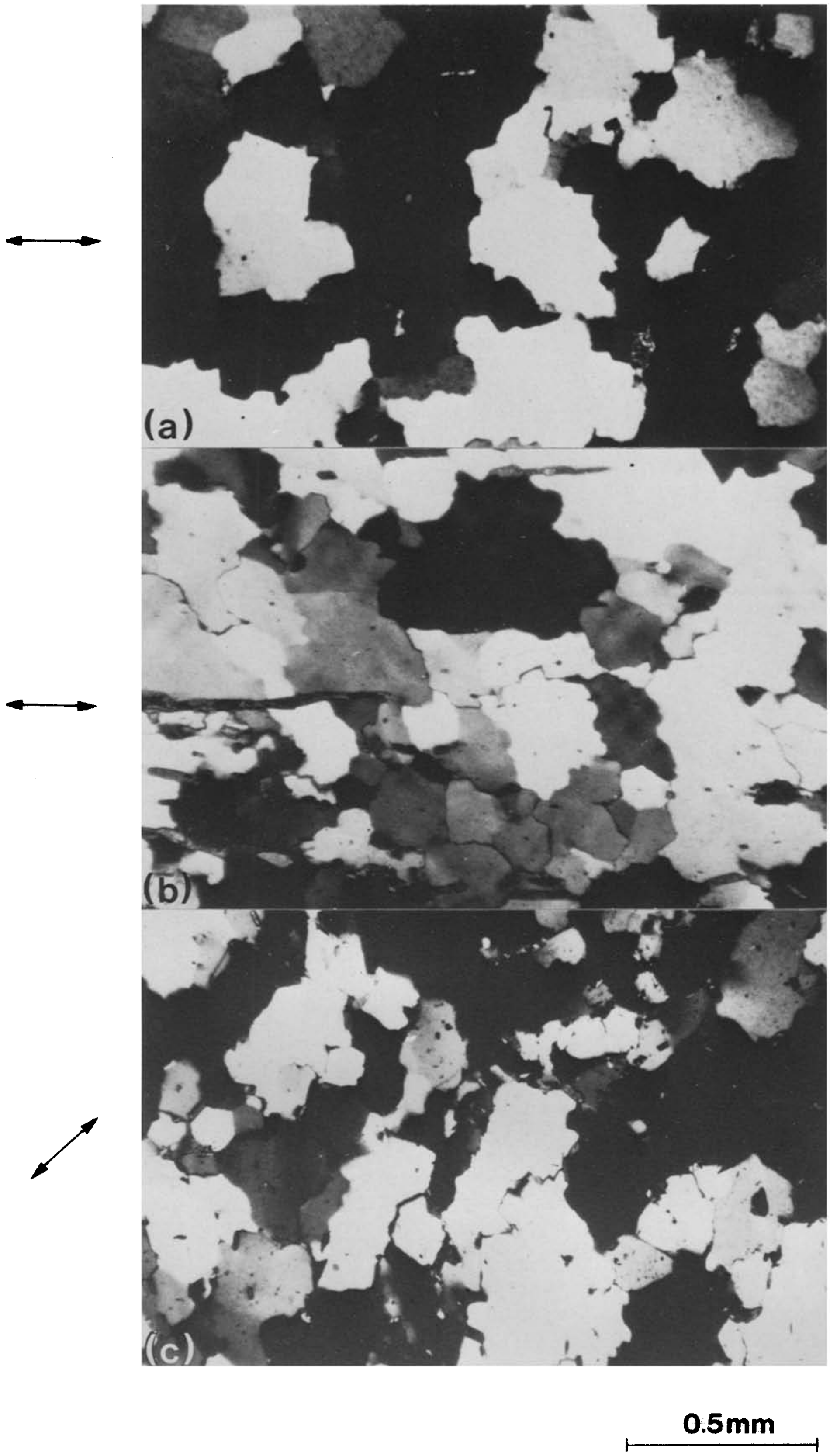
Grain-size sensitivity of strain

We quantitatively verified Thompson's (1950) observation that strain, as measured from the aspect ratio of the clast, and grain size within the clast are related (Fig. 4). Two curves are shown, one for relatively pure quartzite clasts ($>95\%$ quartz) and a second for quartzo-feldspathic clasts. As Thompson observed, and as is a common field observation (Hobbs *et al.* 1976), the quartzo-feldspathic clasts appear to be stronger (by virtue of suffering less strain for a given average grain size than the relatively pure quartzites).

The observation that strain and, by inference, strength, are sensitive to grain size is important in identifying the mechanism of deformation. Deformation by mechanisms variously called Coble creep, Nabarro–Herring creep, grain boundary sliding accommodated by diffusional flow, structural super-plasticity, or pressure solution would be expected to show a grain-size dependence of strength (Stocker & Ashby 1973), whereas deformation by dislocation mechanisms would show a much smaller sensitivity of strength to grain size (Hirth & Lothe 1968). Put in terms of strain rate, grain-size sensitive deformation could be represented by a general equation of the form

$$\dot{\varepsilon}_{\mathrm{GSSM}} = \frac{\dot{\varepsilon}(T,\sigma,P)}{d^m} \tag{5}$$

Fig. 3. Section of the sampled conglomerate cut perpendicular to the intermediate pebble axis. Quartzo-feldspathic and feldspar clasts appear whitish against the darker sericite mica-rich matrix, while 'purer' quartz clasts appear greyish. The scale bar represents 5 cm.


(a)
(b)
(c)
0.5mm

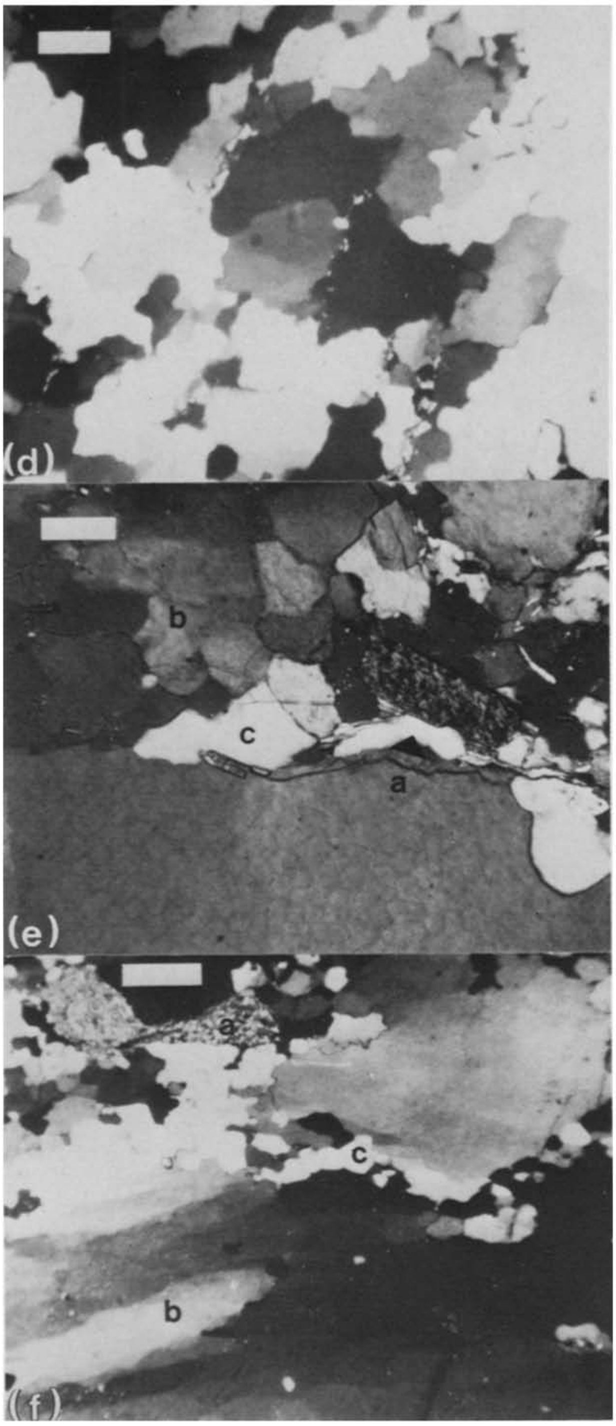


Fig. 4. Grain structure of three deformed clasts (a, b, c), dominantly quartz, and three undeformed clasts (d, e, f), as seen in thin section, transmitted light, crossed polars. The strain suffered by the deformed clasts is as follows: (a) $\varepsilon \cong 1.1$, (b) $\varepsilon \cong 1.6$, (c) $\varepsilon \cong 2.03$. The direction of principal stretching is shown by the arrows next to each micrograph, and the scale for these three micrographs is given. In contrast to the relatively equigranular structure of the stretched conglomerate clasts (a, b, c), the unstretched clasts (d, e, f), frequently show highly sutured grain boundaries, extensive elongate subgrain structure (marked on photos 6e, 6f by the letter b), and a duplex grain structure including a fine-grained fraction (indicated on the photos by the letter c). The phyllosilicate fraction of the unstretched clasts is frequently clumped (shown at a in Fig. 6e and 6f) rather than dispersed as shown for the stretched clasts (Fig. 6a–c). Scale bars for 6d–f are shown for 0.3 mm, 0.3 mm and 0.1 mm respectively.

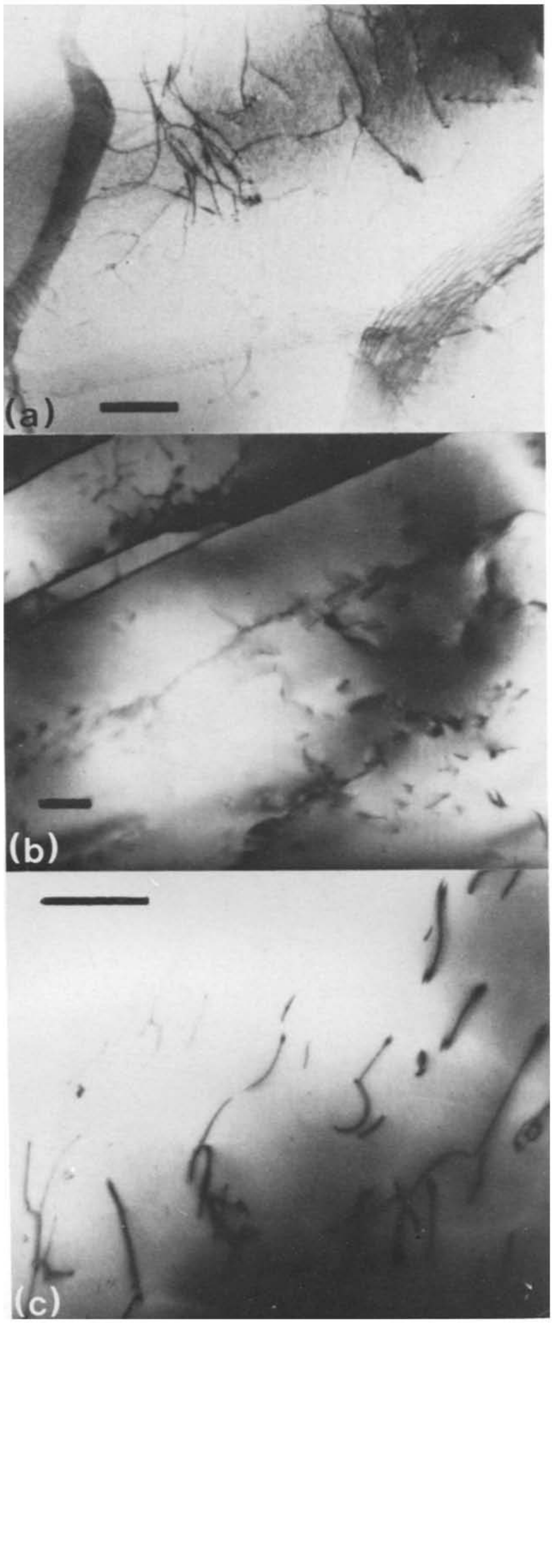


Fig. 7. Dislocation microstructure of three dominantly quartz clasts which tends to indicate cross-slip and climb processes occurring either during or after dislocation generation. Note loops and nodes in (c) and branching subboundary in (b). Strain and dislocation density are $\varepsilon = 2.12, 1.15, 1.15$ and $\rho = 2.8 \times 10^8\ \mathrm{cm}^{-2}, 2.7 \times 10^8\ \mathrm{cm}^{-2}, 2.4 \times 10^8\ \mathrm{cm}^{-2}$ respectively. Scale bars are 1 μm.

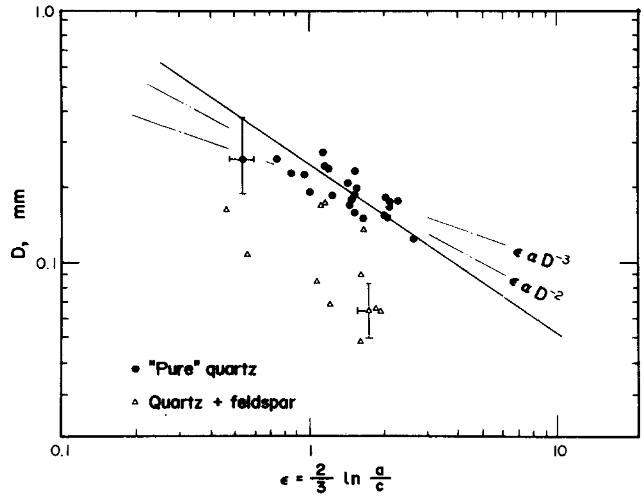


Fig. 5. Observed strain in pebbles, ε, vs grain size, D. Heavy line is least squares fit ($r^2 = 0.62$) to 'pure' quartz data and is $\ln \varepsilon = -1.46 \ln D - 2.1$. Typical error bars are shown. Dashed lines are shown for ε proportional to D^{-2} and D^{-3} respectively.

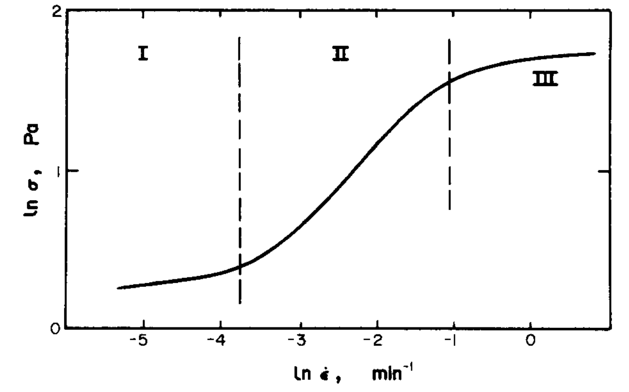


Fig. 6. Typical superplastic behavior (Edington *et al.* 1976), shown here for an Mg–Al alloy, is characterized by rapid increase in flow stress, σ, with strain rate, $\dot{\varepsilon}$. Some phenomenological characteristics of superplastic behaviour corresponding to the regions shown in this figure are given in Table 2. Notice that deformation under conditions appropriate to region III implies a substantial contribution by dislocation motion.

where m is 2–3, depending on the exact mechanism or group of mechanisms operating. In contrast, the strain rate for dislocation mechanisms ($\dot{\varepsilon}_{\text{dis}}$) is experimentally found to depend explicitly on grain size in a weaker fashion (Hirth & Lothe 1968); m is usually found to be about -0.5 for these mechanisms.

It would be difficult, if not impossible, to identify a particular purely-diffusional mechanism on the basis of microstructure alone. Furthermore, some of the processes named above are actually a combination of separate mechanisms. Edington *et al.* (1976) divide the superplastic regime into three regions, the first two of which have high grain-size sensitivity of strain rate and limited dislocation flow, and a third which has low grain-size sensitivity of strain rate and extensive dislocation flow (Fig. 6). In view of the indeterminacy of the exact diffusional mechanism, we prefer to use a generic term, grain-size sensitive mechanisms (GSSM), to distinguish these mechanisms from those involving dislocation motion (including region III of Fig. 5).

Several previous authors have suggested the operation of various GSSM under natural deformation conditions: Boullier & Guegen (1975), Guegen & Boullier (1976), and Twiss (1976) have suggested that superplastic flow may operate in the mantle on the basis of microstructural characteristics, while Elliott (1973), Rutter (1976), Mitra (1977), Kerrich *et al.* (1977), McEwen (1978) and others have discussed the possibility of diffusional flow in crustal rocks. It seems useful, therefore, to compare the textures and microstructures observed during experiments on materials deformed in the GSS field to these natural samples which we suspect to have been deformed by a similar mechanism.

Edington *et al.* (1976) summarized the important changes in the microstructure of metals undergoing superplastic deformation in a thorough review article (a brief summary is given in Fig. 6 and Table 2). For materials deformed in the GSS region, i.e. in regions I and II of Edington *et al.*'s (1976) classification, one should expect to observe more or less equiaxed grains and reduced textures (Table 2). In contrast to the plethora of data on metals, information on minerals in the GSS field is very limited (Rutter 1976, Schmid *et al.* 1977, Twiss 1976,

Table 2. Superplastic behaviour (after Edington *et al.* 1976)

	Region of σ-ε̇ curve		
	I	II	III
Sensitivity of yield stress to grain size	high	high	low
Dislocation activity		single slip	multiple slip
Grain elongation	limited	very limited	significant
Grain boundary sliding	minor	significant	minor
Texture reduction	minor	significant	minor
Texture retention	none	significant	significant

Schwenn & Goetze 1978). In experimentally deformed Solenhofen limestone, Schmid *et al.* (1977) observed grain elongations which were $\sim \frac{1}{3}$ of the specimen strain, grain-size sensitivity of flow stress and, importantly, some texture production. In addition, they observed a within-grain dislocation structure consisting of a relatively low density core surrounded by a mantle of dislocation-rich material near the grain boundary. These characteristics, with the possible exception of texture production, compare quite closely to those described for metals by Edington *et al.* (1976), particularly in region II (Fig. 6 and Table 2).

In addition, White (1977, 1979a) has suggested that superplastic materials deformed in simple shear are characterized by smooth or undulating grain boundaries parallel to the flow direction and by a grain size which is comparable to or smaller than the equilibrium subgrain size. But because the conglomerate pebbles have been flattened considerably, the strains probably differed considerably from simple shear, and thus parallel grain boundaries would not necessarily be expected.

We examined several other elements of microstructure, including the dislocation microstructure, in order to see if they were compatible with the working hypothesis of grain-size sensitivity of strength suggested by Thompson (1950) (and Fig. 5), and to compare the conglomerate characteristics with the structure of materials known to have been deformed in the grain-size sensitive region. Three important questions arise. Is the dislocation structure compatible with the operation of GSSM? Could the apparent grain-size dependence of strain of Fig. 5 be developed with dislocation mechanisms? Are the other microtextural elements compatible or incompatible with the hypothesis of grain-size sensitivity of strength? Discussion of each of these questions is presented below, followed by a comparison with previously published field studies.

Information from the dislocation microstructure

In thin section, the stretched clasts exhibited evidence of dislocation creep, e.g. slightly sutured grain boundaries and wavy and patchy extinction (Fig. 4). As Edington *et al.* (1976) and White (1976) point out, deformation mechanisms may operate simultaneously under certain conditions. Thus, the presence of dislocations does not preclude the operation of GSSM.

Several typical transmission electron micrographs are pictured in Fig. 7. The dislocation microstructures show both straight and slightly curved low angle tilt walls and higher angle subgrain walls. Single straight dislocations are evenly dispersed throughout the tilt walls. The free dislocation segments appear not to be confined to a single plane or set of glide planes, and some have undergone dislocation reactions forming triple points or dislocation nodes. Small loops are observed and the dislocations are frequently highly curved. The presence of uniformly spaced tilt walls strongly suggests that recovery processes have operated, either concurrently with, or subsequent to, dislocation generation and motion. One striking feature which we observed was variability of the density of the dislocation structures by a factor of 10 within a given grain. Similar inhomogeneities of dislocation structure, termed core and mantle structure (Gifkins 1976) have been noted for superplastically deforming metals (Gifkins 1976, Edington *et al.* 1976) and calcite (Schmid *et al.* 1977). White (1976) suggests that such structures occur in naturally deformed quartzites and feldspathic rocks. However, it is certainly true that variability of strain has been shown to occur in rocks deformed experimentally in the dislocation regime to total strains as large as 40% (Tullis *et al.* 1973). Clearly, variability in dislocation density is probably very common in polycrystalline aggregates, even in purely dislocation flow regimes.

Most grains contained many elongate subgrains often arrayed in subparallel orientations.

Information from palaeostress indicators

Recent experimental evidence has raised the possibility of using the dislocation microstructure and the dynamically recrystallized grain size to estimate quantitatively the stress under which those microstructural elements formed (for example, see works by Goetze 1975, Takeuchi & Argon 1976, Twiss 1977, McCormick 1977, Mercier 1977, Goetze 1978). The application of these experimental results to natural conditions is not entirely straightforward, and many uncertainties arise due, among other things, to second phase effects, temperature effects, and alteration of the structure during uplift (White 1979b). In this paper we wished to use the palaeostress indicators to obtain order of magnitude estimates of the stress, fully cognizant of the fact that future refinements in calibration may alter the exact value of the constants in the equations used. In particular, we used the relations for quartz as given by Weathers *et al.* (1979) in their study on the Moine thrust—which were based on unpublished work done by Goetze and published work of Kohlstedt *et al.* (1976) and Mercier *et al.* (1977)—and the equation for subgrains as given by White (1979a):

$$\sigma = 6.6 \cdot 10^{-3} \sigma^{1/2} \quad (6)$$

$$\sigma = 3.8 \cdot 10^{2} D^{-0.71} \quad (7)$$

$$\sigma = 500/d \quad (8)$$

where σ is the inferred differential stress $(\sigma_1 - \sigma_3)$ in MPa, ρ is the dislocation density in number/cm^2, D and d are the grain size and subgrain size in 10^{-6} m respectively. Some of the workers among those cited above have proposed equations with constants different from those given in equations (6)–(8). The differences between the stresses determined by these various relations are an indication of the present order of error of the determinations (at least a factor of 2–5). Further experimental work will undoubtedly refine the uncertainties involved. The stresses calculated are given in Table 1 next to the microstructural parameter from which they are derived.

Interestingly, there is a disparity between stresses estimated from dislocation density and subgrain size on the one hand and stresses estimated from the grain size on the other hand. This situation could have arisen either from a multistage deformation history or from incorrectly calibrated palaeostress scales. Because grain size is thought to be more stable to small strains than dislocation density, the combination of high dislocation density and large grain size may occur following a short high stress pulse. Such a structure cannot result from an annealing episode since the dislocation density would be reduced either before the grain boundaries became mobile or as they swept through the crystal. This constrains the time of production of the observed dislocations to be either concurrent with or subsequent to the last stage of grain growth (or grain reduction). The apparent lack of correlation of dislocation density with either grain size or strain (Fig. 8) seems to indicate that the stresses which generated the dislocations were approximately constant over the relatively small volume of rock which we sampled, but were highly variable on the grain size level. Likewise, the fact that the subgrain spacing is much less than the grain size—rather than roughly equivalent—is consistent with the notion of dislocation overprinting of a GSSM structure.

Are the other microtextural elements of this rock consistent with the operation of a GSSM grain growth?

Grain growth frequently occurs during superplastic deformation (Edington *et al.* 1976). If the grains were left to continue growing they would eventually become so large that dislocation processes would dominate, at which point the grain size and the dislocation density would equilibrate with the same steady-state stress. The natural system at hand apparently did not equilibrate in this fashion, since the clasts maintained different grain sizes for large strains (and periods of time), even though there is evidence that the grain boundaries were mobile and that there was ample strain to allow grain size equilibration. It is common practice in ceramics engineering to subdue grain growth by 'poisoning' the mobility of grain boundaries by the addition of a dispersed second phase, and it has long been known that, for recrystallizing metals, the final grain size is closely linked to the volume fraction, size, and spacing of the massive second phase (Smith 1948). A recent study shows similar results for ice (Wilson 1979). Field geologists have observed a phenomenon which is apparently a natural analogue to this process, viz. the grain size of quartz is often linked to the second phase—frequently mica—content (Hobbs *et al.* 1976). This indeed appears to be the case for the conglomerate clasts. Although there could easily be a systematic error in the determination of volume fraction of impurities, due at least in part to the presence of uncounted small second phase particles, the grain size is correlated inversely with the volume fraction of impurities (Fig. 9). Thus differences in grain size and, hence, in mechanical strengths for the clasts deforming in the GSS region could have been maintained over large strains by differing volume percent impurity content.

Could the grain-size dependence of strain be achieved by a variation of local stress in the conglomerate?

At least two observations argue against this proposal. TEM observations show that while dislocation density may vary by as much as a factor of 10 from grain to grain within a clast or within a given grain, the dislocation density averaged for an entire clast is relatively constant from clast to clast, irrespective of strain or average grain size (see Fig. 8). If the grain size and dislocation density were at steady state with the same differential stress, the palaeostress equations (6) and (7) imply that a decrease by half in average grain size (from 300 to 150 μm would correspond to an increase of dislocation density of $\sim$ 2.7. Although such an increase could perhaps be concealed in the scatter of the data, the average dislocation density seems to be more constant than this. Secondly, we have observed instances where one clast indents a second finer-grained clast. Since it is difficult to argue for large stress variations when clasts are in such close proximity, this observation tends to argue in favour of real strength contrasts. Thus it appears unlikely that the strain–grain

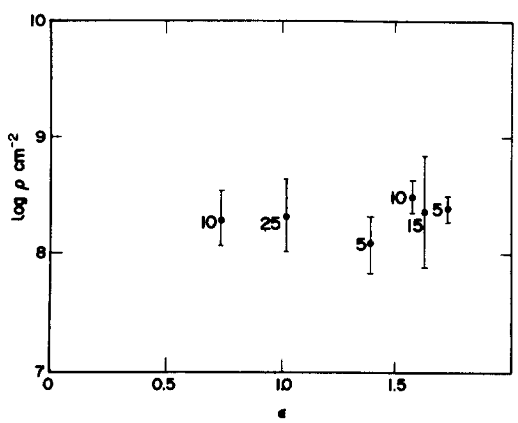


Fig. 8. Dislocation density, ρ, vs strain, ε. The number of plates counted is given beside each point.

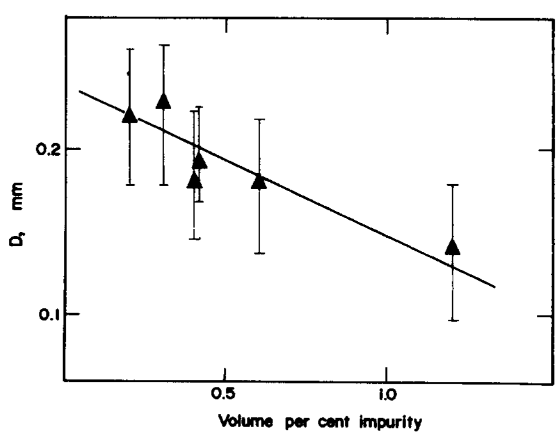


Fig. 9. Grain size of quartzite pebble vs amount of impurity, principally mica.

size relationship arises from dislocation processes.

Grain flattening and grain boundary morphology

Measurements of metals deformed in the superplastic regime show slight grain elongation. Although the grain elongation is much smaller than the total elongation of the specimen, the aspect ratio of the grains may commonly be as large as 1.5: 1 (Edington *et al.* 1976). Schmid *et al.* (1977) observed slight grain size flattening $[(\varepsilon_{grain}/\varepsilon_{total}) \simeq 0.20]$ in their Solenhofen limestone deformed in the superplastic region. Thus, the magnitude of the slight grain elongation observed in the stretched pebbles is certainly not inconsistent with the operation of a grain-size sensitive mechanism (for instance, compare Fig. 6 with Schmid *et al.*'s experiments). The possibility exists that some or all of the flattening could have resulted from a later deformation stage involving dislocation mechanisms. This hypothetical second stage would then be limited in finite strain such that $\varepsilon_{second} \simeq 0.10$–$0.20$. If the current palaeostress indicators are correct within an order of magnitude, the second deformation stage would have been a stress pulse of perhaps ~ 100 MPa, but of insufficient duration in strain to cause significant recrystallization to a smaller grain size. Such a dual phase microstructure is not at all unreasonable, and the fact that a certain amount of strain is necessary to cause the onset of recrystallization is a common field observation. For instance, Weathers *et al.* (1979) observed that during the progressive recrysallization of a mylonite along the Moine thrust, crystal elongations of ~ 2 to 1 occurred when only 5% newly-recrystallized grains were present.

Preferred orientation

Brace discussed the preferred orientation of a suite of samples from seven locations in the Central Vermont area, Dry Hill being among these (Brace 1955, fig. 20). We did no further fabric analysis and only wish to discuss the earlier results in terms of their bearing on the identification of the deformation mechanism. The initial fabric of the Precambrian rocks, which probably have served as the source rocks for the conglomerates, frequently have higher density of C-axes than the deformed conglomerates. Thus there may have been at least partial randomization of texture either by deformation or annealing. Neither the shape of the distribution of C-axes, nor the magnitude of C-axis density, shows any particular correlation with the strain suffered by the clast. In fact, Brace notes that some of the highest percent maxima of C-axis orientations occur in pebbles whose axial ratios range from 6: 1 to 10: 1, whereas in pebbles with ratios nearer to 20: 1 the degree of orientation can be much smaller. The shape of the distribution could be described variously as diffuse point maxima, or single girdles around the axis of extension. If the appropriate strain condition for the pebbles is roughly that of axially symmetric extension, then these patterns are roughly similar in their general features to distributions generated by a Taylor–Bishop–Hill model of dislocation slip (Lister *et al.* 1978). As mentioned above, randomization of texture is consistent with GSS flow. However, because both recrystallization and recovery were probably occurring, the relation between fabric intensity and strain would be far from trivial even if the fabrics were caused exclusively by dislocation processes. Furthermore, as noted above, Schmid *et al.* (1977) observed texture production during grain-size sensitive flow. Therefore, the production or retention of textural elements in these rocks could have occurred during GSS flow (as in region II of Edington *et al.*'s (1976) classification) or during steady state dislocation flow, either instead of GSS flow or subsequent to it during a later stress pulse. Thus one may say only that the textural evidence does not contradict or predict the operation of either GSS flow or purely dislocation flow.

In summary, the microstructural elements of the rock seem to be consistent with the operation of grain-size sensitive flow, either concurrently with some dislocation production or with production of dislocations being caused by a later high stress pulse; with the exception of the preferred orientation, they do not seem to be consistent with deformation entirely by dislocation motion. It is possible that some very complicated history of stress pulses and annealing might also be consistent with the microstructure, but we believe that the present explanation is the simplest history consistent with the evidence and our current knowledge of deformation mechanisms.

Comparison of this result with some previous studies

Elliott (1973) presented a convincing case that diffusional creep operated under some conditions in the crust by examining marker surfaces of inert inclusions, pressure shadows, and indentation of quartz grains in the Weaverton formation from Maryland. Mitra (1977) estimated the relative amounts of strain due to diffusional pressure solution flow and intracrystalline dislocation flow by comparing the geometry of pressure shadows with deformed intragranular rutile needles. Both authors presented convincing arguments that the pressure shadows resulted from diffusional pressure solution processes. We noticed no obvious pressure shadows in these rocks, but it is not clear that the appearance of pressure shadows is required under large strain diffusional flow enhanced by a grain boundary 'fluid' phase. If the effect of the fluid phase on grain boundaries is simply to enhance the grain boundary diffusivity without fundamentally altering the physics of diffusional deformation processes, the microstructures produced with fluid present might be quite similar to those observed in superplastically deformed materials. We did not, however, attempt to analyze impurities in the quartz grains by any means other than an optical examination, so more subtle indications of overgrowths or pressure shadows might be present and unnoticed.

Kerrich *et al.* (1977) estimated the temperature and grain size transitions for natural intercrystalline diffusional creep and intracrystalline dislocation creep using temperature determined by oxygen isotope thermometry and a set of optical microstructural characteristics to distinguish deformation mechanisms. For pressure solution diffusional flow, their criteria were; (1) presence of cleavage shadows, (2) pressure shadows, (3) mass transfer in microbuckles, (4) stylolites and (5) volume loss in cusp structures. Dislocation flow was identified by the presence of undulatory extinction, deformation bands and recrystallization. They estimated from those qualitative optical observational criteria that the transition between those two regimes occurred at $\sim 100\,\mu m$ for a temperature of 450°C. The stress, strain rate and pressure at which the deformation took place were effectively unknown, although the pressure was estimated to have been 110 MPa in one area and 350 MPa in a second region from the estimated thermal gradient.

The data presented here (Fig. 4) argue for grain-size dependence of strength for grain size up to $\sim 250\,\mu m$ in the Vermont conglomerates at temperatures which do not exceed $\sim$ 420°C. This grain size is larger than Kerrich *et al.*'s (1977) estimate by a factor of 2.5. Several things might account for the apparent discrepancy: strain rates and/or stresses may have differed in the two regions, or the partial pressure of water (or the chemical activity of some other chemical species) may have differed. Furthermore, because the scatter in Fig. 4 is rather large, the transition from grain-size sensitivity of strain to grain-size insensitivity of strain could remain unrecognized and we may have overestimated the upper bound of transition grain size. On the other hand, Kerrich *et al.*'s (1977) criteria would tend to underestimate the grain size at the transition since grain-size sensitivity of strength may be present even with limited dislocation flow. We emphasize, however, that both this study and Kerrich *et al*'s indicated that grain-size sensitive flow is important in the temperature regime $\sim$ 350–450°C and at grain sizes of up to 100–300 μm.

CONCLUSIONS

The principal conclusions of this study are as follows:

(1) We have verified quantitatively Thompson's (1950) observation that the strain of the pebbles correlated to the grain size within the pebbles, and we hypothesize that a grain-size sensitive deformation mechanism was responsible.

(2) The grain size within the clasts seems to be correlated in a general way with the mean free path between the second phase particles (λ), even though the λ is apparently much larger than the mean intercept length.

(3) The dislocation microstructure indicates a recovered structure which is not in equilibrium with the grain size if current palaeostress measures are correct to within an order of magnitude. If the palaeostress measures are accurate, the dislocation structure must have been overprinted by a later high stress pulse; whereas, if they are in error by a factor of 10 or more, the dislocation substructure might have been produced concurrently with the GSS deformation mechanism.

(4) The evidence at hand from this study and from other published work seems to indicate that grain-size sensitive mechanisms are capable to operating at geologic strain rates for quartzites in the grain size range of ~ 100–$300\,\mu m$ at temperatures of 370–420°C.

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