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SHORT NOTE

On the influence of porosity on the low-temperature brittle–ductile transition in siliciclastic rocks

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Abstract—A compilation of available experimental data, coupled with new results on Tennessee sandstone, shows that the low-temperature brittle faulting to cataclastic flow transition in siliciclastic rocks takes place at progressively higher confining pressures as porosity is reduced. The collapse of initial porosity compensates for the tendency for brittle deformation to be dilatant. According to a stability criterion, this in turn favours spreading of the cataclastic deformation throughout the rock volume instead of fault localization. At sufficiently high strains, dilatation, and hence fault localization, supervenes. From microstructural observations on Oughtibridge Ganister (7% porosity), deformed under conditions of the faulting to flow transition, the mechanism of pore collapse involves crushing and sliding on shear-oriented grain boundaries, with accumulation of wear products in the pore spaces. This explains the enhancement of pore collapse by differential stress relative to purely hydrostatic compression.

INTRODUCTION

It has progressively become clear that the low-temperature ductility of rocks by cataclastic flow (distributed microcracking on the grain-scale) depends upon the presence of porosity (Schock *et al.* 1973, Shimada 1986, Hirth & Tullis 1989, Evans *et al.* 1990, Wong 1990, Zhang *et al.* 1990), except (a) where a weak second phase is present (Byerlee 1968) or (b) where limited intracrystalline plasticity is possible (Fredrich *et al.* 1989). Hirth & Tullis (1989) showed that low-porosity (0.6%) Heavitree quartzite failed by cataclastic fault localization at all confining pressures up to 1000 MPa, whereas the higher porosity Oughtibridge Ganister (7.0%) shows a transition to cataclastic flow at about 600 MPa (Hadizadeh & Rutter 1983, Hirth & Tullis 1989). Other, more porous siliciclastic rocks become ductile at confining pressures as low as 100–200 MPa (Handin *et al.* 1963, Donath & Fruth 1971, Edmond & Paterson 1972, Hoshino *et al.* 1972). By preloading Kayenta sandstone (initial porosity 17%) under different hydrostatic pressures so that different porosities were produced, Wong (1990) showed that the brittle–ductile transition pressure increased systematically with decreasing porosity.

In this note, we demonstrate the effect of porosity on the brittle–ductile transition for different siliciclastic rocks and, with the aid of new microstructural observations, we discuss qualitatively the mechanism of porosity elimination during ductile flow.

THE POROSITY–TRANSITION PRESSURE RELATIONSHIP

By reference to experiments on Gosford sandstone (Edmond & Paterson 1972), Fig. 1 shows the general phenomenology of volumetric strain during deformation of brittle, porous rocks. The initial part of the inelastic strain history is marked by a substantial volume decrease, much in excess of the elastic volume decrease which arises from the loading, but always rather less than the initial porosity of the material. Local dilatancy is always associated with cataclastic deformation, so that as the rate of compaction decreases with strain, the rate of dilatation eventually supervenes.

Edmond & Paterson (1972) showed that because such substantial volume changes are associated with large-strain cataclastic deformation, the rate at which the external forces do work on the specimen with respect to the axial strain (dW/de_a) is a better description of the resistance to deformation than the differential stress

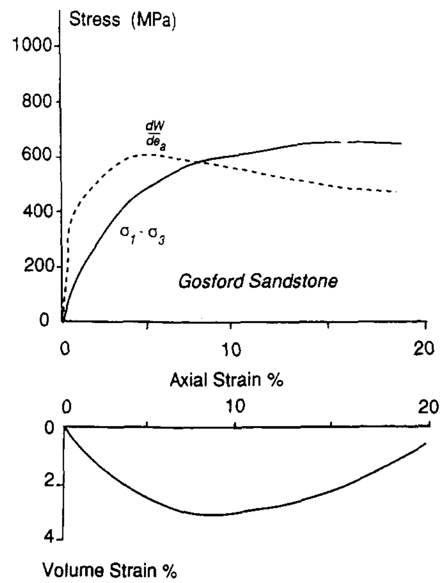

Fig. 1. Phenomenology of cataclastic flow in a porous sandstone (Gosford sandstone, initial porosity 13%, confining pressure 400 MPa) (after Edmond & Paterson 1972). The upper graph shows curves of both differential stress (solid line) and rate of work (broken line) against axial strain. The lower curve shows volumetric strain. In each case the initial period of pore compaction coincides with macroscopically ductile behaviour. When net dilatation supervenes, localized deformation becomes favoured.

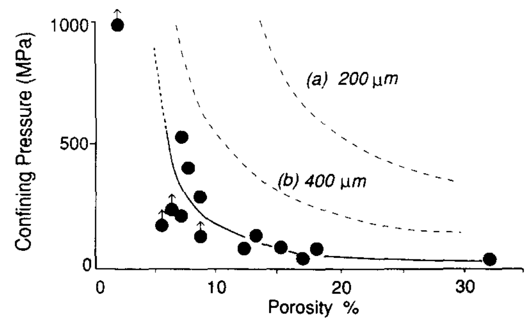

Fig. 2. Assembly of data (see Table 1) to show how the confining pressure at the brittle–ductile transition increases as porosity is reduced. No attempt is made to take into account differences in grain-size, cementation, etc., between the different rock types. The trend in the data is indicated (solid line). Points with upward-pointing arrows are highest test pressures for those rock types, but ductility was not yet attained. Curves (a) and (b) are calculated from the theory of Zhang *et al.* (1990), as the hydrostatic pressure required to cause growth of Hertzian indentation cracks in an aggregate of spherical grains of maximum size 200 and 400 μm, respectively. These curves grossly overestimate the brittle–ductile transition curve, even if mean stress were to be plotted instead of confining pressure.

(Fig. 1). The differential stress curve shows apparent strain-hardening to higher strains than the dW/de_a curve, and the intersection of the dW/de_a and $(\sigma_1 - \sigma_3)$ curves occurs when $de_v/de_a = 0$. They suggested that the condition $d^2W/de_a^2 < 0$ is a criterion for instability leading to localization of the flow. Hobbs *et al.* (1990) showed that for the case of axisymmetric deformation in a triaxial test, this criterion is approximately equivalent to the Hill (1958) stability criterion ($\dot{\sigma} \cdot \dot{e} > 0$).

The availability of collapsable pore volume therefore appears to be largely responsible for the stability of cataclastic flow, up to the point at which net dilatation supervenes. From numerous experimental studies, cataclastic flow is a transient phenomenon, which usually ends after 10–30% axial strain with the formation of a fault zone transecting the deformed material (e.g. Hadizadeh & Rutter 1983). It is not safe to conclude, however, that in experiments the termination of the flow by faulting arises from the evolution of the material behaviour alone. Faulting may be forced through inhomogeneous flow at the corners of the specimen, which arises from specimen/piston friction and spreading of the specimen beyond the corners of the pistons (e.g. fig. 7a of Hadizadeh & Rutter 1983).

In Table 1 and Fig. 2 we assemble data which show how the faulting to cataclastic flow transition depends upon porosity. The sensitivity is most marked below about 12% initial porosity, therefore we have performed additional experiments on Tennessee sandstone (7.5% total porosity) using the same experimental methods as Hadizadeh & Rutter (1983), in order to find the brittle–ductile transition pressure (Table 1). This assembly of data strongly supports the view that collapsible porosity is essential to cataclastic flow on the grain-scale, and demonstrates the striking control that porosity exerts.

DISCUSSION

In recent years substantial progress has been made in the modelling of cataclastic deformation in which elements of 'real' microstructural evolution have been explicitly included (e.g. Nemat-Nasser & Horii 1982, Costin 1983, 1986, Sammis & Ashby 1986, Ashby & Hallam 1986, Zhang *et al.* 1990, Casey & Wust 1990, Wong 1990). We will attempt to focus upon certain aspects which appear to be relevant to the problem of cataclastic flow which explicitly involves pore collapse.

(a) *The role of differential stress in facilitating pore collapse*

Zhang *et al.* (1990) reported experimental data on the collapse of porosity in sandstone under hydrostatic stress. They argued that the onset of compaction corresponded to conditions for the growth of impingement cracks at highly stressed grain contacts, and carried out a fracture-mechanics analysis in terms of the formation of well-behaved 'Hertzian' cracks under a spherical indenter. They showed that this approach predicted behaviour which was consistent with experimental observations, at least on initially relatively porous materials.

In Fig. 2 we show curves constructed from the theoretical model of Zhang *et al.* (1990), which represent the pressure for the onset of crushing vs porosity. There is a grain-size effect, finer grained rocks requiring higher crushing pressures for a given porosity. The crushing pressure curves depend on porosity in much the same way as the brittle–ductile transition. In the latter instance, however, there is no apparent grain-size effect, but this could be an artifact of the limited range of grain-sizes represented, particularly for the lower porosity

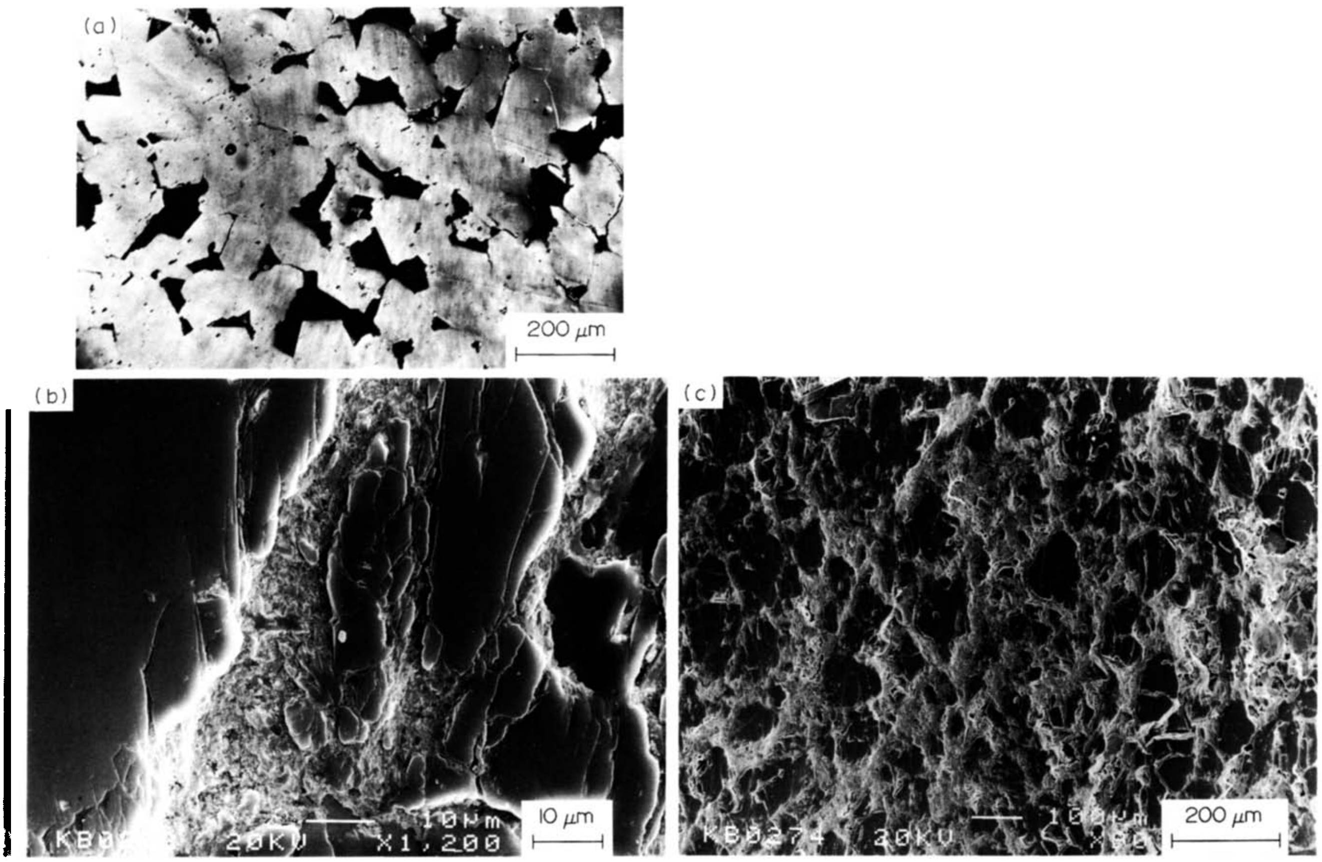

Fig. 3. Micrographs of polished surfaces of Oughtibridge Ganister. (a) Optical micrograph (reflected light, undeformed) showing the porosity structure (black). Pore edges are angular because they are bounded by crystal faces of the quartz overgrowth cement. The grain size is about 110 μm, and the long edge of the photograph is about 0.9 mm. (b) & (c) are scanning electron micrographs of a specimen deformed at 500 MPa confining pressure (the transition pressure), 20°C and at a strain rate of $10^{-5}\ s^{-1}$. Compression direction is parallel to the short side of each photograph. (b) Failure of a shear-oriented grain boundary, involving the formation of intensely crushed material in the interface, through the formation and coalescence of short axial and other indentation cracks which propagate into the grain. Scale bar 10 μm. (c) Lower magnification image (scale bar 200 μm) showing the formation of a dispersion of conjugate arrays of shear failure and crushing of shear-oriented grain boundaries. This accommodates collapse of intergranular porosity and accumulation of wear products in the pore spaces, the remnants of some of which can still be discerned.

Table 1. Brittle–ductile transition data for (mainly) siliciclastic rocks (all stresses in MPa, all rocks tested air or oven dry at a strain rate of about 10^{-5}–10^{-4} s^{-1})

Rock type	Porosity (%)	Transition pressure	Ultimate strength	Source
Oughtibridge Ganister	7	550	1700	Hadizadeh & Rutter (1983)
Porous basalt	7	230	700	Shimada (1981)
Lance sst	8.5	300	750	Schock *et al.* (1973)
Berea sst	18	100	240	Handin *et al.* (1963)
Bunter sst	15	100	350	Gowd & Rummel (1980)
Gosford sst	13	150	250	Edmond & Paterson (1972)
Sandstone	12	100	—	Ismail & Murrell (1976)
Tennessee sst	7.5	400	1250	This study
Heavitree quartzite	0.6	>1000	>3700	Hirth & Tullis (1989)
Kayenta sst	17	50	—	Wong (1990)
Maze sst	6	>220	530	Hoshino *et al.* (1972)
Ohmagari sst	5.3	>190	480	Hoshino *et al.* (1972)
Nanatani sst	8.6	>140	450	Hoshino *et al.* (1972)
Hiwaki sst	30	~10	~20	Hoshino *et al.* (1972)

rocks. From Fig. 2 it is clear that, even taking into account the contribution of differential stress to the mean stress on the sample, the 'well-behaved' Hertzian indentation model for pore collapse requires effective hydrostatic stress levels rather higher than those associated with cataclastic flow. This point was also made by Hirth & Tullis (1989) with particular reference to Oughtibridge Ganister.

(b) *Mechanism of pore collapse*

Hirth & Tullis (1989) interpreted the mechanism of pore collapse in Oughtibridge Ganister to be analogous with the collapse of a void in plaster of Paris subjected to uniaxial loading. This occurs owing to the high compressive stress induced around the void in the plane normal to the applied stress (Lajtai & Lajtai 1975). In plaster of Paris such collapse occurs through the formation of intersecting conjugate shear surfaces adjacent to the void, but from our observations (Fig. 3) this is not what happens in Oughtibridge Ganister (although a certain amount of tensile spalling akin to the formation of borehole breakouts may occur).

In addition to the formation of axial transgranular cracks emitted from pore corners and grain impingements, grain-boundary cracks form readily, owing to lower fracture toughness along such surfaces (Hadizadeh & Rutter 1982), and are also initiated at pore corners. Frictional sliding occurs on shear-oriented cracked grain boundaries, causing cataclastic damage to the grain-boundary region (Fig. 3). Spalled fragments are forced into the pore spaces and the change in the effective grain-shape facilitates pore collapse through the sliding motions of the grains. Attention was drawn to this grain-boundary shearing damage by Hadizadeh & Rutter (1983, fig. 7b), but the new SEM photographs illustrate the process rather better. Such damage would not be induced during purely hydrostatic compaction, and this accounts for the disproportionate efficacy of the differential stress in aiding pore collapse.

The sliding of one grain boundary over another induces asymmetric indentation-microcracking (Hamilton & Goodman 1966, Lawn & Wilshaw 1975), which leads to the formation of wear products as microcracks interact and curve into each other (Fig. 3c). Ashby & Hallam (1986) and Sammis & Ashby (1986) showed that axial growth of symmetric axial cracks is strongly inhibited by confining pressure. However, Hamilton & Goodman (1966) showed that stress intensity at the tip of a crack under a *sliding* indenter could be magnified *ca* 20 times relative to that at the tip of a crack of the same length under an indenter subjected only to normal load. This may help to explain why grain-boundary sliding, with attendant damage, may dominate over transgranular axial cracking as confining pressure is increased, provided sufficient porosity is available to accommodate the consequences of the sliding.

The behaviour of small parts of the complex structure of a rough grain boundary, in which local frictional forces vary from point to point, may be considered analogous to the geometry in which tensile 'wing cracks' form from the ends of a shear-oriented crack, such as was analysed by Nemat-Nasser & Horii (1982) and Ashby & Hallam (1986). The large numbers of observed short 'wing' cracks, such as illustrated on Fig. 3(c), may provide the level of damage required to pulverize the boundary. Comparison of figs 14, 15 and 16 in Ashby & Hallam (1986) with fig. 18 in Sammis & Ashby (1986) shows how, all other factors remaining constant, smaller applied stresses can drive a 'wing' crack to a given length compared to a symmetric axial crack propagating from a pore.

SUMMARY AND CONCLUSIONS

The cataclastic faulting to flow transition in thoroughly brittle quartzites depends upon the rock having sufficient porosity to allow deformation to be accompanied by a net volume decrease. With continued strain, dilatancy supervenes and fault localization occurs, but in experiments to high strains faulting may be less a material characteristic than a product of specimen–machine interaction.

Available experimental data have been gathered together to show the marked sensitivity of the faulting to flow transition pressure to rock porosity. Microstructural studies of Oughtibridge Ganister at the faulting-flow transition show that pore collapse occurs through crushing and sliding of shear-oriented grain boundaries. This provides a qualitative explanation for the facilitation of pore collapse under differential stress compared with hydrostatic stress of the same mean value.

We are unable to account satisfactorily for the form of the porosity vs transition pressure relationship, but suspect that the geometric intensification of applied stress around voids and through area reduction at grain boundaries in porous rocks, facilitating grain-boundary sliding at lower pressures, provides a partial and qualitative explanation for the enhanced ductility of more porous rocks.

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