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Microstructure of deformation bands in porous sandstones at Arches National Park, Utah

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Abstract—At Arches National Park it is possible to distinguish three kinds of deformation bands on the basis of their distinctive microstructure: (1) deformation bands with little or no cataclasis; (2) deformation bands with cataclasis; and (3) deformation bands with clay smearing.

The micromechanics of deformation band development consist of initial dilatancy followed by grain crushing and compaction. This process may be developed to different stages according to the interplay of porosity, confining pressure, clay content and amount of strain. Low porosities and low confining pressures promote the formation of dilatant bands with no cataclasis. High porosities and high confining pressures promote compaction and cataclasis.

Two generations of deformation bands were documented. The older generation has little or no cataclasis and formed in relatively undisturbed sandstone probably under conditions of low confining pressure. The younger generation exhibits cataclasis, appears to be localized in proximity to major faults and seems to have developed under conditions of high confining pressure. The temporal sequence of deformation band development can be related to the regional geology of the area; where the first generation probably formed during growth of the salt anticline, and the second generation during its collapse.

INTRODUCTION

THE subject of this paper is the microstructure of deformation bands in porous sandstones and the identification of the factors relevant to their development. We have used field mapping, optical microscopy and video image analysis to study deformation bands microstructure in different sandstone units as well as different kinds of deformation bands microstructures within the same sandstone unit.

At Arches National Park in southeastern Utah (Fig. 1) deformation bands occur in several different sandstone units near major faults and folds or isolated in relatively undeformed sandstone. We observed and sampled deformation bands in the Triassic Chinle, Wingate and Kayenta Formations, in the Triassic–Jurassic Navajo Sandstone, in the Dewey Bridge, Slickrock and Moab members of the Jurassic Entrada Sandstone, in the Brushy Basin member of the Jurassic Morrison Formation and in the sandstone interlayers of the Cretaceous Mancos Shale. At Arches National Park, faulting developed in detrital rocks of Mesozoic age as a result of growth and subsequent collapse of a NW–SE-trending salt anticline (Doelling 1985). Units younger than the Entrada Sandstone have uniform thicknesses over the area and therefore their deposition does not seem to have been affected by salt diapirism. Salt movement during growth of the anticline has been recorded from Permian to Jurassic time. Salt dissolution and collapse started later and is probably ongoing (Doelling 1985).

Deformation bands (Aydin 1977, Underhill & Woodcock 1987) also called shear fractures (Dunn *et al.* 1973), granulation seams (Heald 1956, Bevan 1985), Lüder bands (Friedmann & Logan 1973, Bahat 1991) or band faults (Cruikshank *et al.* 1991, Zhao & Johnson 1991) are small 'fault-like' structures that develop in porous granular materials. According to Aydin (1977, 1978) and Aydin & Johnson (1978) deformation bands are roughly planar features that appear as thin bands from a few centimeters to a few tens of meters in length. Each band is about 1 mm thick and accommodates a small offset on the order of a few mm to a few cm (Fig. 2). At Arches National Park the strains have been accommodated via dip-slip, oblique-slip and strike-slip deformation bands that formed in different periods of salt movement. Although the bands are approximately planar on the scale of their length, smaller-scale irregularities, forming characteristic eye and ramp structures are common (Fig. 2). The grain-scale characteristics of deformation bands include a thin layer of fine crushed grains and a distinct porosity reduction relative to the host rock. The grain crushing and porosity reduction are caused by cataclasis involving grain breakage and rotation.

Although deformation bands accommodate small offsets, they do not exhibit a well-defined plane of displacement discontinuity. The gradient of displacement within a deformation band is steep but more or less continuous. The large values of shear strain (1–10) across a band, their small thickness and the discontinuous nature of the sandstone grains perhaps can give the impression of a

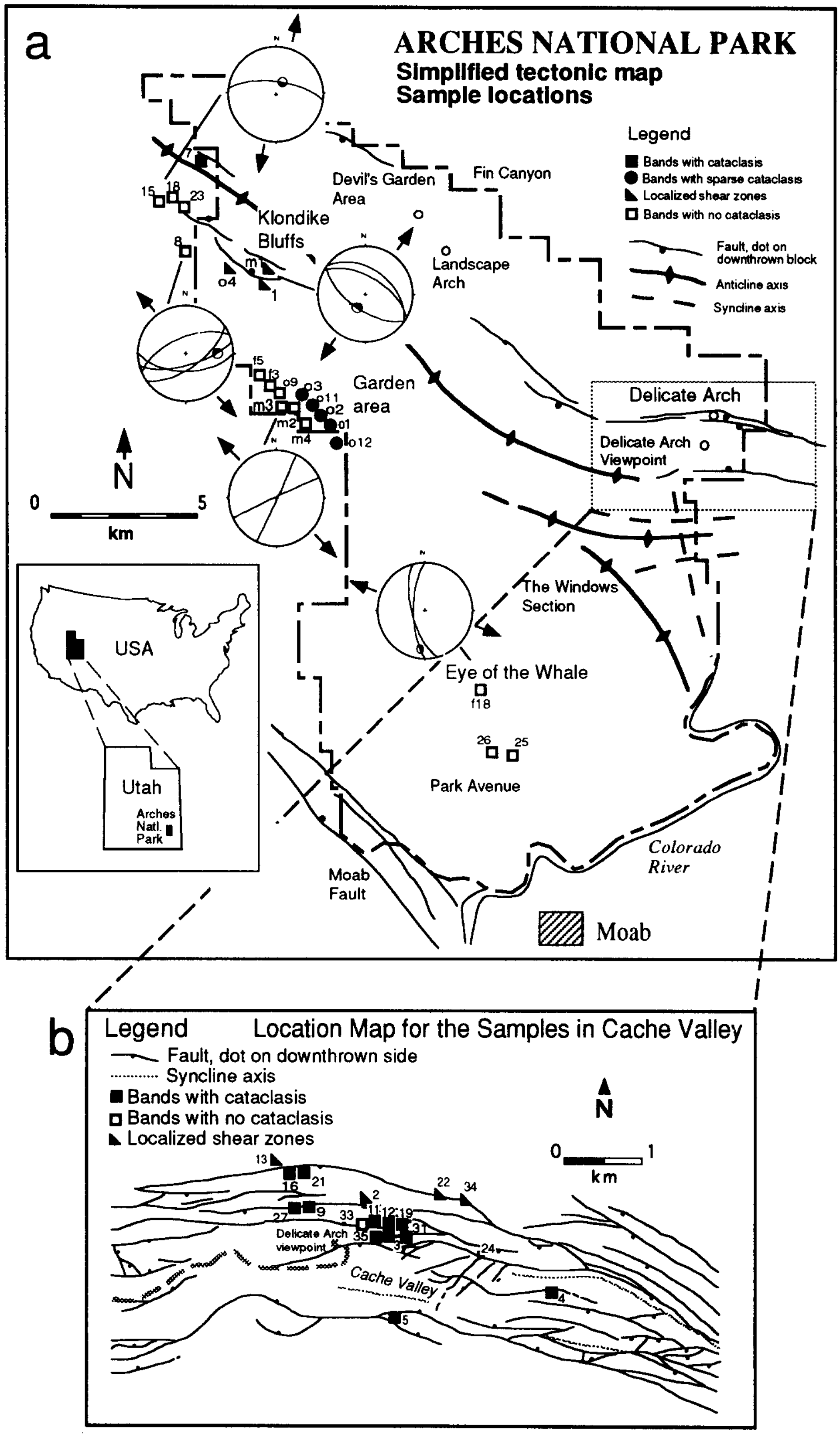

Fig. 1. (a) Schematic map of Arches National Park containing major structural features and selected sample localities. Note the distribution of bands with cataclasis along the anticlinal axis and of bands with no or little cataclasis at greater distances away from the anticlinal axis. The trends of the principal sets of deformation bands have been plotted on equal-area stereonets, together with the direction of net slip represented by the dot (empty half of the dot is on the down-thrown side). The principal direction of extension is indicated by the arrows. (b) The smaller map represents the major faults in Cache Valley (after Doelling 1985) and selected sample localities.

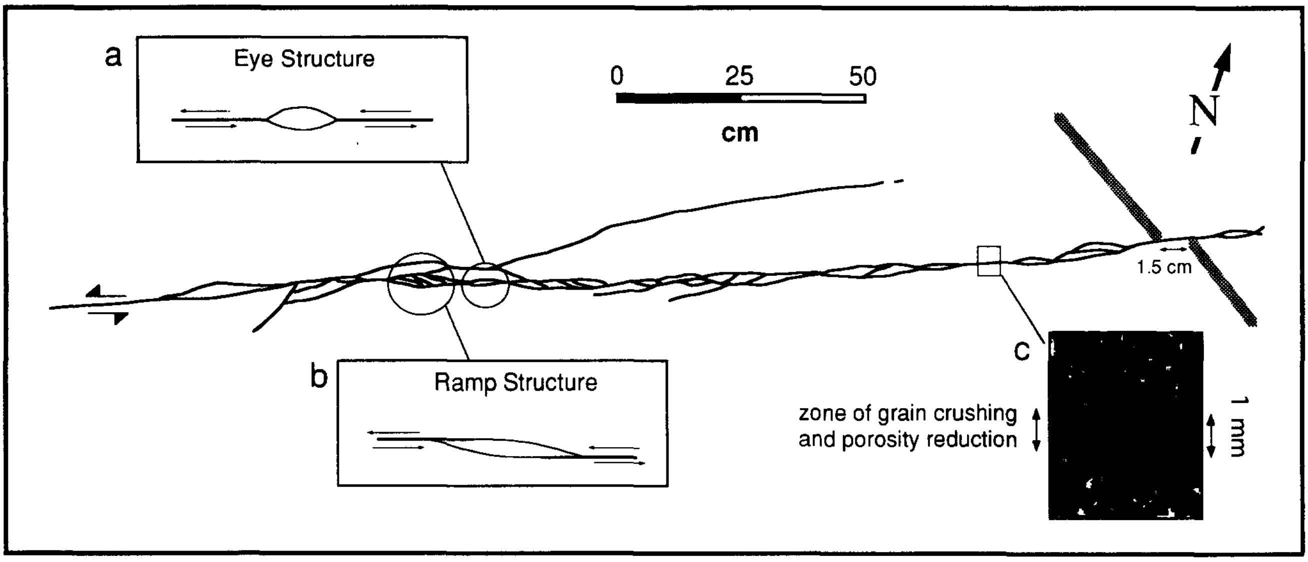

Fig. 2. Outcrop map of a deformation band from the Garden area (Arches National Park). (a) Eye structures where two segments are almost coplanar, (b) ramp structures where segments are offset from one another and (c) single band segment as a zone of grain crushing and compaction.

discontinuity. Aydin (1977, 1978) and Aydin & Johnson (1978) have described how deformation bands tend to cluster in 'zones of deformation bands' up to 40–50 cm in thickness. The zones of deformation bands can accommodate a few dm of offset; further offset is usually accommodated by the development of a slip plane that does represent a displacement discontinuity within the sandstone.

In the literature concerning triaxial experiments the term 'localization' describes the development of a distinct shear fracture along which the rock parts. The development of this surface usually is followed by strain softening (Rudnicki & Rice 1975). Deformation bands are a localized phenomenon common in sandstone, but they do not necessarily represent a discontinuity, nor are they necessarily associated with strain softening. Typically the rock does not part easily along the bands when a specimen is recovered from the outcrop. Their ability to accommodate only limited amounts of offset as single bands, and their tendency to occur in zones, suggest a strain hardening phenomenon (Rudnicki & Rice 1975, Aydin & Johnson 1983). Rudnicki & Rice (1975) suggest that grain crushing and compaction in a band have a stabilizing effect. This stabilizing effect may lead to the localization of other deformation bands in proximity of the first one. Thus deformation bands play an important role in deformation of porous rock masses.

The fine layer of crushed grains and the porosity reduction within a deformation band may provide a natural barrier to fluid flows. Compartmentalization of hydrocarbon reservoirs and fault sealing in porous sandstones often are controlled by these structures (Pittman 1981, Bevan 1985, Nelson 1985, Hardmann & Booth 1991). Thus, where present, deformation bands may play an important role in determining the flow behavior of reservoirs and aquifers.

THE METHOD OF STUDY

The microstructural study of the deformation bands at Arches National Park is part of a broader project that focuses on the structural and fluid flow properties of faulting in sandstone. The field work included mapping and sampling; the results of the mapping will be presented elsewhere (Antonellini *et al.* in preparation). The microstructure of the samples collected has been analyzed with standard optical techniques and image analysis. Ninety thin sections, 75 of which were impregnated with blue epoxy to enable porosity determinations, have been examined. Seventy-eight thin sections are from samples collected at or near Arches National Park (Fig. 1). To have a broader range in sandstone types and stress state conditions the thin section analysis was extended to deformation bands in four samples from the Mio-Pliocenic Pismo Formation (Central California), in three samples from the Ordovician St. Peter Sandstone of the Kentland Quarry (Indiana), in two samples from the Jurassic Aztec Sandstone of the Valley of Fire (Nevada) and in three samples from the Eocenic Loiano Formation of the Northern Apennines (Italy).

Geometric properties, such as band thickness, offset, grain size of the crushed zone and grain size of the host rock, were measured with an optical micrometer. The composition and textural characteristics were determined by optical microscopy and by point counting.

The deformation bands are so thin (1 mm) that the standard procedures of porosity determination (Bourbié *et al.* 1978) do not provide the correct estimate of void volume within and in the neighborhood of the band. Therefore, the porosity was measured in the host rock, in the deformation band, and in an intermediate zone directly on the thin section via an image analysis system supported by an RTI vision engine (Recognition Tech-

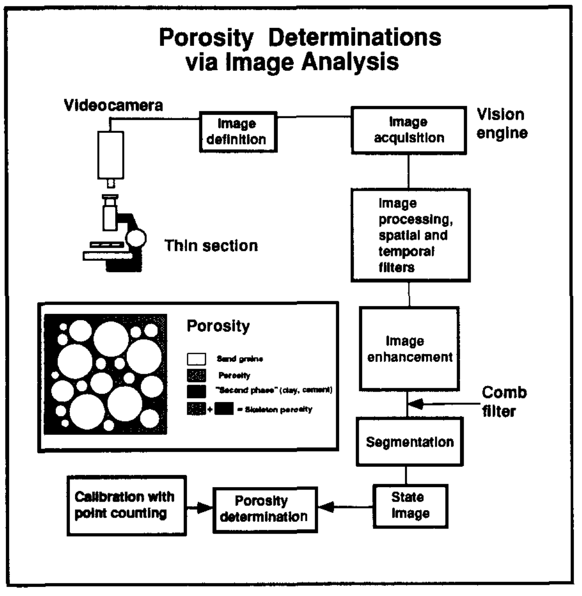

Fig. 3. Flow chart representing image processing sequence required to obtain two-dimensional porosity determinations from a thin section. The inset shows the relationship between sand grains, porosity, second phase and skeleton porosity.

nologies Inc.) and a video camera (Fig. 3). The image of a thin section is transmitted from a video camera to an image analysis system where it is processed and finally segmented to separate minerals from pore space using a comb filter on the gray levels of the pixels. This filter can separate the blue of the epoxy from the color of the minerals. Once the image has been segmented the porosity is determined by automatic pixel counting on selected areas of the image. The porosity measured with the image analyzer was checked and calibrated by direct point counting on the thin section, and by direct superposition of the segmented image on the original image of the thin section.

The definition of porosity is the ratio between the volume of voids and the volume of sample. However, in some places the pore space is filled by 'second phases' such as authigenic clays, coatings, iron oxides or cement precipitations. Typically these 'second phases' are the result of weathering and diagenesis and they were not present at the time of formation of the deformation bands. Therefore we introduce another definition of porosity that includes this second phase (Fig. 3); following Bourbié *et al.* (1987) we call this extended definition of porosity 'skeleton porosity'. The porosity of the clay cannot be estimated by any visual method, however its contribution is of lesser importance to our microstructural study.

TYPES OF DEFORMATION BANDS AND THEIR DISTRIBUTION

In the following the types of deformation bands observed in this study are described and classified (see Table 1). We base the classification of deformation bands on microstructural characteristics such as development of cataclasis and observed porosity changes with respect to the host rock. Then, we consider the distribution of deformation bands in terms of their spatial arrangement and of their position relative to major structures. The common characteristics among all the deformation bands observed are the small amounts of offset (mm to dm), the lack of a well-defined discontinuity plane and the typical connecting structures among different segments (eye and ramp structures) similar to those described by Swanson (1988) and by Cruikshank *et al.* (1991).

Microstructure and classification of deformation bands

On the basis of their microstructure and localization characteristics it is possible to define three major kinds of deformation bands (see also Table 1): (1) deformation bands with little or no cataclasis (with dilatancy, with compaction and with no volume change; Figs. 4a & b); (2) deformation bands with cataclasis (Fig. 5); and (3) deformation bands with clay smearing (localized shear zones; Figs. 6 and 7).

(1) *Deformation bands with no cataclasis.* At Arches National Park deformation bands with few or no crushed grains formed in relatively undisturbed layers of the Moab and Slickrock members of the Entrada Sandstone. These bands can be identified in thin section by the porosity and/or fabric differences that they display with respect to the surrounding host rock. They appear as: (a) bands showing positive dilatancy resulting in a porosity increase with respect to the host rock; (b) bands showing compaction (negative dilatancy) resulting in a porosity decrease with respect to the host rock; and (c) in the case of no detectable change in volume, as an area with a distinctive fabric defined by the alignment of elongate grains (grains that tend to have their longer axis in the plane of the deformation band). These bands are usually more distinct in outcrop than in thin section. Cement, authigenic clays and/or iron oxide precipitation is common in this kind of band, especially where the bands are associated with positive dilatancy. These second phases are not the result of weathering of crushed grains in the deformation bands because no evidence of grain cracking or cataclasis is observed and because these phases are also present in joints.

The thickness of these deformation bands is directly proportional to the mean grain size of the host rock (Fig. 8a). A linear relation was observed by Roscoe (1970) in experiments on loosely consolidated granular materials; in particular he found that the thickness of shear bands was approximately 10 times the average grain diameter. Mühlhaus & Vardoulakis (1988) used bifurcation analysis (Rudnicki & Rice 1975, Vardoulakis 1980, 1981, 1983) and Cosserat's theory (Coleman & Hodgon 1985, Triantafyllidis & Aifantis 1986) to show that, for a given shear strain, the thickness of a band should be a small multiple (5–10) of the grain size (diagram p. 282 of Mühlhaus & Vardoulakis 1988). Laboratory experiments on a medium-grained sand by Vardoulakis *et al.* (1985) show that the band thickness is about 16 times the

Table 1. Classification of deformation bands

Major groups	Subgroups
1. Deformation bands with no cataclasis	1a. Deformation bands with dilatancy
	1b. Deformation bands with no volume change
	1c. Deformation bands with compaction
2. Deformation bands with cataclasis	2a. Deformation bands with poorly-developed cataclasis
	2b. Deformation bands with well-developed cataclasis
3. Deformation bands with clay smearing (localized shear zones)	3a. Deformation bands with clay smearing in high porosity (>20%) and low clay content (~5–7%) host rocks—with cataclasis
	3b. Deformation bands with clay smearing in low porosity (<15%) and high clay content (~15%) host rocks—with poorly developed cataclasis

average grain diameter for a shear strain of 0.1. The difference between the experimental result of Vardoulakis *et al.* (1985) and that observed for the bands at Arches (Fig. 8a) is probably caused by lower strain in the sandstones compared to that in the samples in the experiments (Vardoulakis *et al.* 1985) and to that modeled by Mühlhaus & Vardoulakis (1988, diagram p. 282).

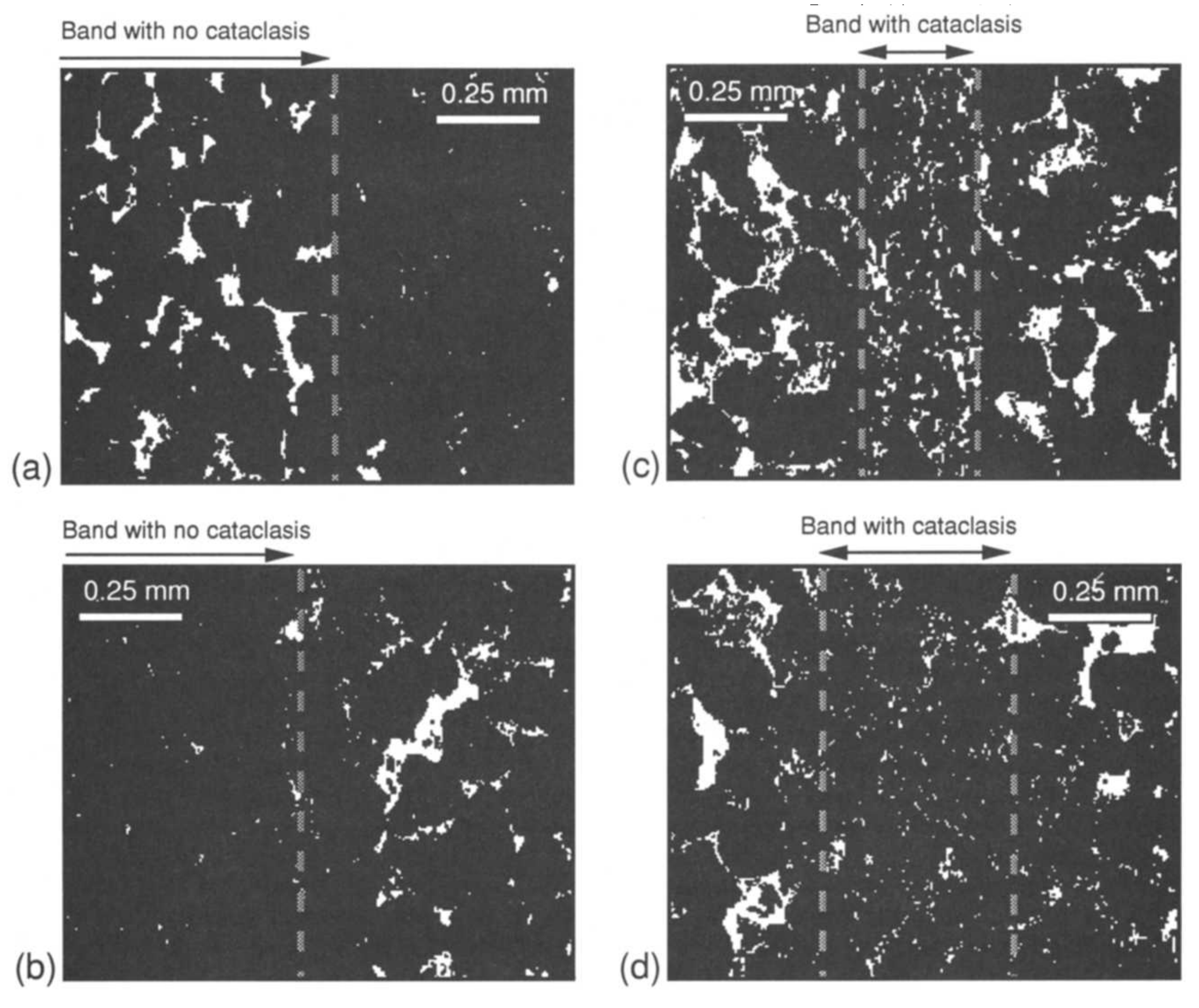

Fig. 4. Segmented images of typical deformation bands from black and white video frames. White represents pore space, black represents solid phases (mineral grains, clay, coatings). (a) Deformation band with no cataclasis showing dilation in sample m4 of the Moab member of the Entrada Sandstone in the Garden area. (b) Deformation band with no cataclasis showing compaction in sample f12 of the Moab member of the Entrada Sandstone collected in the Garden area. (c) Deformation band with cataclasis in sample 3c of the Moab member of the Entrada Sandstone in the Delicate Arch area. (d) Deformation band with cataclasis in sample o1 at the contact between Moab and Slickrock members of the Entrada Sandstone.

(1a) *Deformation bands showing positive dilatancy.* Deformation bands showing positive dilatancy (Table 1), but few or no crushed grains, have been observed in the Garden, Little Valley and Klondike Bluffs area at Arches National Park. These bands cluster in thin (0.5–3 cm) zones accommodating a 1–5 cm offset. The zones are distributed over wide areas (several hundreds meters to a few km) and have a spacing ranging from 10 to 50 m. These deformation bands show a higher porosity than the host rock (Figs. 4a and 9a). For cases where the porosity in the host rock is 4–5%, the porosity in the band is up to 12–13%. Positive dilatancy in the bands is usually observed where the porosity in the host rock was less than 12%, but in a few cases host rock porosity was up to 18–22%. The skeleton porosity is most effective in documenting the increase in porosity related to the original deformation within the band, because it does not include authigenic clays and iron oxides precipitated preferentially in the pore space within the band (cf. Figs. 9a & b).

Other characteristics of the sandstone in which these bands form include tight packing, generally good sorting and low clay content (<1–2%). The amount of polycrystalline quartz within the band is the same as in the host rock and the number of feldspar grains with a small size in the band is also comparable to that of the host rock. In some samples it is possible to recognize an increase in grain cracking within the band with respect to the host rock, but only occasionally can one see crushed grains. The offset measured on these bands ranges from several mm to 3–4 cm. An exception to this observation is represented by the deformation bands examined in the Valley of Fire that have little or no appreciable offset (<1 mm; Hill 1989).

Bands showing positive dilatancy usually do not weather out in positive relief unless the increased pore space within a band is filled by mineral precipitate. The pore filling of the deformation band often consists of authigenic clays and/or iron oxides precipitated by circulating fluids. These minerals give a characteristic brown or red color to the deformation bands in outcrop.

The absence of crushed grains together with the observed positive dilatancy suggest that these deformation bands formed under conditions of low mean compressive normal stress. Later we will discuss this point making comparisons with triaxial lab experiments.

(1b) *Deformation bands showing compaction.* Deformation bands showing compaction, but little or no cataclasis (Table 1), have been observed in the Garden, Delicate Arch and Eye of the Whale areas at Arches National Park. Their spatial distribution is similar to that of bands showing positive dilatancy. These bands are marked by decreased porosity with respect to the host rock (Fig. 4b). The porosity in the host rock is usually larger than 18% (Figs. 9c & d) and ranges from 9 to 29%. Within the band the porosity drops to a value ranging from 1 to 18%.

Other characteristics of the sandstone containing these bands include loose packing and variable sorting (from very poor to good). The amount of polycrystalline quartz and of small-grain-size feldspars within the band is the same as in the host rocks. The clay content is less than 2–3%, except for the samples in the Slickrock where it can be as high as 7–8%. These bands have offsets up to 3–4 cm, but generally on the order of a few mm. In some bands it is possible to recognize a slight increase in grain cracking at contact points between grains, but seldom does one observe crushed grains. In some places one can identify a distinct fabric in the deformation band because of the alignment of grains with an oblate shape within the plane of the band.

(1c) *Deformation bands with little or no volume change.* These deformation bands do not exhibit cataclasis nor do they display any appreciable volume change (Table 1). They can be recognized by the alignment of grains with an oblate shape within the plane of the band. Such bands were recognized in a few samples collected from the Dewey Bridge member of the Entrada Sandstone at Arches National Park (samples 25, 26, 33 in Fig. 10) and in the Wingate sandstones at Canyonlands National Park. In the deformation bands of this latter unit, however, there is also a very small amount of cataclasis (samples 29 and 30 in Fig. 10).

In these bands the porosity is equivalent to the porosity of the host rock (1–5%). The packing varies from medium to tight; the sorting in the host rock is very poor and the clay content is high (5–15%). These bands are usually associated with conspicuous amounts of offset (on the order of several cm). The amount of polycrystalline quartz and small-grain-size feldspars is about the same in the band as in the host rock.

(2) *Deformation bands with cataclasis.* Bands with considerable cataclasis have been observed in the Chinle, Wingate, Kayenta and Navajo Formations, in the Moab and Slickrock members of the Entrada Sandstone, in the Morrison and Dakota Formations (Fig. 10) at Arches National Park. These bands are localized in proximity to major faults with offsets ranging from 5 to 1500 m. They tend to cluster in zones 10–30 cm thick that may be distributed over several meters or tens of meters near the fault zone.

A characteristic of these deformation bands is the presence of grain crushing and porosity reduction (Figs. 4c & d). They are similar to deformation bands described by Aydin (1977, 1978) and Aydin & Johnson (1978) in the San Raphael Desert (Utah). The amount of cataclasis is variable from a few patches of crushed grains in pods and pockets aligned along a line (Fig. 5a) to a fully developed cataclastic zone about 1 mm thick (Fig. 5b). The amount of cataclasis is greater in bands closer to slip planes, or in zones near slip planes where the host rock has been compacted (Fig. 5c). However, there is no apparent relation between the amount of offset and the degree of grain crushing as measured by Engelder (1974) for fault gouge in the Precambrian sandstones of the Uinta Mountain Group (Utah).

Deformation bands with cataclasis have a thickness

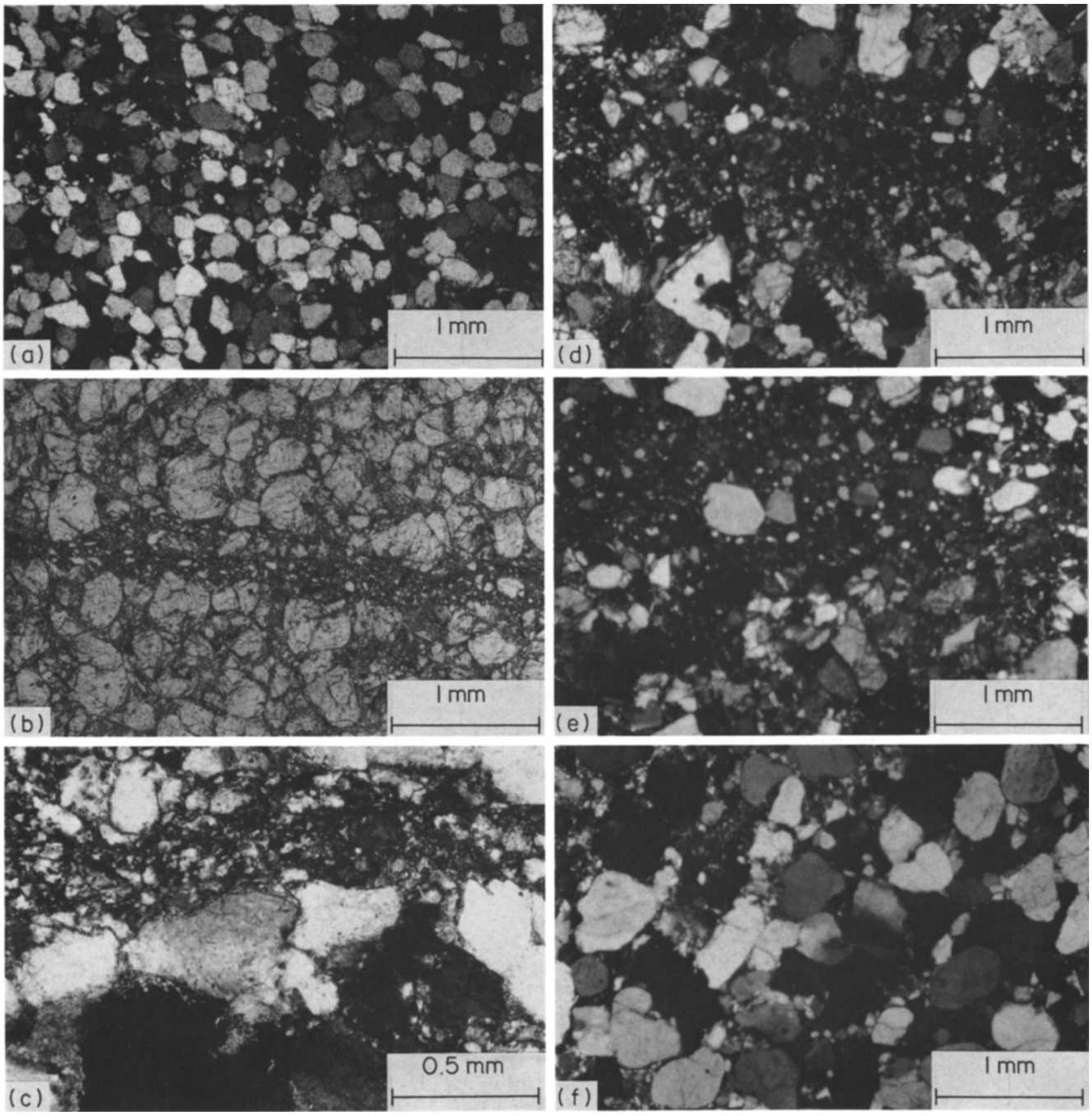

Fig. 5. (a) Deformation band with very sparse cataclasis at the contact between Moab and Slickrock members of the Entrada Sandstone in the Garden area (Arches National Park). (b) Deformation band with well-developed cataclasis in the St. Peter Sandstone (Indiana). Note the intensity of grain cracking in the host rock and the small width (0.2 mm) of the band. (c) Deformation band with well-developed cataclasis in the Moab member of the Entrada Sandstone at Delicate Arch. The band formed in compacted host rock (5% porosity) between two slip planes. The band is less than 1 mm thick. The change in grain size between band and host rock is very distinct. (d) Deformation band with fine cataclasis in the Loiano Formation, Northern Apennines (Italy). The sandstone is very rich in feldspars and the intensity of grain cracking in the host rock is high. (e) Deformation band with very fine cataclasis in the Loiano Formation, Northern Apennines (Italy). Note that a few of the larger grains are enclosed within the finer cataclasite and remain relatively uncracked. (f) Grain cracking at points of contact between grains in a deformation band in the Aztec Formation of the Valley of Fire (Nevada).

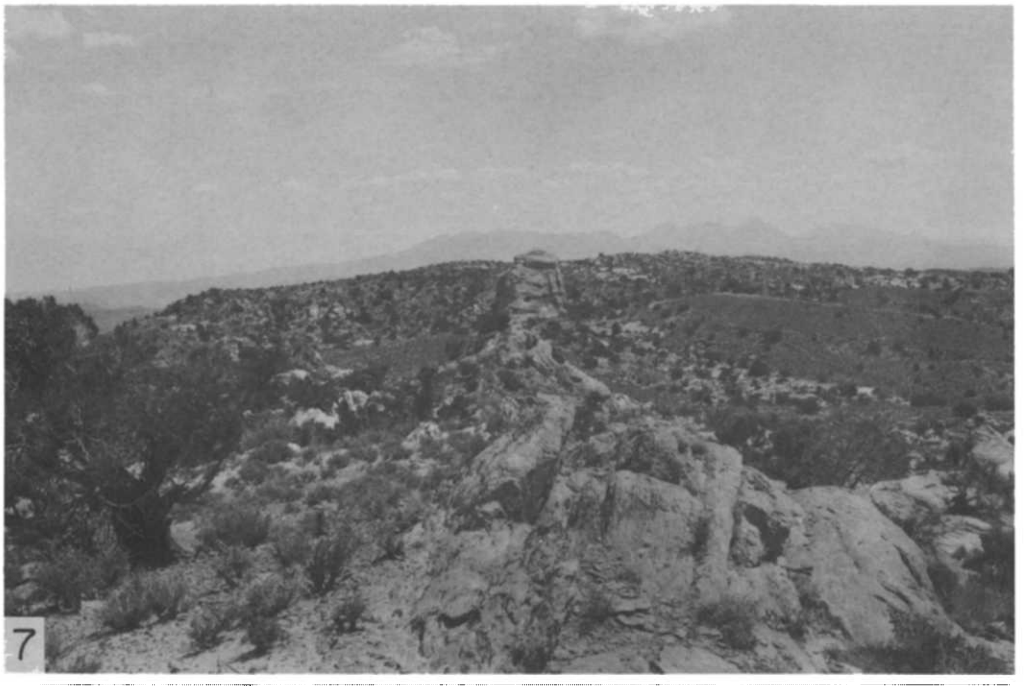

Fig. 6. Localized shear zone in the Navajo Sandstone at Klondike Bluffs. These zones appear as colorful bundles of deformation bands with color ranging from white to orange and purple. The thin colorful cross beds in the Navajo sandstone are sheared and stretched along these localized shear zones.

Fig. 7. Localized shear zone (~3 m thick) at Klondike Bluffs that accommodates an offset on the order of 10 m. No slip planes are present. This zone is also detectable from aerial photographs.

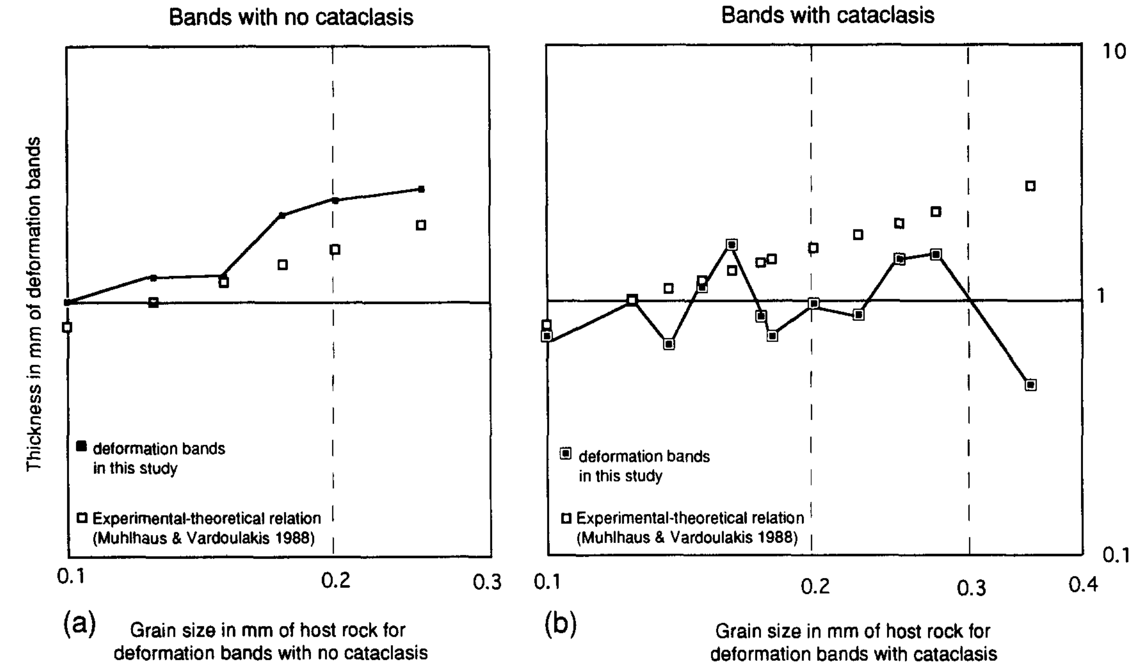

Fig. 8. Relation between grain diameter and band thickness in log–log plot. The empty squares represent the experimental relation observed by Vardoulakis *et al.* (1985) in an unconsolidated granular material. The black squares represent the mean of measurements made on deformation bands grouped by average grain diameters. (a) Deformation bands with no cataclasis in samples from the Garden, Little Valley and Klondike Bluffs area, (b) deformation bands with cataclasis in samples from the Delicate Arch, North Salt Valley, Garden and Klondike Bluffs area.

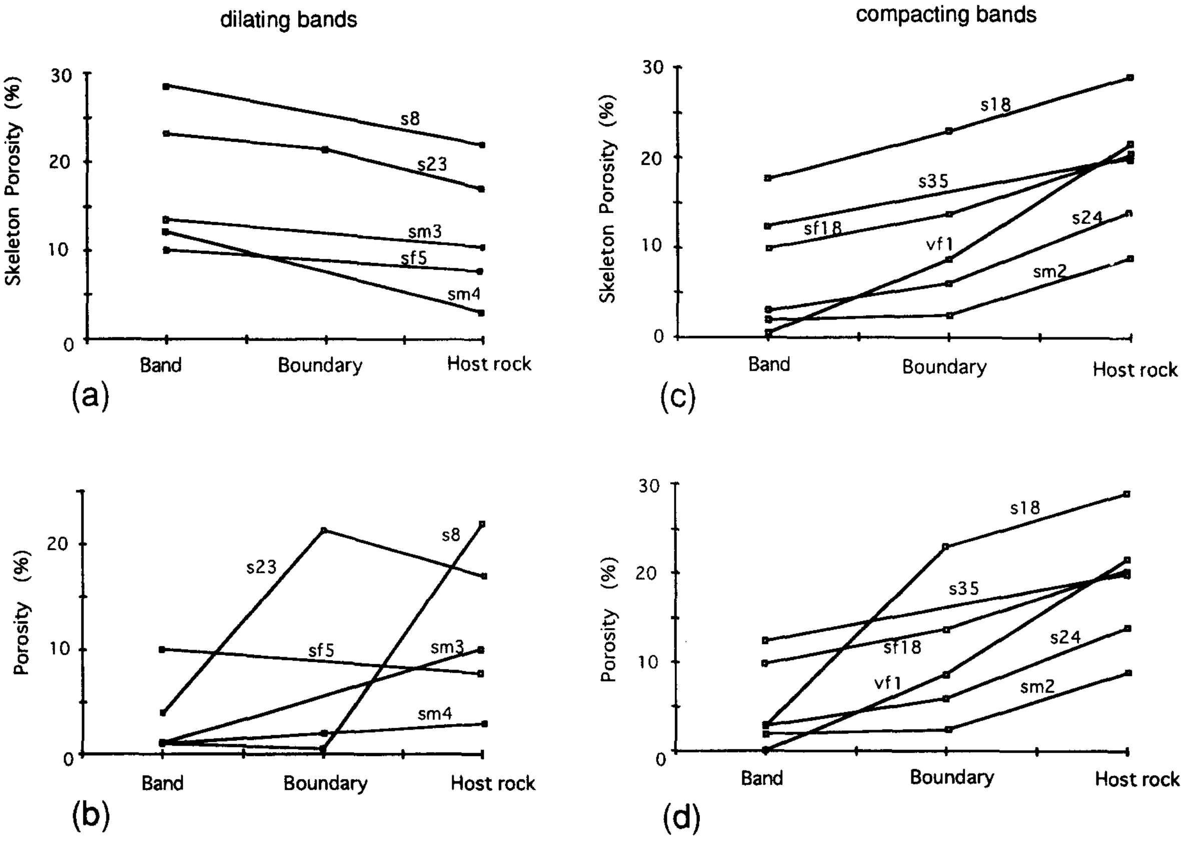

Fig. 9. Porosity profiles across deformation bands with no cataclasis. The porosity has been calculated via image analysis in the band, in the boundary between band and host rock and in the undisturbed host rock. (a) Skeleton porosity profiles in bands showing dilation. (b) Porosity profiles in bands showing dilation. (c) Skeleton porosity profiles in bands showing compaction. (d) Porosity profiles in bands showing compaction.

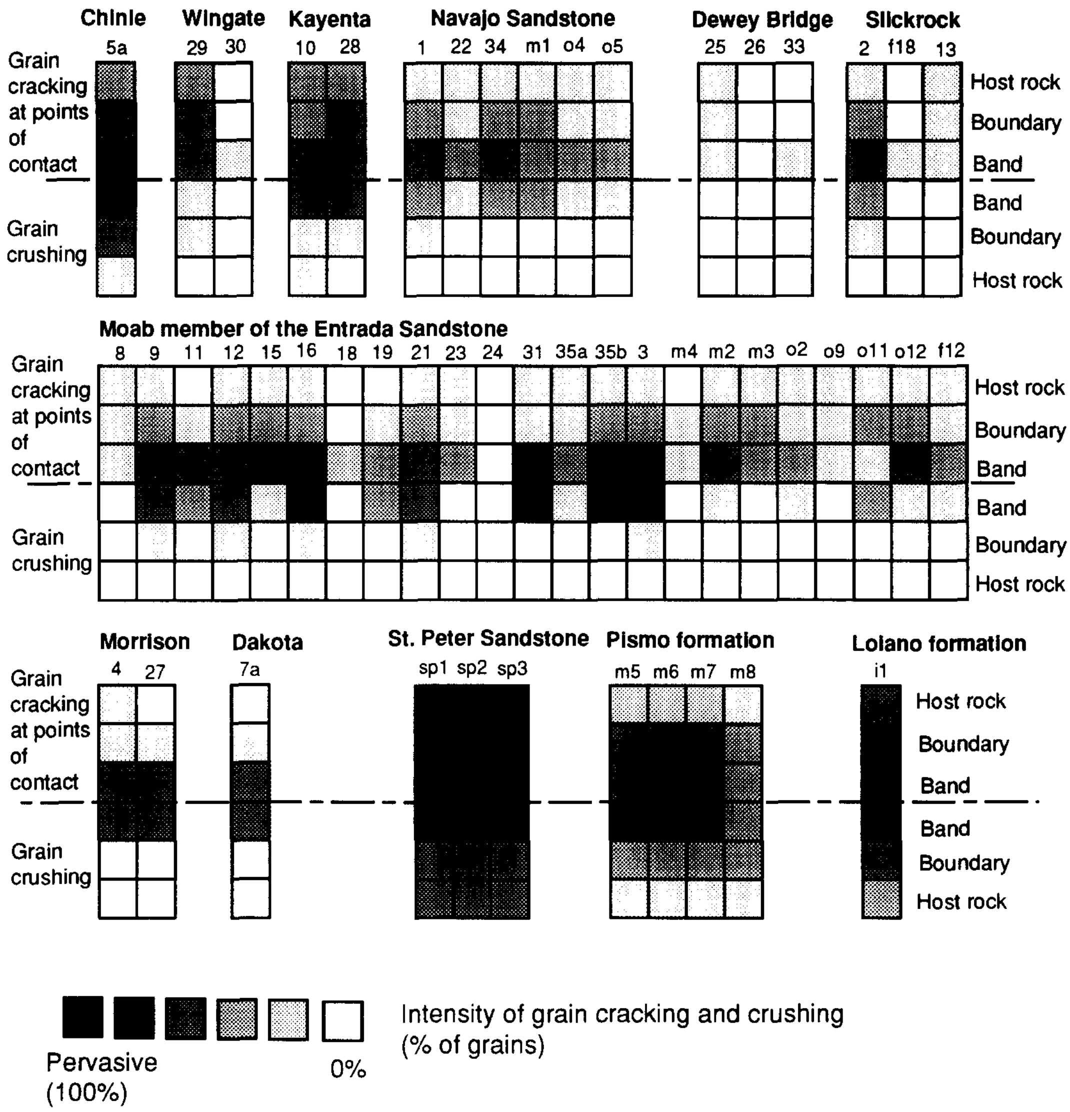

Fig. 10. Grain cracking and crushing associated with all types of deformation bands from a variety of locations. The intensity (% of cracked or crushed grains) is estimated in the band, in the boundary zone, and in the host rock.

that is not dependent on grain size (Fig. 8b). Their thickness varies from 0.2 to 3 mm regardless of grain size in the host rock. These bands do not follow the linear trend observed in experiments with unconsolidated granular materials made by Roscoe (1970) and by Vardoulakis *et al.* (1985), nor do they follow the relation proposed by the Mühlhaus & Vardoulakis (1988) model.

On the other hand, there is a weak relationship between porosity in the host rock and thickness of bands with cataclasis (Fig. 11a). The samples cluster in two groups, one with a host rock porosity between 1 and 7% the other with a porosity between 15 and 30%. For the first group the 90% confidence interval for mean band thickness is from 0.51 to 0.8 mm, whereas for the second group it is from 1.1 to 1.63 mm. Where the porosity in the host rock is >15%, the bands tend to be about 0.5 mm thicker. For bands with no cataclasis (Fig. 9b) there is no apparent correlation between porosity and band thickness.

The porosity loss in bands with crushed grains is usually very pronounced (Figs. 4c & d). For example, in the Moab member of the Entrada Sandstone the porosity of the host rock is on average around 17%; whereas the porosity in the band is always less than 7% (Figs. 11c & d). The decrease in porosity within the band appears to be more accentuated where the porosity in the host rock is larger.

Examining the distribution of the skeleton porosity in Fig. 11(c) it is interesting to note how a few of the bands (so12, s3, sf3) in the Moab member of the Entrada Sandstone show a positive dilatancy, especially in the boundary zone of the band. The grain crushing in these bands is very poorly developed, probably because they represent a stage of transition from bands with no cataclasis to bands with cataclasis. The microstructure of these samples suggests that positive dilatancy during deformation band development is an early condition, while grain crushing and compaction is a subsequent condition, that takes over as shearing progresses. Positive dilatancy in the boundary is not recognizable in the

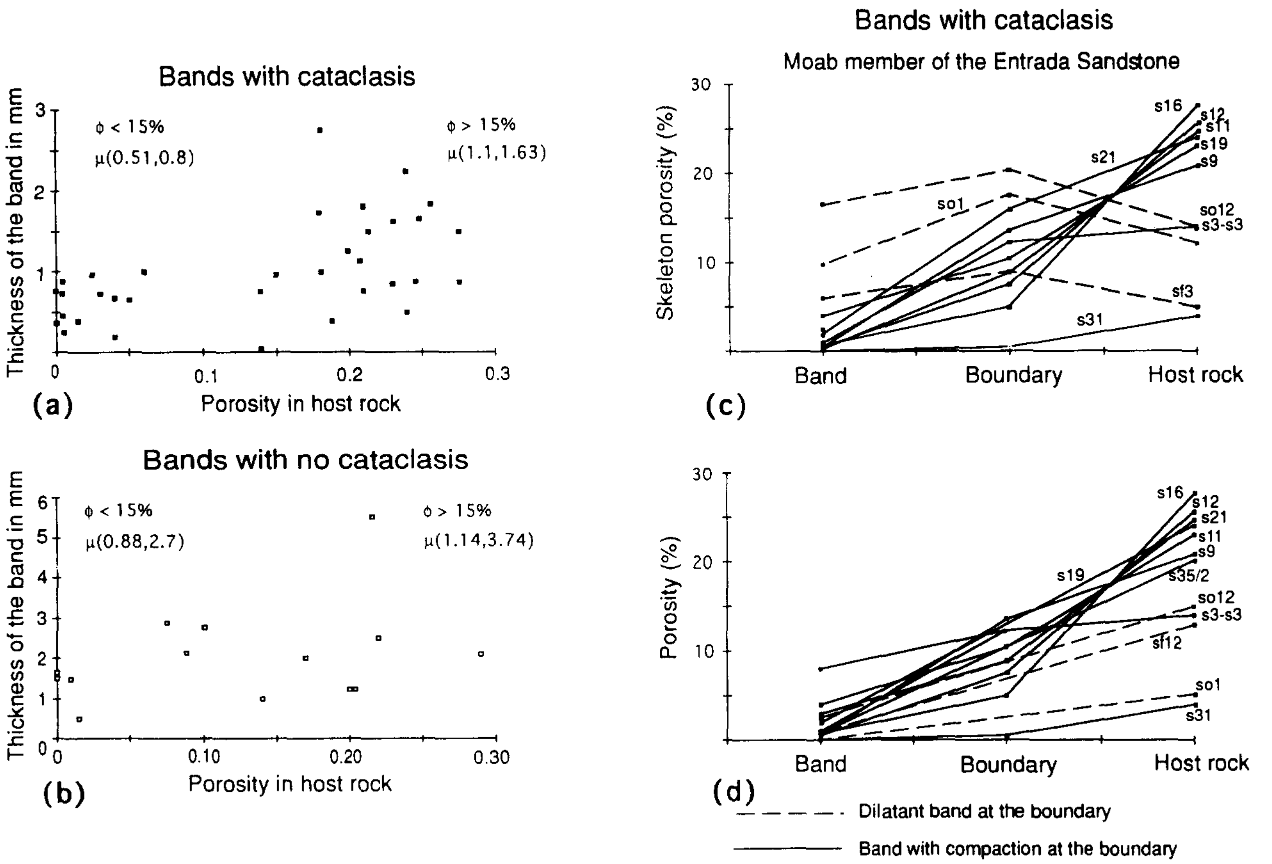

Fig. 11. Relation between porosity in the host rock and band thickness for deformation bands with cataclasis (a) and for bands with no cataclasis (b). Diagrams (c) & (d) represent skeleton porosity and porosity profiles respectively across bands with cataclasis in the Moab member of the Entrada Sandstone. Note how the porosity decrease in the band and in the boundary zone is more accentuated for deformation bands developed in host rock with high porosity. Some bands also show dilatancy in the boundary zone; these bands show very poorly developed cataclasis.

bands with well-developed cataclasis, rather one observes a zone of compaction with very little grain crushing (Aydin 1978). The amount of compaction in the boundary zone is also larger, if the host rock has a higher porosity.

The amount of polycrystalline quartz in bands with respect to the host rock decreases as the amount of cataclasis increases. It seems that polycrystalline quartz is weaker than single monocrystalline quartz grains and so tends to split into individual smaller grains during deformation. The amount of small-grain-size feldspar tends to increase in bands with respect to the small-grain-size quartz as the amount of cataclasis increases. The feldspars, and to a certain extent the rock fragments, tend to crush preferentially because of cleavage or pre-existing discontinuities at grain boundaries. Apparently feldspar grains are the first to crush or crack in the host rock and are among the most frequently crushed grains in proximity of the band.

When the host rock has >30% feldspar and rock fragments, and if the confining pressure was large, as suggested by overburden thickness, the amount of grain cracking in the host rock is large (Fig. 5d). In the samples of the Pismo and Loiano Formations the amount of cataclasis is very large and the deformation band is about 1.5–2 mm thick, a value slightly larger than for the bands in quartz-arenites. Other conditions such as degree of lithification, porosity and clay content were comparable in these two sandstones. The cataclasite is made up of a very fine material that is partially replaced by clay, indicating a high feldspar content.

Sandstones with well-developed cataclasis are well sorted. Grain cracking within a band is accentuated where grains of the same size are in contact; grains of large size surrounded by smaller grains, often are relatively uncracked (Fig. 5e).

The offset across bands exhibiting cataclasis varies from a fraction of a mm to 1–2 cm. The bands with largest absolute offset in clean sandstones form where the porosity in the host rock is high (>20%); however the relation between porosity of the host rock and offset across the band (Fig. 12) is not as clear as the linear trend presented by Smith (1983). Among the bands with cataclasis there are two clusters of data; one for sandstones with porosity less than 5% and one for sandstones with porosity larger than 13%. In each of these two clusters offsets increase with porosity, but maximum and minimum values are comparable in two clusters.

In the sandstone types that were investigated, porosity is not the only factor controlling the amount of offset. For example, the clay content, if not authigenic, appears to be an important factor. Typically, the amount of clays and iron oxides in the host rock is less than 5%, but occasionally it exceeds this amount (e.g. in the

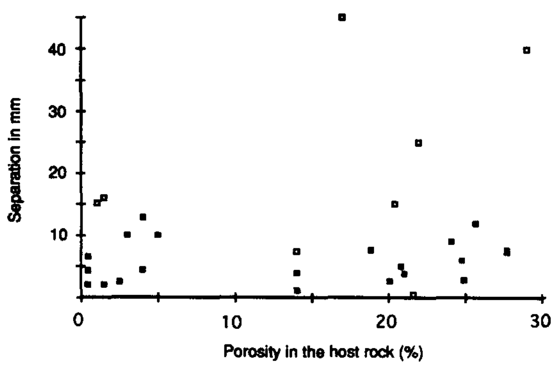

Fig. 12. Relation between deformation band offset and porosity in the host rock. Solid boxes are for bands with cataclasis, empty boxes are for bands with no cataclasis. No relation is apparent between amount of offset and porosity.

Slickrock member of the Entrada Sandstone and in the Wingate Formation) and is associated with larger offsets on the order of a few cm. The data for bands in clay-rich sandstones have been omitted from the plots in Fig. 12 to avoid complication.

Deformation bands with cataclasis often weather out in positive relief, forming characteristic long and narrow ridges in outcrop, due to the effect of compaction and to particular diagenetic phenomena that preferentially take place within the band. In the tiny pore spaces of bands with cataclasis it is possible to detect small amounts of carbonate or silica cement. According to Wollast (1971) and Berner (1980) authigenic cementation starts at grain contacts. Less free energy is required to nucleate the cementing material as the distance between the contacts decreases. If this distance is sufficiently small it is possible to precipitate cement from an undersaturated solution. Apparently the very tiny pore spaces among the crushed grains in a deformation band provide sites where nucleation of cement can be triggered more readily than in the host rock.

(3) *Deformation bands with clay smearing or localized shear zones*. High clay content and/or high porosity in the host rock are relevant factors in controlling deformation band microstructure and localization characteristics. These factors are so important that it is necessary to introduce a third kind of deformation band to account for these characteristics.

At Klondike Bluffs and at Delicate Arch in the Navajo Sandstone and in the Slickrock member of the Entrada Sandstone, very colorful bundles of deformation bands occur with color ranging from white to orange and purple. These zones are usually narrow (10–50 cm thick) and are often in proximity of slip planes. Observed in a slip-parallel cross-section these 'zones of deformation bands' look like a single, or a set of narrowly-spaced shear zones (Figs. 13a & b). The thin clay-rich colorful cross beds in the Navajo sandstone are sheared and stretched (smeared) along the shear zones so that in outcrop exposure they appear as coloured deformation bands (Fig. 6). We call these structures 'deformation bands with clay smearing' or 'localized shear zones'.

The deformation bands in these outcrops have different characteristics from those observed elsewhere in the Park. The thickness of deformation bands with clay smearing is dependent on the amount of offset they accommodate, they may grow thicker than an individual deformation band, and they may accommodate offsets in excess of 15 cm. In some places, the bands constitute a single entity a few mm to a few dm wide, elsewhere they are separated one from the other by thin lenses of undisturbed host rock. Rock within the localized shear zones is characterized by a reduction of porosity caused by pore collapse, cataclasis and compaction. The shear zones may cluster in narrow zones next to fault planes where they look very different from the broad zones of deformation bands in proximity of slip planes within the Moab member of the Entrada Sandstone (Fig. 13c).

There is a striking difference between zones of deformation bands in the Moab member of the Entrada Sandstone and shear zones in the Navajo Sandstone. The porosity in the host rock within the Moab member of the Entrada is on average 17%, whereas in the Navajo sandstone it is on average 23%. The clay content in the Moab is less than 1%, whereas in the Navajo can be up to 5%. The Moab member of the Entrada Sandstone is about 30 m thick, whereas the Navajo Sandstone is about 300 m thick in the Arches area (Doelling 1985). At Klondike Bluffs a shear zone (Fig. 7) is located about 0.8 km west of the faults bounding the down-dropped part of the salt-cored anticline in Salt Valley. The shear zone is about 3 m thick (about 10 times the usual zone of deformation bands thickness) and accommodates a normal offset on the order of 10 m, yet it is not associated with a slip plane. The total offset is accommodated by distributed shear across the zone (Fig. 13a).

A transect across Salt Valley (Antonellini *et al.* in preparation) shows that closer to the axis of the valley, where the strains due to anticline collapse have been larger, the width of the zones was narrower (20–30 cm) and the zones were systematically associated with slip planes. In these examples, slip plane localization seems to be controlled by the amount of strain.

Shear zones with similar structure, although much narrower (2–10 cm), have also been observed in the Slickrock member of the Entrada Sandstone. Here the only controlling factor seems to be the high clay content, which is about 15%. It therefore is possible to define two subgroups for these narrow shear zones (Table 1), one with high porosity and low clay content in the host rock, the other with high clay content and low porosity in the host rock.

Distribution and age relations of deformation bands

Deformation bands occur in two distinct spatial arrangements at Arches National Park: widely and uniformly distributed over large areas, or narrowly-spaced in proximity to major faults. Deformation bands with no cataclasis occur over large areas, whereas deformation

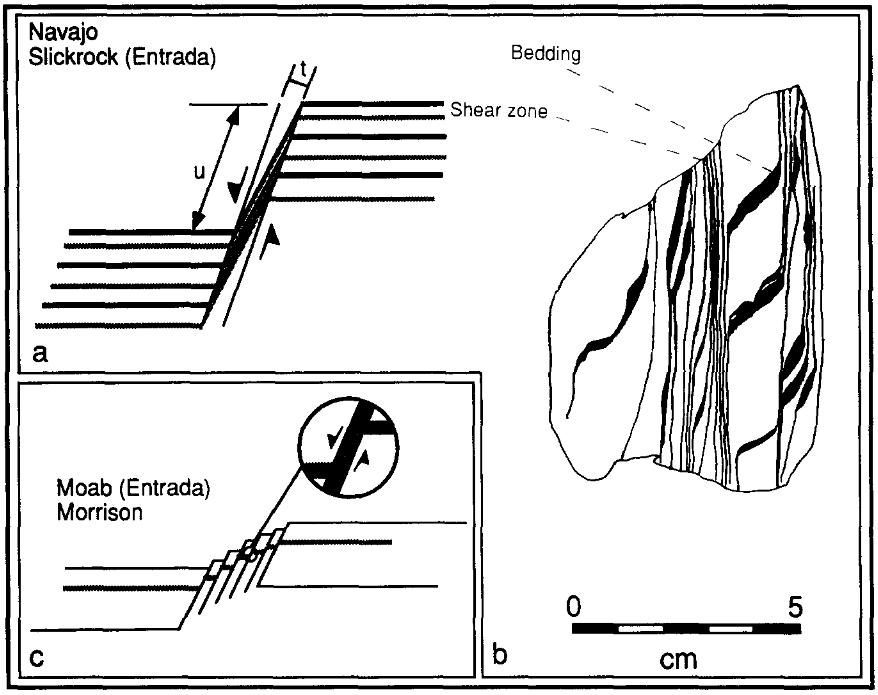

Fig. 13. (a) Zones of deformation bands peculiar to the Navajo Sandstones and the Slickrock member of the Entrada Sandstone. These zones are usually narrow (10–50 cm thick) and were localized in proximity to slip planes. Observed in cross-section parallel to the slip vector these 'zones of deformation bands' look as a single localized shear zone. (b) Sketch of a sample from one of these shear zones. The face represented is perpendicular to the shear zone and parallel to the slip vector. (c) Zones of deformation bands typical of the Moab member of the Entrada Sandstone and of the Morrison Formation. The bands are spread over a zone that may span 0.2–10 m in width and that may be associated with one or more slip planes. Each individual band is about 1 mm thick and represents a small shear zone.

bands with cataclasis occur in proximity to fault zones. In the Moab member of the Entrada Sandstone at the Delicate Arch Viewpoint (Fig. 1) it is possible to observe bands with these two different spatial distributions in the same rock type. The first set of deformation bands has no crushed grains, an orientation (N60E) at about 30° to the major fault in the area (parallel to Cache Valley, N90E) and belongs to a system of bands which is widely spaced within the sandstone unit (30–50 m). The second set of deformation bands has well-developed cataclasis and belongs to a system of bands localized in a highly deformed zone near the major fault of the area. From cross-cutting relations bands with no cataclasis are consistently older than bands with cataclasis (Fig. 14). These relations suggest that there are at least two generations of deformation bands at Arches National Park. For example, the deformation bands described by Zhao & Johnson (1991) in the Garden area of the Park have wider spacing and do not appear to be associated with any major fault. These bands have, in fact, very little cataclasis and fall into the older generation.

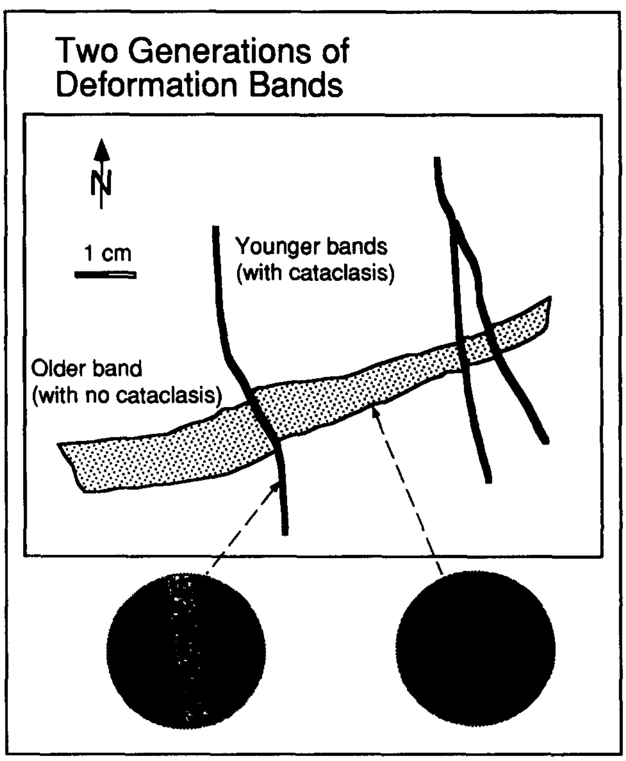

Fig. 14. Two generations of deformation bands at Arches National Park. Map of sample 35 collected in the Delicate Arch viewpoint area. The band of the first generation has no or very little cataclasis, whereas the bands of the second generation have a well-developed cataclastic zone.

DEFORMATION BANDS IN TRIAXIAL EXPERIMENTS AND INFLUENCE OF STRESS STATE AND POROSITY

In the following we summarize the results of triaxial lab experiments that allow us to constrain the effect of stress state and porosity on the development of cataclasis and dilatation during straining of a soil sample and/or

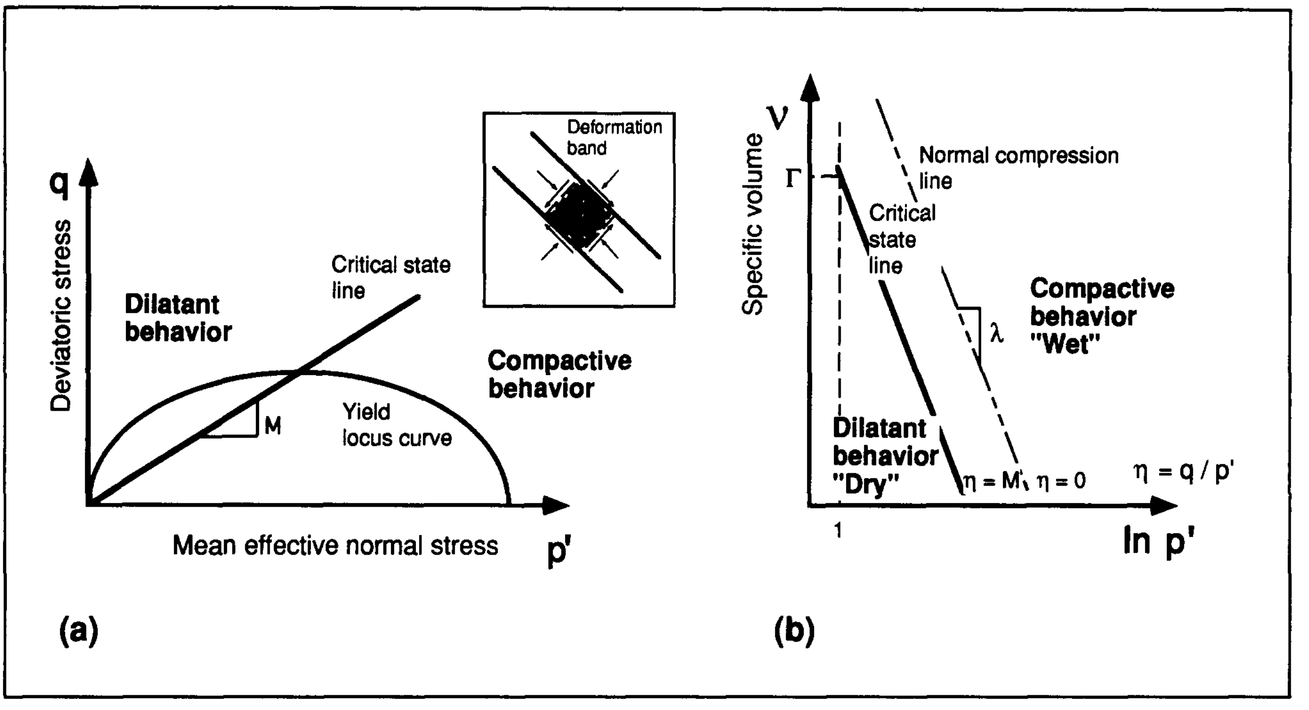

Fig. 15. (a) Critical state line is plotted in 'deviatoric stress' – 'mean effective normal stress space'. The slope of the line M, represents a frictional coefficient, which is a constitutive property of the material. The yield locus for a granular material is shown by the curve. In the field to the left of the critical state line the material dilates after yielding, in the field to the right it compacts. (b) Critical state line in 'specific volume' – 'log of the mean effective normal stress space'. Γ is the ordinate of the critical state line and λ is the gradient of compression line. Inset shows the boundary conditions.

deformation bands localization. In the light of the experimental results we discuss the conditions of development of the deformation bands described earlier.

Deformation bands and soil behavior in triaxial experiments

The relations between porosity in the host rock and microstructure in the deformation bands can be analyzed in terms of critical state theory (Schofield & Wroth 1968). This theory is based on constitutive properties of the material such as porosity and angle of internal friction, and on the stress state during deformation. Extensive triaxial experiments on unconsolidated granular material support this theory (Wood 1990). The critical state concept proposes that a soil or a granular material, when continuously strained, can flow like a fluid with no volume change if it is in a well-defined critical state (Schofield & Wroth 1968). The critical state is described by two equations (Schofield & Wroth 1968):

$$q = Mp' \tag{1}$$

$$v = \Gamma - \lambda \ln p' \tag{2}$$

where q is the deviatoric stress, p' the mean effective normal stress, v the specific volume, and M, Γ, λ are constitutive properties of the material defined in Figs. 15(a) & (b).

Equation (1) defines the critical state with a line in deviatoric stress (q)—effective mean normal stress (p') space (Fig. 15a). Equation (2) defines the critical state with a line in specific volume (v)—log of the effective mean normal stress (p') space (Fig. 15b). In Fig. 15 the critical state line divides the field in which the material has positive dilatant behavior (left of the line) from the field where it has a compactive behavior, when it is strained beyond its yield point (represented by the yield locus curve). It is apparent from Fig. 15 that large values of the mean effective compressive normal stress promote compaction whereas small values promote positive dilatancy. Figure 15(b) shows the relation between specific volume and the log of the effective mean compressive normal stress. Low porosity samples (low values of the specific volume) when strained tend to have a dilatant behavior, even at moderate effective mean normal stresses. On the other hand high porosity samples (large specific volumes) show compaction after yielding.

These relations can be used to interpret the thin section observations and the porosity determinations of several types of deformation bands that occur in Arches National Park. In particular, by considering the band in its post-bifurcation state (after strain localization into a band; see inset in Fig. 15), one can apply the theory to understand the behavior of the material within the band. Deformation bands with no cataclasis in low porosity sandstone have the largest positive dilatancy (a porosity differential of about 8% relative to a host rock with original porosity around 4%). Low porosity in the host rock also controls the development of positive dilatancy in deformation bands with poorly developed cataclasis. Deformation bands with and without cataclasis in high porosity host rock (porosity larger than 20%) almost invariably show compaction and voids reduction. Bands formed in sandstones with very poor sorting, and high clay content deform with negligible or no volume change. It appears that deformation with flow in a critical state (no volume change) is more likely to take place in an immature unsorted shaly sandstone. The effect of the mean compressive normal stress on deformation bands is harder to test from the thin section observations alone. However, the results obtained from laboratory experiments are helpful in interpreting the effects of mean compressive normal stress on dilatancy and compaction within a deformation band if the porosity in the host rock is known.

Influence of stress state and porosity on grain crushing

The thickness of the sedimentary cover on top of the Entrada Sandstone has been calculated from stratigraphic sections in neighboring areas to be in a range from 1000 m to more than 2400 m (Stokes 1987). These figures would give a confining pressure on the order of 20–50 MPa. This estimate does not consider the presence of pore fluid pressure which would decrease the effective confining pressure.

The observations of sandstone microstructure, when compared to the results of lab experiments, help to infer the stress state at the time of deformation band growth. Zhang *et al.* (1990) and Wong & Davis (1992) have shown with triaxial experiments that higher confining pressures cause an increase in the amount of grain cracking and crushing within sandstone specimens. Increased grain cracking at points of contact with higher confining pressure is also observed in triaxial experiments on sandstones in the cataclastic flow regime (Wong & Davis 1992). Zhang *et al.* (1990) have made hydrostatic compaction tests on porous sandstones that showed how the number of microcracks at grain contacts increased as the pressure increased, and that a critical pressure existed above which grain crushing and pore collapse in the sample became pervasive. This critical pressure is lower in high porosity sandstones. Localization of deformation bands in high porosity sandstone occurs at lower uniaxial loads (confining pressure constant at 1 MPa) than in low porosity sandstone (Dunn *et al.* 1973).

Dilatation during shear band development in a porous sandstone may be caused by the movement of the grains that try to slide past each other and to a lesser extent by grain cracking. Triaxial experiments on sandstone show the occurrence of significant amounts of positive dilatation at low values of confining pressure (Robinson 1959, Gowd & Rummel 1980, Fischer & Paterson 1989, Bernabe & Brace 1990) with peak values just before shear band localization. The amount of dilatation usually decreases with increasing values of the confining pressure. From these experiments it appears that grain sliding may cause dilatation at low confining pressure whereas grain cracking may cause dilatation at larger confining pressures.

We conclude that the pervasive grain cracking and crushing in the younger deformation bands is indicative of higher values of confining pressure, whereas the little grain cracking and crushing in the older deformation bands is indicative of lower values of confining pressure.

In addition, Dunn *et al.* (1973), Hirth & Tullis (1989), Bernabe & Brace (1990) and Rutter & Hadizadeh (1991) have investigated the effect of confining pressure and porosity on the brittle–ductile transition in siliciclastic rocks. The experiments of Hirth & Tullis (1989) considered the effect of strain. Their results show that high porosities and high confining pressures cause the deformation in the samples to remain distributed, and in the form of cataclastic flow accompanied by work hardening. However, the cataclastic flow is only a transient phenomenon that is followed by strain softening and slip plane localization when the strains become larger than about 20%. Slip plane localization, associated with positive dilatancy and strain softening, is also predicted by Rudnicki & Rice (1975) on a theoretical basis.

These experiments offer an explanation for the thickness and the distribution of the localized shear zones observed in the Navajo Sandstone at Klondike Bluffs. From a transect mapped perpendicular to the axis of Salt Valley (Antonellini *et al.* in preparation), and passing through the Klondike Bluffs area, it appears that the thicker zones (>1 m) form far from the axis of Salt Valley where the strains are smaller, whereas the thin zones (20–30 cm) associated with slip planes form in proximity to the axis of the valley where the strains are larger.

MICROMECHANICS OF DEFORMATION BANDS

From thin section observations reported in the previous literature (Aydin 1978, Aydin & Johnson 1978, Pittman 1981, Jamison & Stearns 1982, Underhill & Woodcock 1987, Jamison 1989) deformation bands have been described with a very distinct shape and a well developed cataclastic zone. These descriptions are essentially correct but incomplete: they have omitted (or perhaps de-emphasized) the initial processes that took place during the formation of a band. To understand the micromechanics of deformation band development the tip regions of five strike-slip deformation bands in the Garden area were cored and a total of 10 thin sections were examined for detecting any changes in the sandstone microstructure in the tip region.

It is often possible to identify patches of sandstone in outcrop exposure that have been selectively weathered in the tip regions of deformation bands. Authigenic clays and fluids containing iron oxides have preferentially infiltrated these tip areas. This localization of weathering is particularly accentuated along bands of the dilatant type with no cataclasis.

In three of the samples examined it was possible to recognize a distinct porosity increase (Fig. 16a) in patches (0.5–5 cm long) aligned and coplanar with the tip of bands with poorly developed cataclasis (Fig. 16b). The porosity in the host rock is about 10%, whereas in the patches it increases to 19%.

From these few examples one cannot generalize to all

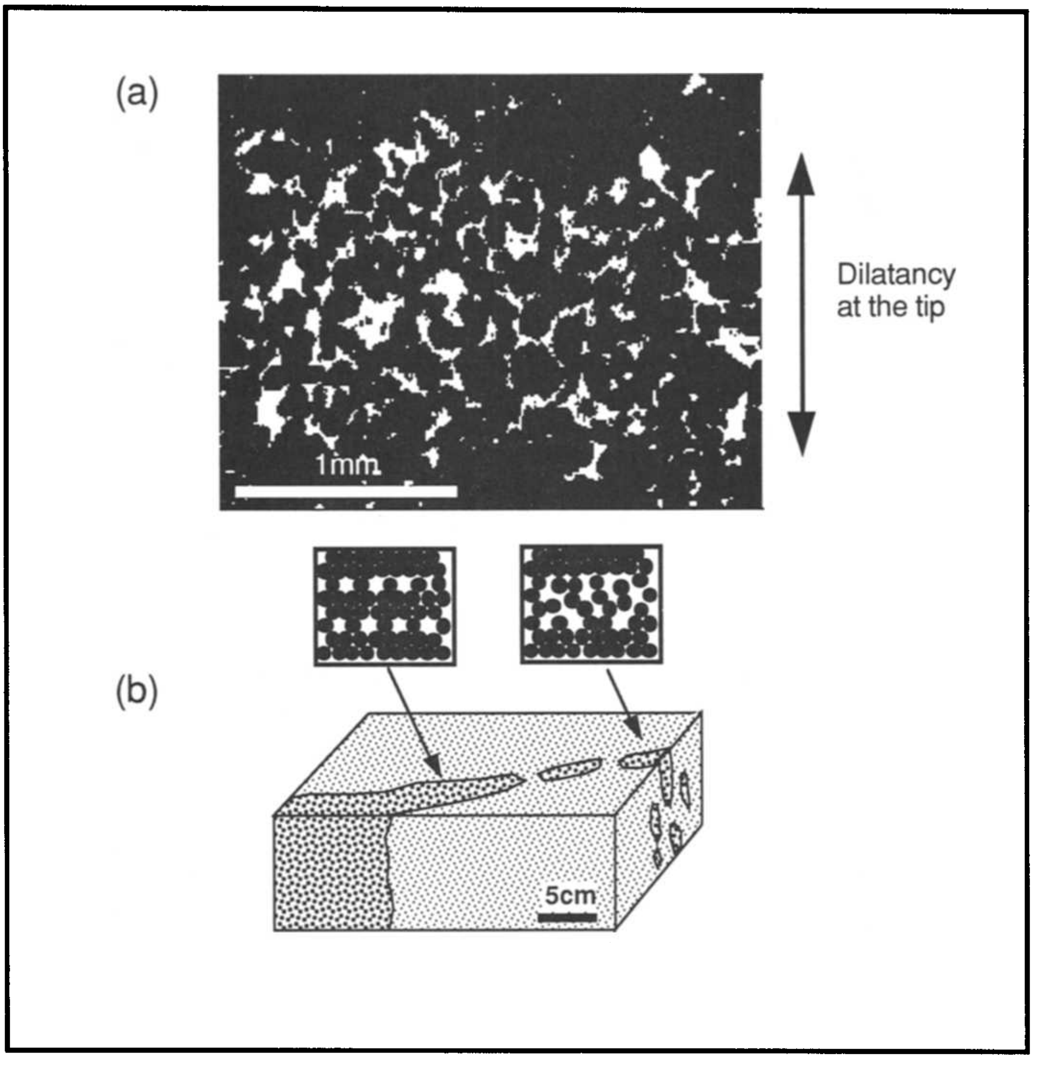

Fig. 16. Tip of a deformation band in Entrada Sandstone (Garden area, Arches National Park). (a) Segmented image of a band tip showing an increased porosity, (b) higher porosities are detectable in coplanar patches aligned with the tip of the band. The porosity increases from about 10% in the host rock to nearly 19% in the tip region.

types of deformation bands, but for the case of those studied in the Garden and Eye of the Whale area the following hypothesis can be framed. Provided that the porosity of the host rock is low enough (<10%) and the mean compressive normal stress is small enough such that contact forces should not be able to cause grain crushing, the initial micromechanism of deformation at the tip causes a positive dilatancy. This occurs as the grains tend to slide and roll around each other (Rowe 1962). Some bands apparently develop with positive dilatancy, whereas for others the increased porosity may be just a transient phenomenon. In the latter case continued shearing leads to porosity reduction and cataclasis. This hypothesis is supported by the porosity determinations (Fig. 11c) in the boundary region of some bands sampled in the Garden area. The central part of these bands shows compaction but the initial dilation is preserved within the band boundary.

During this initial positive dilation the porosity increase caused by sliding and rolling of grains lead to a decrease in the number of contact points among grains. This decrease in contact points among grains leads to a load concentration at the remaining contacts and hence this mechanism can trigger grain crushing via opening-mode splitting of the grains (Fig. 5f). The splitting occurs along the trajectories of maximum compressional stress connecting the points of contact (Timoshenko & Goodier 1970, Gallagher 1974). Figure 17(a) is an idealization of how the process of cataclasis and pore collapse may develop from a tight hexagonal packing. The particles (here modeled as spheres) slide and roll from their initial position increasing the distance from their centers from $2R - d$ to $2R$. The contact points diminish from four to three and tensional cracking starts at the points of contact. Subsequent shearing leads to cataclasis (Fig. 17b). However, if the initial packing is very loose and the porosity is high (Fig. 17c), grain splitting at contact points may lead directly to pore collapse and porosity reduction without an initial stage of positive dilatation. Grain splitting at contact points may be aided by pre-existing flaws or cleavage planes in the grains.

High values for mean compressive normal stress result in large point forces at grain contacts and may induce grain cracking without an initial reduction of the contact points. This was recognized in triaxial laboratory experiments on sandstone samples by Gowd & Rummel

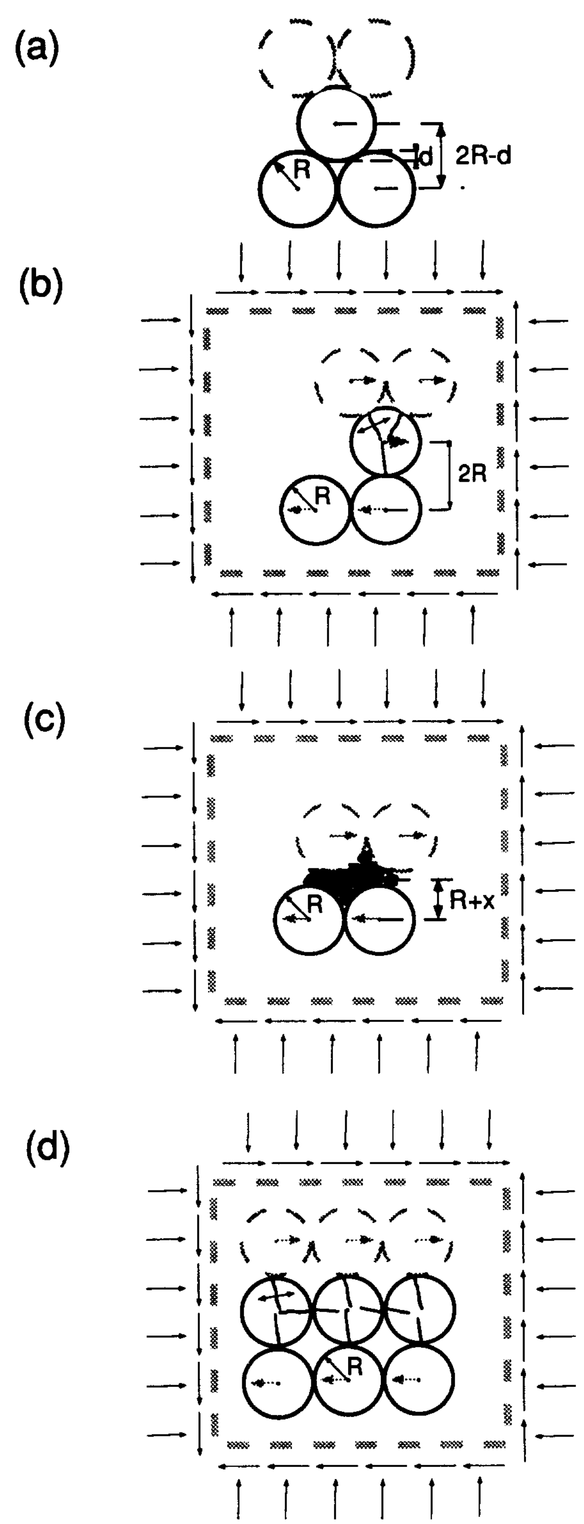

Fig. 17. Micromechanisms involved in the formation of deformation bands. (a) Initial packing of spherical grains prior to shear deformation. (b) Positive dilatation and increase in porosity during shearing with grain cracking caused by stress concentration at fewer points of contact. (c) Grain crushing and pore collapse leading to compaction. (d) In high porosity sandstone grain collapse can begin by cracking at grain contacts without an initial dilational stage.

(1980). In the thin sections examined here, the bands with better developed cataclasis formed where the grains in the host rock were already cracked to a considerable extent (Pismo and Loiano Formations samples) or where the band formed in close proximity to a slip plane. In both cases it is reasonable to assume that the mean compressive normal stress had high values. These thin section observations also correlate with experimental work done by Zhang *et al.* (1990) and by Wong & Davis (1992), who showed an increase in grain cracking within sandstone samples deformed at increasing confining pressure.

From our thin section observations one can see that as the amount of cataclasis increases some large grains are surrounded by finer grains and remain relatively intact with little grain cracking (Fig. 5e). It appears that larger grains surrounded by grains of smaller size are in a relatively stable condition. In this regard Sammis *et al.* (1987), have proposed the theory of the 'nearest neighbor', developed from observations of grain size distributions in the gouge zone of a fault with hundreds of meters of slip. According to this theory each particle's fracture probability is determined solely by the relative size of its nearest neighbor, and it does not depend on the absolute particle size or mineralogy. The principle of this theory is that many particles of different size distribute the load among themselves at many points of contact, while similar particles in contact concentrate the load at fewer points of contact. Thus if two particles of the same size are next to each other one will fracture (Biegel *et al.* 1989). Similarly, in our study, grains of similar size in contact often were cracked and large grains within a fine matrix were relatively uncracked. However, it appears that mineralogy and grain size are also important controls. In fact feldspars, polycrystalline quartz and rock fragments are generally more susceptible to cracking, regardless of the size of their nearest neighbor.

DEFORMATION BANDS AND GROWTH OF THE SALT ANTICLINE, ARCHES NATIONAL PARK

In a previous section we have discussed how deformation bands with no cataclasis are widely distributed over large areas, whereas deformation bands with cataclasis are narrowly spaced in proximity to major faults; we have also mentioned that the bands with no cataclasis are older than the bands with cataclasis.

An interesting hypothesis can explain the observed relations between bands and regional structures, and their temporal sequence. At Cache Valley the bands of the younger generation (bands with cataclasis) are localized in proximity to the major faults. Their origin therefore is inferred to be related to the collapse of the anticline caused by salt dissolution during which the major faults localized (Doelling 1985). On the other hand the bands of the older generation (bands with no cataclasis) may have formed during the growth of the anticline when the sediments were not yet fully lithified and when the mean compressional normal stress had low values. In this period the stress field was probably controlled by salt diapirism.

The stereogram representing the systems of bands and the fault solutions for the Little Valley area (close to sample location 8) show oblique-slip (Fig. 1). A similar situation, but with a slightly larger component of dip-slip, is present at North Klondike Bluffs (close to sample location 15). In the Garden area (close to sample location m3; data from Zhao & Johnson 1991), in the Eye of the Whale area (close to sample location f18) and at Park Avenue the strain is accommodated by prevalent strike-slip. In all these areas the deformation bands have little or no crushed grains. The first two localities are closer to the anticlinal axis whereas the latter two are farther away. A stereogram and a fault solution for the easternmost Klondike Bluffs area (Fig. 1) uses bands associated with major faults bounding the collapsed anticline. These have almost pure dip-slip solutions and a very well-developed cataclastic zone. These younger bands may have formed during the collapse of the salt anticline.

Underhill (1988) used conjugate systems of deformation bands to assess the direction of maximum extension in the vicinity of a salt diapir in the island of Zakinthos (Greece). His plots and maps of the band attitudes and of the maximum direction of extension suggest that dip-slip and oblique-slip are dominant adjacent to the diapir whereas farther away (1–2 km) the strain is accommodated by strike-slip faulting. The observations of Underhill (1988) were relative to bands formed in proximity to a salt diapir which has not collapsed and therefore it is reasonable to assume that they formed during the growth of the diapir. His observations are also consistent with the structural data collected in proximity of the salt anticline at Arches National Park, as represented by the stereonets and by the arrows denoting the direction of maximum extension in Fig. 1.

CONCLUSIONS

On the basis of their microstructure, deformation bands may be classified according to the occurrence of cataclasis and dilation or compaction (see Table 1). Three types can be identified: (1) deformation bands with no cataclasis; (2) deformation bands with cataclasis; and (3) deformation bands with clay smearing. The common characteristics among the structures observed are the small amounts of offset (mm to dm), the lack of a well-defined discontinuity plane and the typical connecting structures among different segments (eye and ramp structures).

Deformation bands may cluster in zones that have different characteristics according to the initial porosity of the host rock and to the clay content. In proximity to faults and folds, where the host rock has a high clay content and/or a high porosity, the zones tend to be narrow and localized. Where the clay content and the porosity in the host rock are lower, the bands tend to be distributed in wider zones. Higher clay content in the host rock also correlates with larger offsets across the bands.

Thin-section studies have documented relations between host rock grain size and band thickness, host rock porosity and band thickness, host rock porosity and amount of offset, grain size in the cataclastic zone and the intensity of microcracking in proximity to deformation bands. Feldspars, microcrystalline quartz and rock fragments are broken and shattered preferentially with respect of monocrystalline quartz.

Provided a tight enough grain packing and moderate mean compressive normal stresses, the micromechanics of deformation band growth appears to be controlled by initial dilatancy as the grains slide and roll around each other. This initial dilatancy may be a transient phenomenon that is followed by grain splitting at points of contact and porosity reduction as shearing progresses. This mechanism probably leads to grain crushing and rotation. High values of the mean compressive normal stress may inhibit positive dilatancy and promote grain crushing from the onset of shearing. This micromechanical model is supported by porosity determinations at the tip of deformation bands and by the porosity variations in profiles across deformation bands with poorly developed cataclasis.

Two generations of deformation bands were documented at Arches National Park. The older generation consists of bands with little or no cataclasis that are distributed in relatively undeformed sandstone. The younger generation consists of bands with well-developed cataclasis that are localized in proximity to major structures. Microstructural observations and inferences from lab experiments suggest that the two generations of deformation bands developed under different stress states. This leads to the hypothesis that the first generation of bands formed during the growth of the salt anticline, whereas the second generation formed during its collapse.

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