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Deformation bands in sandstone: a review

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Abstract: Deformation bands are the most common strain localization feature found in deformed porous sandstones and sediments, including Quaternary deposits, soft gravity slides and tectonically affected sandstones in hydrocarbon reservoirs and aquifers. They occur as various types of tabular deformation zones where grain reorganization occurs by grain sliding, rotation and/or fracture during overall dilation, shearing, and/or compaction. Deformation bands with a component of shear are most common and typically accommodate shear offsets of millimetres to centimetres. They can occur as single structures or cluster zones, and are the main deformation element of fault damage zones in porous rocks. Factors such as porosity, mineralogy, grain size and shape, lithification, state of stress and burial depth control the type of deformation band formed. Of the different types, phyllosilicate bands and most notably cataclastic deformation bands show the largest reduction in permeability, and thus have the greatest potential to influence fluid flow. Disaggregation bands, where non-cataclastic, granular flow is the dominant mechanism, show little influence on fluid flow unless assisted by chemical compaction or cementation.

Deformation of stiff, low-porosity rock in the uppermost few kilometres of the Earth's crust occurs primarily by fracturing. This can result in extensional fractures, such as joints and veins, or shear fractures such as slip surfaces, which generally form the primary deformation elements of faults in low-porosity rocks. The process of fault formation and propagation in brittle low-porosity rocks has been described in terms of linking of microfractures and the reactivation or linking of mesoscopic joints (e.g. Pollard & Fletcher 2005). The key element in a fault is the slip surface, where the majority of offset has accumulated. Surrounding fractures constitute an enveloping damage zone (Caine *et al*. 1996). Slip surfaces and extension fractures, structures that will be referred to in this paper as ordinary fractures, typically represent mechanically weak structures that are prone to reactivation and continued slip during subsequent stress build-up.

Strain in highly porous rocks and sediments is not initially accommodated by extensional fractures or slip surfaces. Instead, strain localization occurs by the formation of deformation structures commonly referred to as deformation bands. Localized (higher offset) faults subsequently form by the failure of deformation band zones.

Deformation bands in porous rocks are low-displacement deformation zones of millimetres to centimetres thickness (Fig. 1) that tend to have enhanced cohesion and reduced permeability compared with ordinary fractures. Quaternary geologists find them in glacially or gravitationally deformed sand, where they may reveal information on the local glacial history. Sedimentologists frequently encounter them in sandstones, where they may be generated during soft-sediment deformation or post-burial faulting. Petroleum geologists and hydrogeologists (should) look for them in cores from clastic reservoirs and aquifers because of their potential role as barriers or baffles to fluid flow (Pittman 1981; Jamison & Stearns 1982; Gabrielsen & Koestler 1987; Antonellini & Aydin 1994, 1995; Beach *et al*. 1997; Knipe *et al*. 1997; Gibson 1998; Antonellini *et al*. 1999; Heynekamp *et al*. 1999; Hesthammer & Fossen 2000; Taylor & Pollard 2000; Lothe *et al*. 2002; Shipton *et al*. 2002, 2005; Sample *et al*. 2006) and because they commonly indicate proximity to a larger offset fault. From an academic point of view, deformation bands deserve attention because they provide important information on the unique way that faults form in porous sandstones (e.g. Aydin & Johnson 1978; Johnson 1995; Davis 1999) and on progressive deformation in porous rocks in general (e.g. Wong *et al*. 2004; Schultz & Siddharthan 2005). In this paper we review the existing literature on deformation bands, present a classification of deformation bands based on deformation mechanism and discuss how the distinctive characteristics of deformation bands relate to burial depth, lithology and fluid flow.

Characteristics of deformation bands

The term deformation band has long been used in different ways in fields such as material science (e.g. Brown *et al*. 1968) and crystal–plastic deformation of rock (e.g. Passchier & Trouw 1996); however, it was first applied in the context of sandstone deformation by Aydin and co-workers (Aydin 1978; Aydin & Johnson 1978, 1983). Since then, the term has gradually been adopted to encompass terms such as microfaults (Jamison & Stearns 1982), cataclastic faults (Fisher & Knipe 2001), faults (Manzocchi *et al*. 1998; Fisher *et al*. 2003), (micro)fractures (Borg *et al*. 1960; Dunn *et al*. 1973; Gabrielsen & Koestler 1987), shear bands (Menéndez *et al*. 1996), deformation-band shear zones (Davis 1999), Lüders' bands (Friedman & Logan 1973; Olsson 2000), cataclastic slip bands (Fowles & Burley 1994), and granulation seams (Pittman 1981; Beach *et al*. 1999; Du Bernard *et al*. 2002*b*). The most important characteristics of

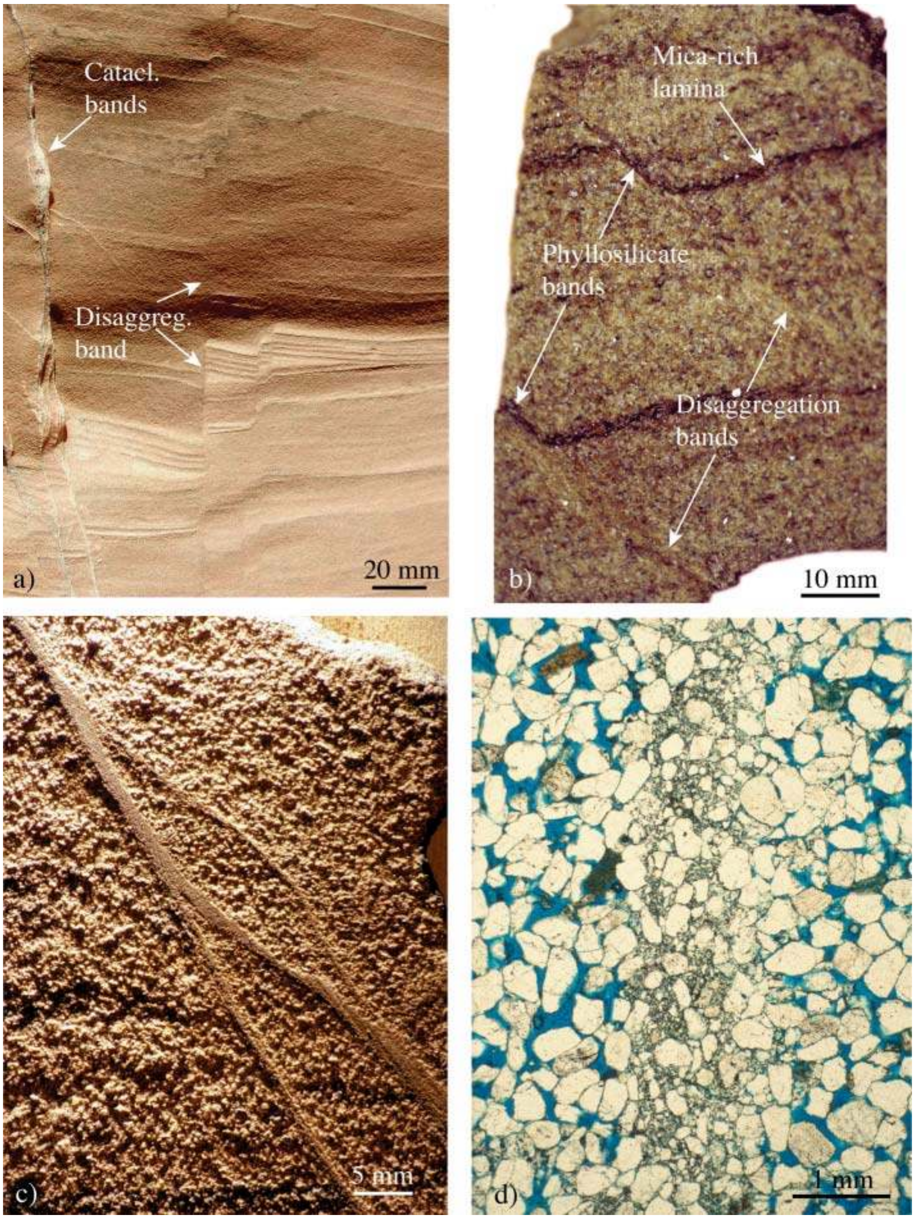

Fig. 1. (**a**) Disaggregation bands (centre, locally invisible) cut by cataclastic deformation bands (white) in the Navajo Sandstone, Utah. (**b**) Rapid variation from phyllosilicate band to disaggregation band in sandstone in Jurassic sandstone (Gullfaks Field, North Sea). Mica-rich layers are local sources of phyllosilicate minerals. (**c**) Phyllosilicate band, Brent Group, Gullfaks Field. The positive relief (increase of cohesion) and loss of porosity in the band should be noted. (**d**) Photomicrograph of a single cataclastic deformation band, showing a low-porosity cataclastic core mantled by a zone of compaction. Blue indicates pore space.

deformation bands (in the context of porous rock and sediment deformation) are summarized as follows.

(1) Deformation bands are restricted to porous granular media, notably porous sands and sandstones. The formation and evolution of a deformation band involves a significant amount of grain rotation and translation, and this process, whether it includes grain crushing or merely rotation and frictional sliding along grain boundaries, requires a certain amount of porosity. If porosity is too low, then tension fractures, stylolites and/or slip surfaces will preferentially form.

(2) A deformation band does not represent a slip surface. Slip surfaces can, however, form within bands or, more commonly, along or within zones of deformation bands, but this represents a more mature stage in the development of deformation band faults.

(3) Deformation bands occur hierarchically as individual bands, as zones of bands, or within zones associated with slip surfaces (also known as faulted deformation bands).

(4) Individual deformation bands rarely host offsets greater than a few centimetres even when the bands themselves are 100 m long. Localized higher-offset faulting in porous rocks commonly occurs by the failure of existing deformation band zones along a slip surface.

(5) Deformation bands are found in many upper-crustal tectonic and non-tectonic regimes (Fig. 2).

There are several important characteristics that distinguish deformation bands from ordinary fractures (such as slip surfaces or extension fractures). First, they are thicker and exhibit smaller offsets than classical slip surfaces of comparable length. Also, whereas cohesion is lost or reduced across ordinary fractures, most deformation bands maintain or even increase cohesion. Furthermore, deformation bands often exhibit a reduction in porosity and permeability, whereas both slip surfaces and tension fractures are typically associated with a permeability increase. Strain hardening behaviour, commonly associated with deformation band formation, also contrasts to the strain softening associated with classical fractures. These differences in mechanical evolution and structural expression may significantly influ-

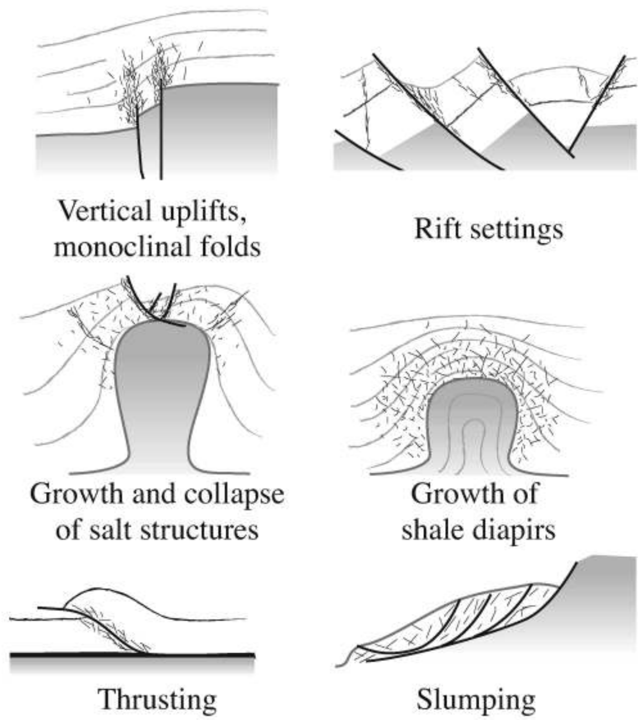

Fig. 2. Some different settings where deformation bands commonly develop: vertical uplifts and related monoclinal drape folds (Jamison & Stearns 1982); rift settings (Fisher & Knipe 2001); around salt structures (Antonellini *et al*. 1994); above shale diapirs, around thrusts and reverse faults (Cashman & Cashman 2000); glaciotectonic settings (Hooke & Iverson 1995); areas of gravity-driven collapse (Hesthammer & Fossen 1999).

ence fluid flow and therefore have direct implications for the management of the porous hydrocarbon and groundwater reservoirs in which they are very likely to occur.

Kinematically, deformation bands can be classified (Fig. 3) as dilation bands, shear bands, compaction bands or hybrids of these types (e.g. Aydin *et al*. 2006). The majority of deformation bands described in the geological literature are shear bands with attendant compaction (compactional shear bands) caused by grain reorganization with or without cataclasis. This compaction contributes to strain hardening and the creation of a localized band network or zone that precedes faulting (e.g. Schultz & Balasko 2003; Shipton & Cowie 2003). Early stages of shear band formation may also involve a component of dilation, and, although rare, dilational shear bands have been observed in experiments and in the field (Antonellini *et al*. 1994; Bésuelle 2001; Borja & Aydin 2004; Okubo & Schultz 2005). Pure compaction bands have been described in experiments and theory (Olsson 1999; Olsson & Holcomb 2000; Issen & Rudnicki 2001; Wong *et al*. 2001; Baud *et al*. 2004) and have been recognized in the field (Hill 1989; Mollema & Antonellini 1996; Sternlof *et al*. 2005). They are favoured in high-porosity (20–30%) coarse sand and sandstone (Mollema & Antonellini 1996) and have been reported to occur in the contractional (leading) quadrants of faults (e.g. Mollema & Antonellini 1996; Du Bernard *et al*. 2002*a*).

Classification of deformation bands by mechanisms

Although a kinematics-based classification (Fig. 3) is logical, it is also useful to classify deformation bands in terms of the dominant deformation mechanism operating during their formation (Fig. 4). Deformation mechanisms depend on internal and external conditions such as mineralogy, grain size, shape, sorting, cementation, porosity and stress state. Different mechanisms produce bands with different petrophysical properties. Thus, such a classification is particularly useful where permeability and fluid flow are an issue. The dominant deformation mechanisms are: (1) granular flow (grain boundary sliding and grain rotation); (2) cataclasis (grain fracturing and grinding or abrasion); (3) phyllosilicate smearing; (4) dissolution and cementation.

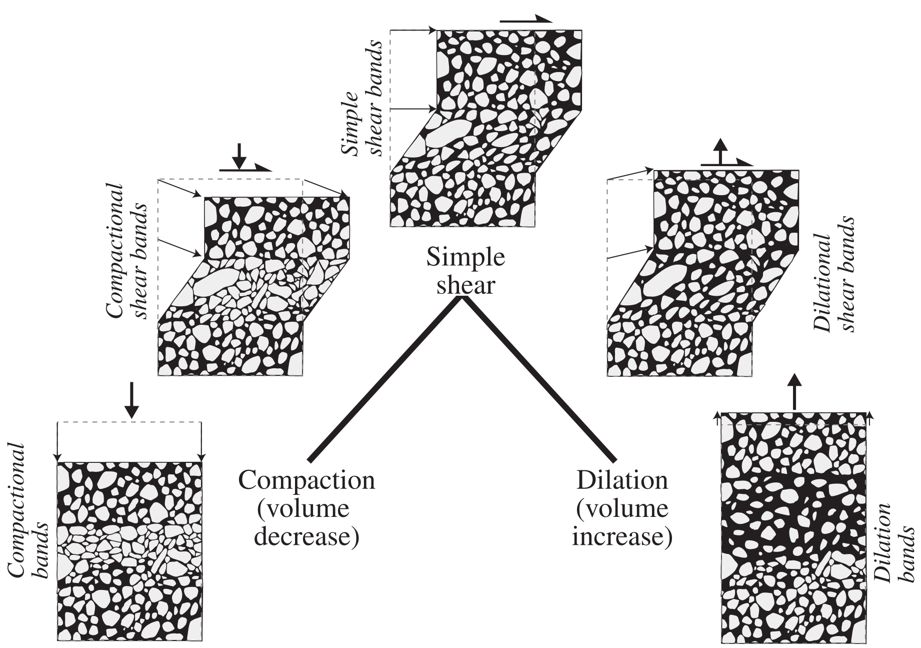

Fig. 3. Kinematic classification of deformation bands.

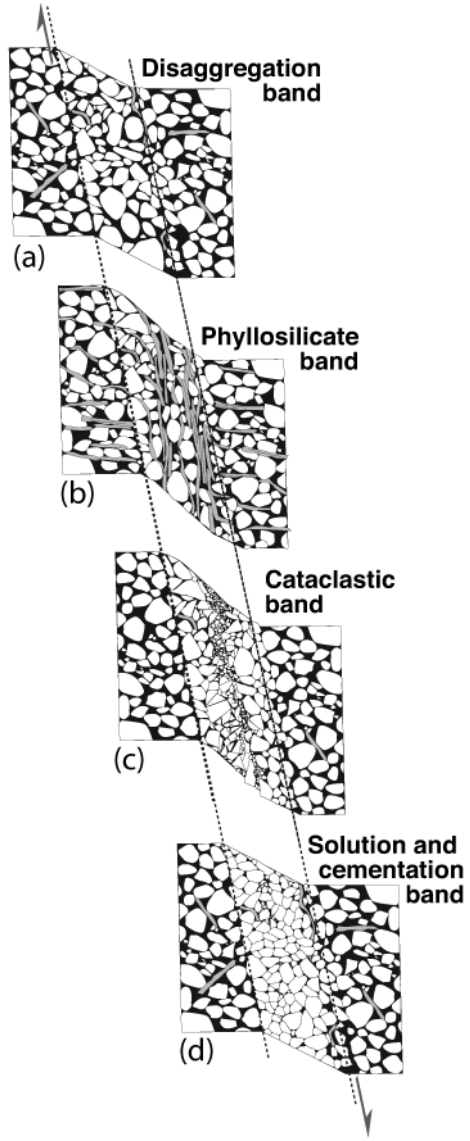

Fig. 4. The principal types of deformation bands, based on deformation mechanism.

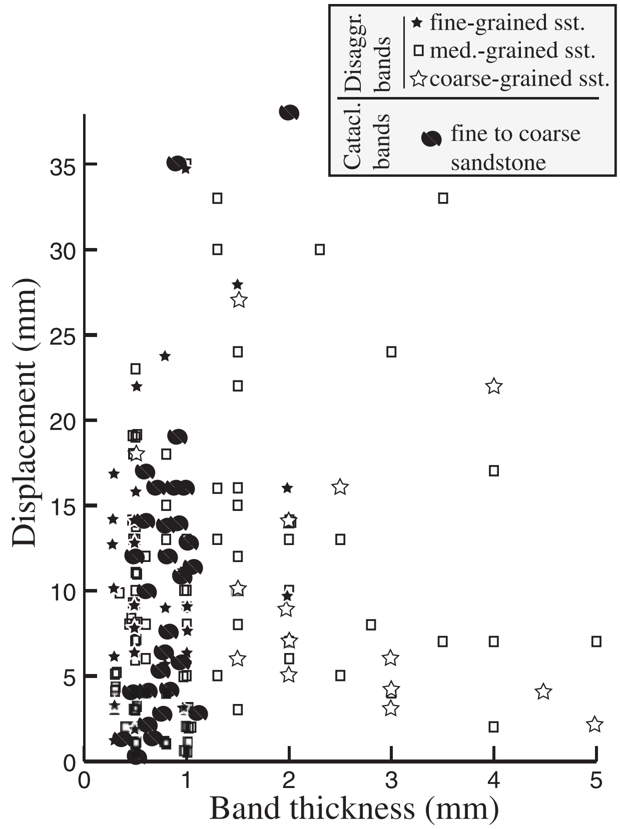

Fig. 5. Deformation band thickness v. displacement, plotted for different lithologies. Cataclastic deformation bands are from the Entrada Sandstone, San Rafael Desert, Utah. Data from disaggregation–phyllosilicate bands are from Jurassic sandstones in the Gullfaks Field reservoir, northern North Sea. It should be noted that fine-grained bands are thinner than coarse-grained bands.

Disaggregation bands

Disaggregation bands (Figs 1a and 4a) develop by shear-related disaggregation of grains by means of grain rolling, grain boundary sliding and breaking of grain bonding cements, a process referred to as granular flow (e.g. Twiss & Moores 1992) or particulate flow (e.g. Rawling & Goodwin 2003). They are commonly found in sands and poorly consolidated sandstones (Mandl *et al*. 1977; Du Bernard *et al*. 2002*a*; Bense *et al*. 2003), and form the 'faults' produced in sandbox experiments (e.g. McClay & Ellis 1987). Disaggregation bands can be almost invisible in homogeneous quartz sand(stone)s, but may be detected where they cross and offset laminae (Fig. 1a). Their true shear offsets are typically some centimetres, their lengths less than a few tens of metres, and their thicknesses vary with grain size of the host (Fig. 5). Fine-grained sand(stone)s develop bands *c*. 1 mm thick, whereas coarser-grained sand(stone)s host single bands that may be at least 5 mm thick.

Macroscopically, disaggregation bands are effectively ductile shear zones in the sense that sand laminae can typically be traced continuously through the band. The amount of shearing and compaction (pore-space collapse) that actually occurs along disaggregation bands depends on the nature and properties of the sandstone. Most pure and well-sorted quartz sand deposits are already compacted to the extent that the initial stages of shearing involves some dilation (Antonellini & Pollard 1995; Lothe *et al*. 2002), although continued shear-related grain reorganization may reduce the porosity at a later point.

Phyllosilicate bands

Phyllosilicate bands (framework phyllosilicate bands of Knipe *et al*. 1997) form in sand(stone) where the content of platy minerals exceeds 10–15%. They can be considered as a particular type of disaggregation band where platy minerals promote frictional grain boundary sliding (Fig. 1b) rather than grain fracturing (cataclasis).

Where clay is the dominant platy mineral, the clay minerals tend to mix with other mineral grains by a process referred to as deformation-induced mixing (Gibson 1998). The resulting bands are fine-grained, low-porosity zones (Fig. 1c) called deformation bands with clay smearing by Antonellini *et al*. (1994). Coarser phyllosilicate grains align to form a local fabric within the bands as a result of shear-induced rotation (Fig. 1b). Such phyllosilicate bands tend to show rotation of mica-rich laminae into the band.

In general, phyllosilicate bands can accumulate greater offsets than other types of deformation bands. This is due to the smearing of the platy minerals along phyllosilicate bands that counteracts strain hardening from interlocking of grains. They are easily detected, as the aligned phyllosilicates give the band a distinctive colour or fabric. An ordinary disaggregation band may transform into a phyllosilicate band where the phyllosilicate content of the rock increases (Fig. 1b).

If the clay content of the host rock is high enough (>40% according to Fisher & Knipe 2001), the structure becomes a clay smear. A clay smear is a continuous surface or thin zone of clay that forms by reorientation, flow, and/or extrusion of clay minerals. Striations seen on many clay smears indicate that they act (and should be classified) as slip surfaces rather than deformation bands. Field examples of cataclastic deformation bands becoming phyllosilicate bands or clay smears as they cross sandstone–siltstone boundaries are relatively common (Johansen & Fossen 2007).

Cataclastic bands

The classic cataclastic deformation bands described by Aydin (1978), Aydin & Johnson (1983) and Davis (1999) occur when mechanical grain fracture is a significant deformation mechanism. These bands consist of a central cataclastic core, commonly within a volume of compacted rock (Fig. 1d). The core is characterized by a wide grain-size distribution and high matrix content because of grain-size reduction, angular grains and a distinct absence of pore space. The surrounding volume is typically characterized by compaction (as a result of granular flow) and gentle fracture of grains. As pointed out by Aydin (1978), the crushing of grains during cataclasis results in extensive grain interlocking, promoting strain hardening. Strain hardening may explain the somewhat smaller displacements observed on cataclastic deformation bands (<3–4 cm), compared with disaggregation bands of similar lengths (Fig. 6).

Cataclastic bands are found in porous sandstones throughout the world; for example, in the Suez rift (Beach *et al.* 1999; Du Bernard *et al.* 2002*b*), France (Wibberley *et al.* 2000), the UK (Underhill & Woodcock 1987; Beach *et al.* 1997; Knott 1993), Ordovician sandstones of Oklahoma (Pittman 1981), Permian sandstones of the southern North Sea (Fisher & Knipe 2001) and southeastern Norway (Lothe *et al.* 2002), and the Jurassic sandstones of southwestern USA (e.g. Aydin 1978; Jamison & Stearns 1982; Davis 1999). They are mostly observed in rocks that have been buried to depths of 1.5–2.5 km; hence it is assumed that most cataclastic bands form at such depths; that is, after lithification but prior to uplift. Interestingly, cataclastic bands are also observed in unconsolidated or poorly consolidated sands in accretionary prism sediments (Lucas & Moore 1986; Karig & Lundberg 1990; Ujiie *et al.* 2004), Californian marine terrace sand (Cashman & Cashman 2000) and loose sandstones of the Rio Grande Rift (Heynekamp *et al.* 1999; Rawling & Goodwin 2003). Cataclastic bands have also been observed in non-welded ignimbrites and tuffs (Wilson *et al.* 2003), Rhine Graben and Roer Valley loess (Bense *et al.* 2003) and subglacial till (Hooke & Iverson 1995; van der Meer *et al.* 2003). Rawling & Goodwin (2003) suggested that cataclastic bands formed in sediments at shallow depths are characterized by grain spalling and flaking whereas the deeper cataclastic bands commonly exhibit transgranular fracturing and grain crushing. However, Cashman & Cashman (2000) showed that cataclastic bands

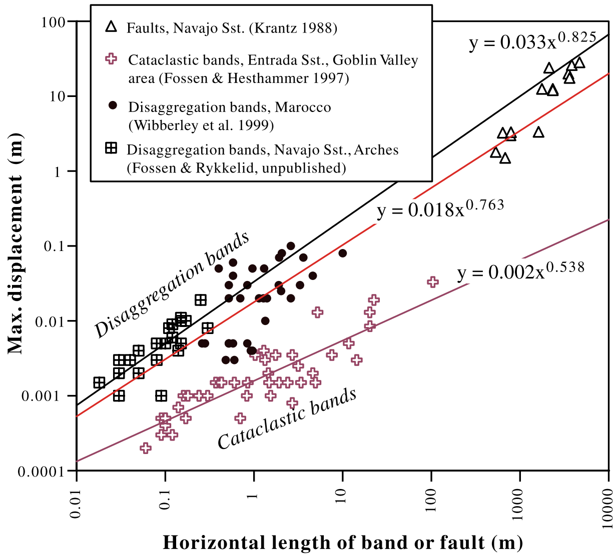

Fig. 6. Displacement–length relationship for cataclastic deformation bands and disaggregation bands. The two categories of deformation bands occupy different fields of the diagram, and define different semi-linear trends; cataclastic bands are much longer than disaggregation bands with respect to offset.

formed in unconsolidated marine sand buried no deeper than 50 m exhibit grain crushing.

Dissolution, cementation and diagenesis

Dissolution and cementation may occur preferentially along a deformation band during or, more commonly, after deformation. If solution, also referred to as chemical compaction or pressure solution, is significant the term 'solution band' is warranted. Solution bands (Gibson 1998) typically consist of tightly packed grains smaller in size than the matrix, but showing little evidence of cataclasis. Although quartz dissolution accelerates at >90 °C (Walderhaug 1996; i.e. depths greater than *c*. 3 km), dissolution is a common feature of deformation bands formed at shallower depths. Whereas dissolution is promoted by clay minerals on grain boundaries, cementation in deformation bands is promoted by fresh and highly reactive surfaces formed during grain crushing and/or grain boundary sliding. Cementation is particularly pronounced in deformation bands where undeformed host sand grains are coated by diagenetic minerals such as chlorite (Ehrenberg 1993) and illite (Storvoll *et al.* 2002). The coating prevents cementation except in the deformation bands, where the coating is broken by fracturing and sliding to expose fresh quartz surfaces (Leveille *et al.* 1997; Hesthammer *et al.* 2002). Cementation may also be promoted by localized tensile fracture in the centre of a deformation band (Gabrielsen & Koestler 1987; Leveille *et al.* 1997) and subsequent precipitation of minerals such as calcite, anhydrite, salt, hydroxides and quartz. Fisher & Knipe (2001) suggested such cementation to be discontinuous and restricted to extensional jogs in deformation band samples from the southern North Sea.

Petrophysical properties

Permeability measurements across deformation bands (Fig. 7) have led many workers to conclude that deformation bands

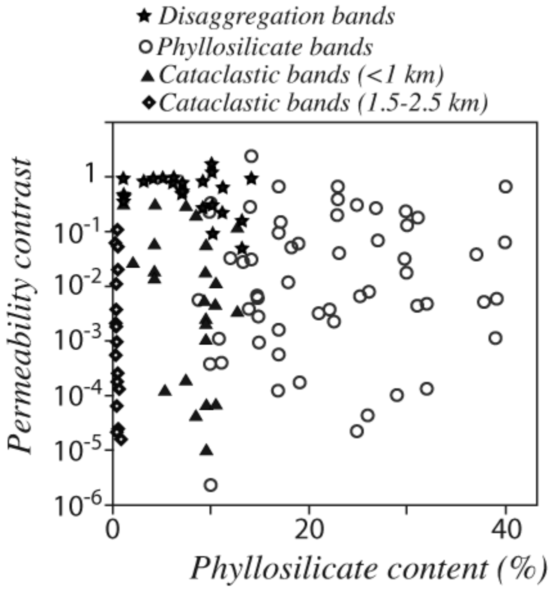

Fig. 7. Permeability data (plug measurements) for deformation bands. Data from Fowles & Burley (1994), Sigda *et al*. (1999), Fisher & Knipe (2001) and Lothe *et al*. (2002).

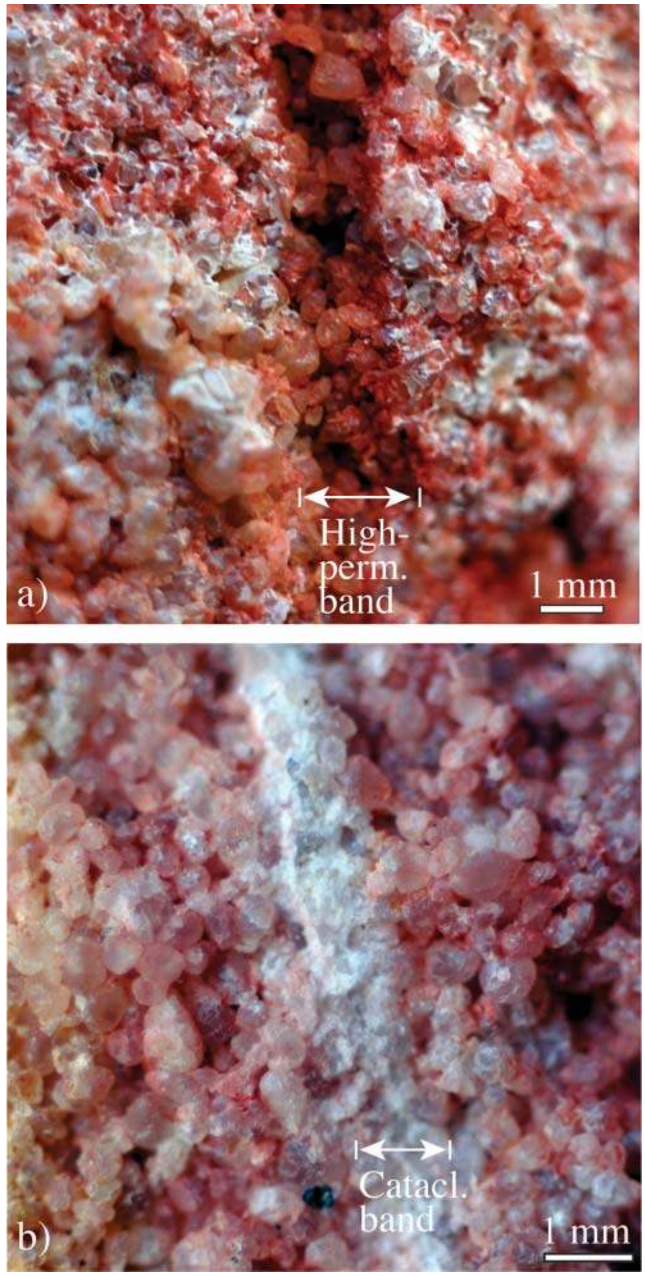

Fig. 8. Two deformation bands in the same layer of the Nubian Sandstone, Tayiba Red Beds, Sinai. (**a**) Disaggregation (dilation) band where the porosity is higher than that of the host rock. (**b**) Cataclastic band showing considerable porosity collapse. Whereas the cataclastic deformation band represents a low-permeability structure in the sandstone, the disaggregation band represents a conduit for fluids.

reduce transmissibility in a reservoir. However, there are other cases where deformation bands appear to be conduits for fluids (e.g. Parry *et al*. 2004; Sample *et al*. 2006). The influence of deformation bands on fluid flow depends on their internal permeability relative to the surrounding rock. Two extreme cases are shown in Figure 8, the first representing a deformation band with a higher porosity and permeability than the host rock (Fig. 8a), the second with considerably lower porosity and permeability than the host (Fig. 8b). In these two examples the mineralogy is almost identical, and the difference lies in deformation mechanism: disaggregation with little grain fracture in the first case, compared with intense cataclasis in the second (Fig. 8b). We now highlight the influence of specific deformation mechanisms on the resulting petrophysical properties of deformation bands.

Disaggregation bands

Disaggregation bands can result in an enhancement or reduction of porosity depending on whether they have a dilational or compactional component. Du Bernard *et al*. (2002*a*) reported that their pure dilation bands represent an increase of porosity of 7%, although the pore space has later been filled with clay-rich cement in this case. Antonellini *et al*. (1994) found a similar figure of 8% porosity increase in dilatant shear bands in Arches National Park. Mollema & Antonellini (1996) reported that compaction bands reduced the porosity from 25% in the host rock to less than a few per cent in the compaction band. Du Bernard *et al*. (2002*a*) suggested that the increased porosity of dilation bands should be transient, because of the increased infiltration of clays into the enhanced pore network. These observations agree with field evidence for preferred fluid flow along dilation bands, as reported by Bense *et al*. (2003) and Sample *et al*. (2006). Other disaggregation bands may be less porous and permeable than the host rock. A permeability reduction of up to one order of magnitude has been observed in phyllosilicate-bearing sandstones (Fisher & Knipe 2001). However, most of these porosity and permeability contrasts are relatively low, and disaggregation bands generally have little influence on the permeability of sandstone reservoirs (Fig. 7).

Phyllosilicate bands

Phyllosilicate bands typically reduce permeability by an amount depending on phyllosilicate abundance, phyllosilicate type, phyllosilicate distribution, displacement along the band, and grain size (Knipe 1992). On average, the reduction in permeability for North Sea reservoirs is around two orders of magnitude but can be up to five orders of magnitude where the phyllosilicate grains are small (<0–5 μm; Fisher & Knipe 2001). The reduction is caused mainly by mixing and alignment of platy minerals, and depends on the specific arrangement of platy minerals and thus the shear strain. Typically, these factors, and therefore also permeability, vary along the deformation bands, depending on the local source of phyllosilicates. Hence, the effective influence of phyllosilicate bands on fluid flow is controlled by the points of lowest and highest permeability. Estimates of permeability reduction associated with phyllosilicate bands from core plugs may therefore incorrectly reflect their effective influence on fluid flow during production of a hydrocarbon reservoir.

Cataclastic bands

The majority of published studies of petrophysical properties in deformation bands have focused on cataclastic deformation bands dominated by shear deformation with or without additional compaction. The porosity of cataclastic deformation bands is reduced by up to an order of magnitude by grain crushing and resulting change in grain-size distribution. The reduction of porosity produces a corresponding decrease in permeability of two to three, and locally as much as six, orders of magnitude with respect to the host rock (Pittman 1981; Jamison & Stearns 1982; Harper & Moftah 1985; Knott 1993; Antonellini & Aydin 1994; Gibson 1994, 1998; Knipe *et al*. 1997; Crawford 1998; Antonellini *et al*. 1999; Fisher & Knipe 2001; Jourde *et al*. 2002; Shipton *et al*. 2002). The very low (<1%) porosity core of some well-developed cataclastic shear bands results in permeabilities as low as 0.001 mD (Freeman 1990; Antonellini & Aydin 1994; Knipe *et al*. 1997; Fisher & Knipe 2001; Shipton *et al*. 2002). Cataclastic compaction bands produced experimentally by Holcomb & Olsson (2003) showed a reduction in permeability of around two orders of magnitude.

The effect of dissolution and cementation

Cementation and dissolution in deformation bands may significantly increase the reduction of porosity and permeability caused by mechanical crushing and reorganization of grains (Ngwenya *et al*. 2000; Ogilvie & Glover 2001). A transient increase in permeability occurred in the experiments of Main *et al*. (2000), associated with initial dilation (e.g. Mandl *et al*. 1977; Bernabe & Brace 1990). This provides a way for fluids to enter the deformation band, and the entrance of reducing fluids at this stage offers an explanation of the bleaching of deformation bands (Parry *et al*. 2004). A similar mechanism may explain cementation within low-porosity cataclastic deformation bands (Fowles & Burley 1994; Labaume & Moretti 2001; Parnell *et al*. 2004; Sample *et al*. 2006).

Cementation probably occurs after, rather than during, the formation of deformation bands, and the solution and precipitation of quartz accelerates after burial and heating to above *c*. 90 °C (Walderhaug 1996). Fisher & Knipe (2001) reported a general decrease of permeability with depth for cataclastic deformation bands that have experienced post-deformational burial in the southern North Sea. As a result of the enhanced chemical reactivity of fresh broken or abraded grain surfaces in the cataclastic bands, precipitates probably include efficient permeability-reducing clay minerals in addition to quartz. Precipitation of secondary minerals such as carbonates and anhydrite along fractures during deformation band reactivation has also been described. Fisher & Knipe (2001) cautioned, however, that the discontinuous nature of many such cements in the North Sea reservoirs makes them a less significant influence on fluid flow than may be expected from thin-section or hand-sample investigations. An additional effect of quartz cementation is to lower porosity and increase the strength of the host rock such that subsequent deformation may lead to the development of ordinary fractures that actually represent fluid-flow conduits.

Effect on fluid flow

It has been shown that the majority of deformation bands show some reduction in permeability, some by as much as several orders of magnitude. However, their practical effect on fluid flow is not clear. For single-phase flow (i.e. water flowing in a water-saturated rock or oil flowing in an oil-saturated rock), the thickness and permeability of the deformation band zone are the controlling factors on fluid flow (Darcy flow). Simple numerical analyses demonstrate that the number of deformation bands (i.e. thickness of the zone) and/or the permeability reduction must be significant for deformation bands to seriously effect fluid flow (Matthai *et al*. 1998; Walsh *et al*. 1998). Nevertheless, complex zones of deformation bands have been blamed for reduced productivity in some oil wells (e.g. Harper & Moftah 1984).

For two-phase flow (i.e. oil flowing through a water-saturated rock, or groundwater flowing through the vadose zone), capillary pressure becomes relevant. In hydrocarbon reservoirs the capillary threshold pressure of the fault rock determines how much oil can accumulate on one side of the fault before across-fault migration occurs. Calculations predict that deformation bands cannot hold much more than a 20 m (Harper & Lundin 1997) or perhaps up to 75 m (Gibson 1998) high column of hydrocarbons.

Regardless of whether one- or two-phase flow is considered, the practical consequence of deformation bands depends on other factors than permeability contrasts. In particular, their continuity or variation in thickness and permeability in three dimensions is critical. Field observations of deformation bands indicate that their thickness and porosity change significantly even along single bands. The same is the case with deformation band clusters. Clearly, the weakest point of the deformation band network influences its effect on flow. The physical connectivity of bands is a related factor, and they both undermine the effect of deformation bands as sealing and flow-reducing structures. However, the presence of deformation bands and deformation band zones may still change the flow pattern if they have a preferred orientation. Sigda *et al*. (1999) observed that low-porosity deformation bands can act as preferential groundwater flow paths through the vadose zone. Similar channelization can be visualized during production of a petroleum reservoir. During oil production stimulated by water injection, pockets of residual oil may also remain in 'shadow zones' as a result of capillary trapping (Manzocchi *et al*. 2002). This effect should be considered during planning of wells and simulation of oilfields where low-permeability deformation bands are a concern.

Formation conditions of deformation bands

Given the range of deformation band characteristics and their influence on fluid flow, considerable attention has been devoted to understanding the conditions that control their formation. A number of factors are important, including confining pressure (burial depth), deviatoric stress (tectonic environment), pore fluid pressure and host rock properties, such as degree of lithification, mineralogy, grain size, sorting, and grain shape.

Some of these intrinsic host rock properties are approximately constant for a given sedimentary rock layer. However, they may vary dramatically from one layer to another, resulting in rapid changes in deformation band style across lithological boundaries.

Factors such as porosity, permeability, confining pressure, stress state and cementation are likely to change with time; hence deformation bands may record a temporal evolution associated with, for instance, increasing burial depths. The temporal sequence of deformation structures in a given rock is an important geological signature that reflects the physical changes experienced during burial, lithification and uplift.

Temporal sequence of deformation in sandstones

The earliest forming deformation bands in sandstones are typically disaggregation bands (Fig. 9). These structures form at low confining pressures (i.e. shallow burial) when forces acting across grain contact surfaces are low and grain bindings are weak (Fig. 10). Early disaggregation bands are often related to local, non-tectonic gravity-controlled deformation, such as local shale diapirism, underlying salt movement, gravitational sliding and glaciotectonics (e.g. Antonellini *et al*. 1994).

Cataclastic deformation bands can form in poorly consolidated sands at <1 km burial depths (e.g. Lucas & Moore 1986; Cashman & Cashman 2000; Rawling & Goodwin 2003). Shallow cataclasis is promoted where well-sorted and well-rounded grains lead to high grain-contact stress, and in feldspar and other minerals that have well-developed crystallographic cleavage (e.g. Zhang *et al*. 1990). In general, shallowly formed cataclastic bands show less intense cataclasis than those formed at greater (1–3 km) depths. Abundant examples of cataclastic deformation bands found in the Jurassic sandstones of the Colorado Plateau (Fig. 1d) highlight a clear temporal evolution from early disaggregation bands to later cataclastic bands (e.g. Antonellini *et al*. 1994). Phyllosilicate bands may form at a variety of depths if enough (*c*. 15% or more) phyllosilicates are present.

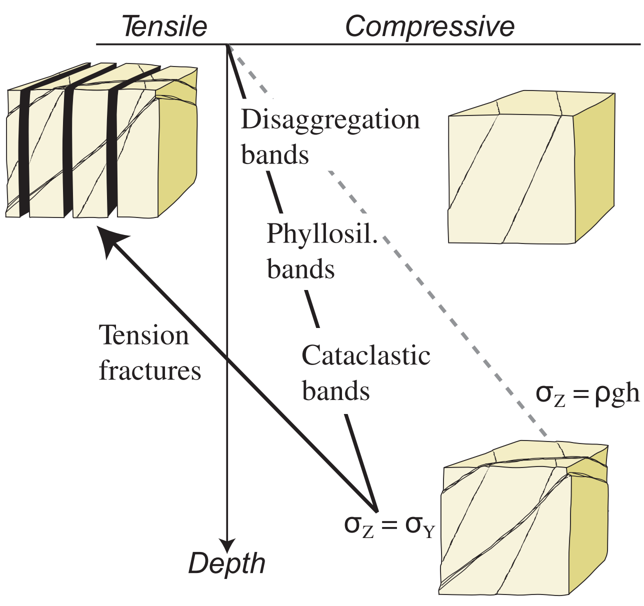

Fig. 9. Theoretical stress history for a simple burial and uplift history of sandstones (Engelder 1993) in relation to structural development. In contrast to shales, sandstones enter the tensile regime during uplift, and tension fractures (joints) form.

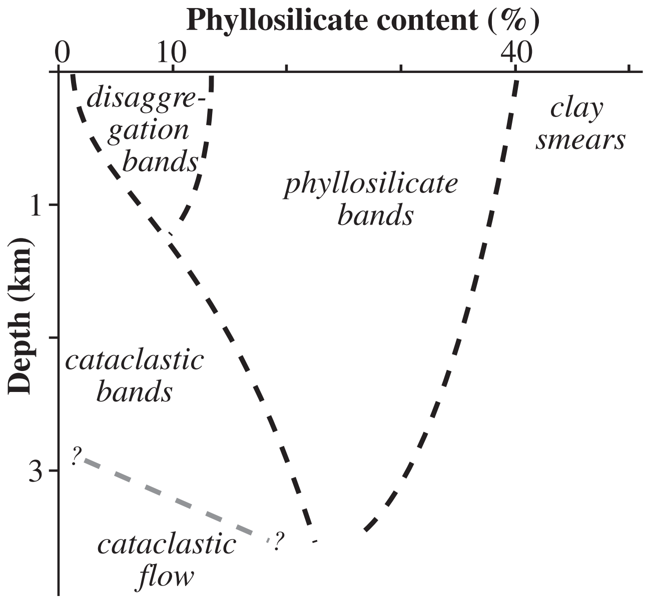

Fig. 10. Schematic illustration of how the different deformation band types relate to phyllosilicate content and depth. Many other factors influence on the boundaries outlined in this diagram, and we do not know in any great detail how they interact. In addition, the transitions are gradual, so the boundaries drawn should be considered as uncertain.

Once a rock becomes a cohesive lithology with reduced porosity, deformation tends to occur by crack propagation instead of pore space collapse, and slip surfaces, joints and mineral-filled fractures can form. Slip surfaces associated with deformation bands probably form as a result of significant porosity reduction within many deformation band zones. Joints and veins can also form in sandstones that have lost porosity because of lithification and quartz cementation. Thus, slip surfaces, joints and veins almost invariably postdate both disaggregation bands and cataclastic bands in sandstones. Because quartz cementation-related porosity reduction may vary locally, deformation bands and joints may develop simultaneously in different parts of a sandstone, but locally the temporal sequence is: (1) deformation bands; (2) faulted deformation bands (i.e. slip surface); (3) jointing; (4) reactivation of joints as faults (e.g. Johansen *et al*. 2005).

Lithification and loss of porosity are not the only reasons why extension fractures occur as late structures in most exposed porous sandstones. Such sandstone sequences tend to portray regional joint sets influenced by removal of overburden and related cooling during regional uplift. Although clearly important, such features are unlikely to be developed in subsurface petroleum reservoirs unless they have been significantly uplifted. Thus, knowing the burial and uplift history of a basin in relation to the timing of deformation events is very useful when considering the type of small-scale structures present in, say, a sandstone reservoir. Conversely, examination of the type of deformation structure present also gives information about deformation depth and other conditions at the time of deformation.

Sensitivity to lithological variations

Field observations of deformation bands crossing lithological contacts in layered sedimentary sequences provide important information on lithology control on deformation band style. We assume that a single deformation band that crosses adjacent layers formed at approximately the same geological time, depth and stress conditions in all layers. Field observations clearly reveal that the transformation of a single structure from a cataclastic or disaggregation band into a phyllosilicate band coincides with the transition from well-sorted sandstone to poorly sorted sandstone with a higher percentage of phyllosilicates. Another common observation is that cataclastic bands change into disaggregation bands as they enter more fine-grained and poorly sorted sandstones. Such transition clearly depends on factors such as porosity, mineralogy and grain size. If the contrast is high enough, the deformation band may actually terminate at the contact between two layers or transform into a slip surface (Schultz & Fossen 2002).

These observations are consistent with experimental work indicating that a decrease in porosity and grain size would inhibit cataclasis in sandstone (e.g. Chuhan *et al*. 2002). Laboratory and field observations suggest that the presence of silica cement promotes microfracturing and finer-grained cataclastic deformation bands (Johansen *et al*. 2005). However, the presence of a hematite grain coating (Main *et al*. 2001) encourages grain boundary sliding, thus favouring disaggregation bands rather than cataclastic deformation bands. Primary grain mineralogy will also play a role, and cataclasis is observed to be more intense in lithic or feldspathic sand than in quartz sand (e.g. Chuhan *et al*. 2002; Rawling & Goodwin 2003). Flodin *et al*. (2003) argued that porosity is a primary control on deformation band structure, where increased porosity leads to high grain contact stress, and thus favours the formation of cataclastic deformation bands.

For a subsurface petroleum or groundwater reservoir, the ability to make an accurate prediction of deformation structures and their permeability characteristics at various stratigraphic levels from a basic input of lithology and burial history is highly desirable. Although important advances have been made in this direction, more experimental and field-based work is required to properly understand the coupling between the many factors that control deformation in sandstones and other porous media.

The connection between deformation bands, faults and damage zones

Field data show that deformation bands occur as isolated structures, linked systems, complex zones of multiple, interconnected deformation bands, and in fault damage zones (Fig. 11) (e.g. Aydin & Johnson 1983; Hesthammer & Fossen 2001). Laboratory observations (Mair *et al*. 2000) have confirmed field-based predictions that the number of distinct deformation bands increase with increasing strain. Detailed mapping of outcrops of faulted cataclastic deformation bands shows that slip surfaces tend to nucleate in small patches in deformation band zones that propagate, link up, and ultimately form through-going slip surfaces with accumulated strain (Shipton & Cowie 2001). Mature, through-going slip surfaces are commonly associated with a thin (millimetre thick) core of ultracataclasite (e.g. Aydin & Johnson 1978; Shipton & Cowie 2001). Intense localized grain crushing also occurs within zones of deformation bands prior to slip-surface development (Shipton & Cowie 2001, fig. 11; Johansen *et al*. 2005, fig. 9), suggesting that grain crushing is an incipient stage in the formation of slip surfaces.

The number of deformation bands formed locally at the time of slip-surface formation is probably sensitive to several factors, including porosity, grain size, cement, mineralogy and overburden stress (i.e. depth). Small-scale (5–20 m throw) faults in fluvial to shallow marine North Sea sands deformed at <1 km depth typically exhibit 10–15 deformation bands on either side of the slip surface (Hesthammer & Fossen 2001), whereas small faults in aeolian sandstones deformed at *c*. 2 km depth may have 50–100 bands or more (Aydin 1978). This indicates that more substantial fault damage zones form at greater burial depths (Mair *et al*. 2002*a*). The influence of host rock lithologies is demonstrated in the Moab area of Utah, where the Navajo Sandstone develops considerably more deformation bands than the Entrada Sandstone for a given strain. This relationship is also seen on bed-scale, where deformation band frequency may vary dramatically from bed to bed (Fig. 12), depending on the lithological factors discussed above. Preliminary field data suggest that high-porosity, well-sorted sandstones develop the widest damage zones around minor faults.

The length of the deformation band process zone ahead of a fault tip also varies depending on lithology. This zone is most extensive (up to >100 m) in well-sorted and highly porous sandstones such as the Entrada and Navajo Sandstone (Shipton & Cowie 2001; Rotevatn *et al*. 2007). Such deformation band process zones (Fig. 11) may therefore influence fluid flow in regions ahead of seismically mapped fault tips. Given that fault offsets less than 10–20 m are not resolved in commercial seismic surveys, it is common to use displacement–length scaling relations to extend seismically resolvable fault tips (Pickering *et al*. 1997). The presence of deformation band process zones should also be included in this type of analysis (although see below for a discussion of displacement–length scaling).

Once a continuous slip surface has formed, strain accumulates predominantly by frictional sliding. If subsequent fault growth and strain accommodation was dominated by strain softening, then damage zone thickness should be independent of fault displacement. However, in many cases, large faults appear to have wider damage zones than small faults (Fig. 13), suggesting that damage zones are still active during localized fault slip (Shipton & Cowie 2003). This may be caused by fault locking as a result of non-planar or interfering slip surfaces (Rykkelid & Fossen 2002). Therefore, structural elements in damage zones around deformation band faults may be both remnants from the pre-faulting stage as well as syn-faulting damage (Schultz & Siddharthan 2005).

The orientations of deformation bands in damage zones will clearly influence the permeability structure. Conjugate sets of deformation bands are common in places such as the Colorado Plateau (e.g. Berg & Skar 2005; Fossen *et al*. 2005), the Permian basins of the southern North Sea–UK area (e.g. Fowles & Burley 1994) and North Sea Middle Jurassic reservoirs (Hesthammer *et al*. 2000), with one set subparallel to the main slip plane and the other dipping in the opposite direction. Mutual cross-cutting relationships show that these conjugate sets form contemporaneously (Zhao & Johnson 1991; Olsson *et al*. 2004). It is anticipated that fluid flow parallel to the strike of these conjugate bands would be easier than flow across damage zones. In detail, complications in damage zone structure often increase at fault branch points or stepovers (Antonellini & Aydin 1995; Tindall & Davis 1999; Johansen *et al*. 2005). The complex variation of deformation band geometry in damage zones has the potential to influence flow in a complicated manner.

Deformation band mechanics

A considerable amount of theoretical work has been carried out on the development of deformation bands in rocks (e.g. Rudnicki & Rice 1975; Rudnicki 1977; Aydin & Johnson 1983; Issen &

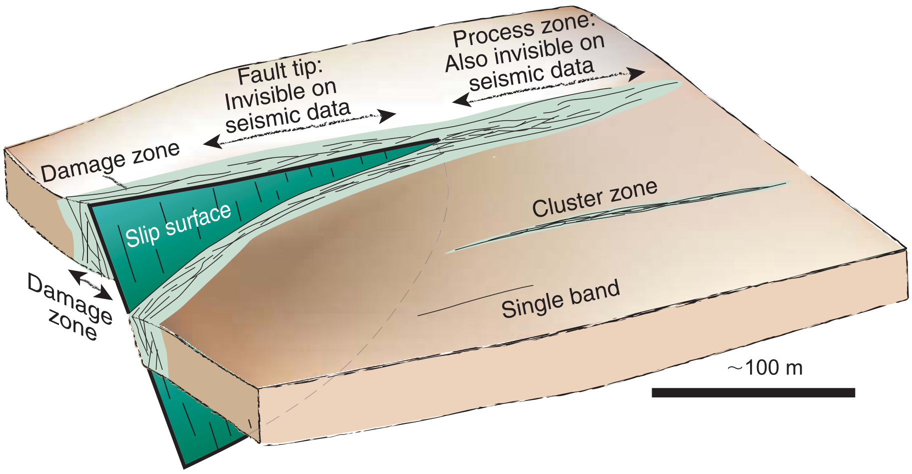

Fig. 11. Fault tip in porous sandstone. Deformation bands are formed in the 'process zone' ahead of the fault tip. The length of the zone varies from lithology to lithology, but can be up to several hundred metres and thus should be considered when evaluating fluid flow in many faulted hydrocarbon reservoirs.

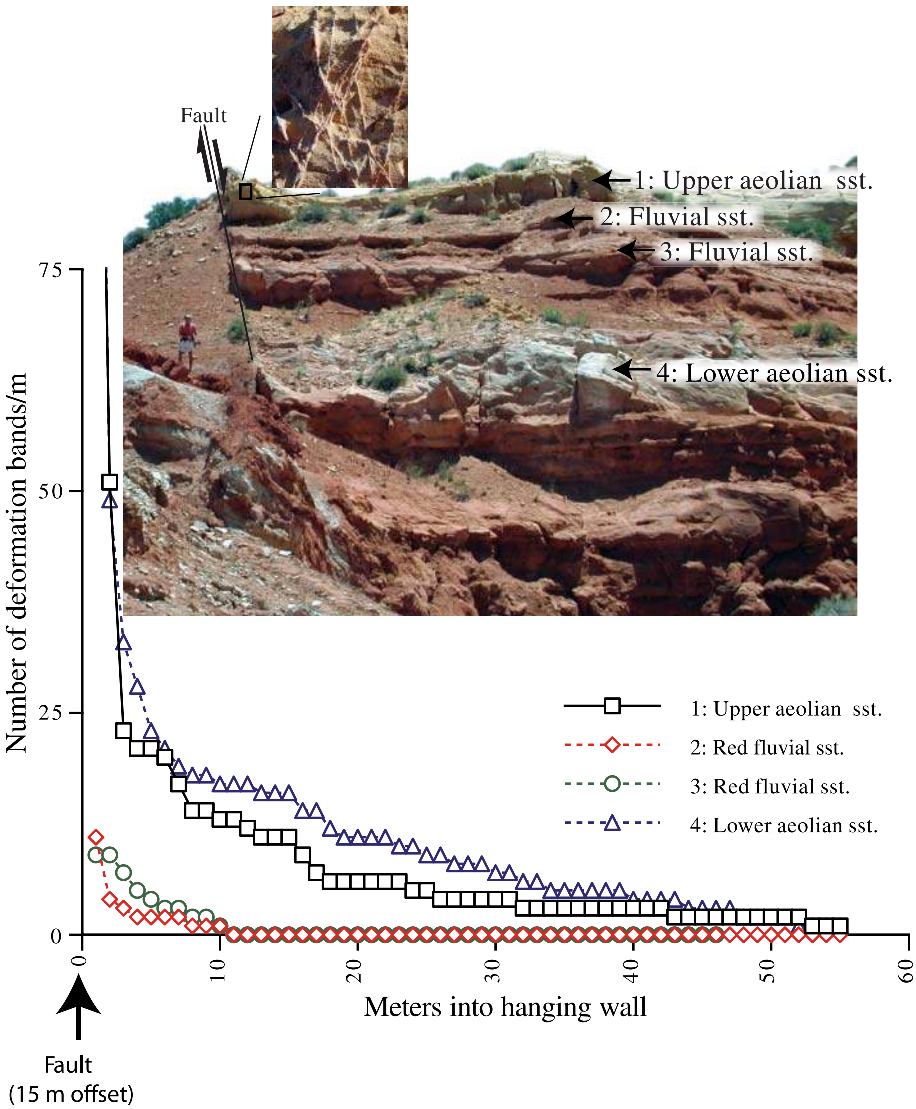

Fig. 12. Characterization of the damage zone for a minor fault in layered sand–siltstones of the Entrada Sandstone (San Rafael Desert). Band frequency varies from sandstone layer to sandstone layer, and is highest in the well-sorted, aeolian layers (1 and 4).

Rudnicki 2000; Borja & Aydin 2004), providing a firm foundation for understanding these structures. An approach called the 'Cam cap' model of yielding and band formation is now widely used (Wong *et al.* 1992, 2004; Borja & Aydin 2004; Schultz & Siddharthan 2005; Aydin *et al.* 2006). This approach provides a consistent framework for understanding the development of deformation bands, damage zones, and attendant faulting.

The model is best described using a q–p stress diagram (Fig.

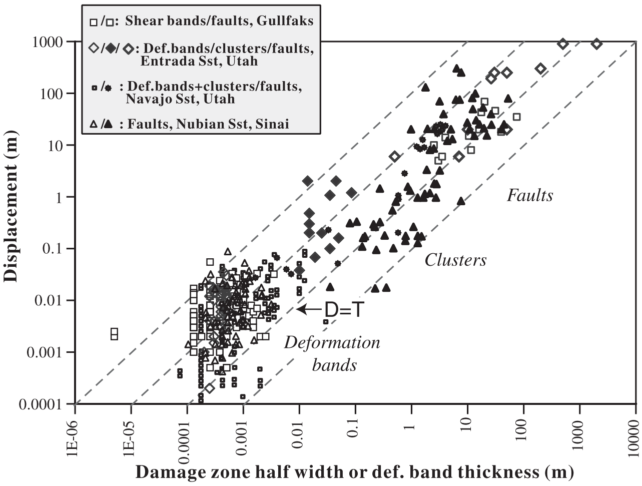

Fig. 13. Half-width of damage zone plotted against displacement in log–log space for faults, together with thickness–displacement data for deformation band zones and individual deformation bands in some porous sandstones. Gullfaks data from Fossen & Hesthammer (2000), Entrada data from Fossen (unpubl.), Navajo data from Shipton & Cowie (2001) and Fossen (unpubl.), and Sinai data from Beach *et al*. (1999) and Rotevatn & Fossen (unpubl.)

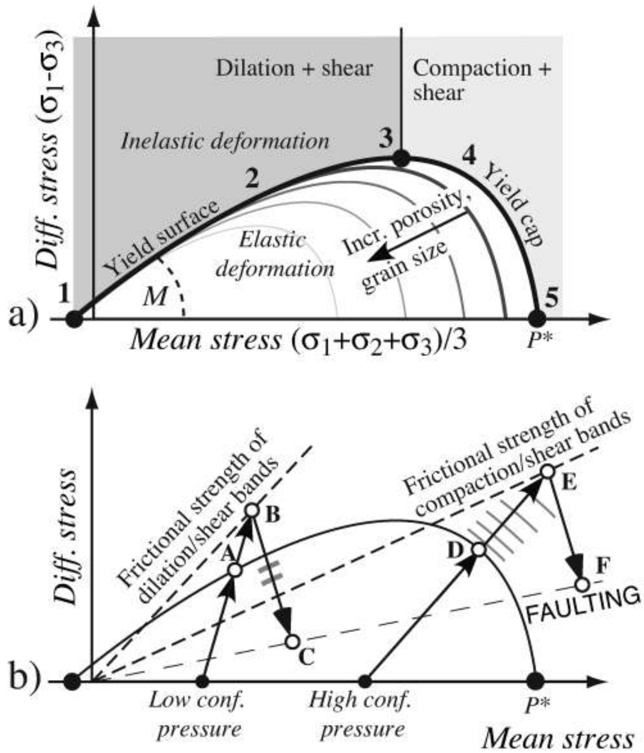

Fig. 14. The q–p diagram (differential stress–mean stress) applied to porous rocks (after Schultz & Siddharthan 2005). (**a**) Inelastic yielding of the host rock (bold curve) produces dilatant bands (1), dilatant shear bands (2), shear bands (3), compactional shear bands (4), and compaction bands (5). The yield surface or cap depends on porosity and grain size as indicated. (**b**) Localization of deformation bands for a compressive remote stress state. Stress paths A–C are for dilatant shear bands, and D–F for compactional (i.e. cataclastic) shear bands. Dashed grey lines in (**b**) depict moving yield surfaces in the direction of the frictional sliding line, leading to faulted deformation bands.

14), where the coordinate axes q and p represent the differential stress and the mean stress, respectively. The yield surface (Fig. 14) separates the elastic (recoverable strain) from the inelastic (deformation-band-forming) regimes. Its shape depends on the physical properties of the deforming rock (e.g. Wong *et al*. 1992; Issen & Rudnicki 2000; Mair *et al*. 2002*a*; Borja 2004; Borja & Aydin 2004). The type of deformation band that forms will depend on the state of stress at the moment of inelastic yielding; that is, on the point of intersection between the loading path and the yield surface. For example, dilatant shear bands are formed at relatively low confining pressures (segment 2 of Fig. 14a) whereas compactional shear bands are formed at higher confining pressures (segment 4). The critical pressure, P^* (point 5 in Fig. 14a), is the pressure at which compaction occurs in the absence of shearing. This value scales approximately with the product of grain size and porosity (Zhang *et al*. 1990; Wong *et al*. 1997) such that as grain size and/or porosity increase, the critical pressure and the yield surface decrease (Fig. 14a). Thus unconsolidated sand can form compaction bands at relatively shallow depth whereas consolidated sandstone requires much higher confining pressures (Mair *et al*. 2002*a*).

As an illustration, compactional shear bands (path D–E in Fig. 14b) form at relatively high confining pressure when differential stress increases and the rock begins to compact and shear. As this occurs, the grain-to-grain contacts experience a much larger compressive stress, eventually promoting grain crushing and fracturing (Zhang *et al*. 1990). Grain crushing leads to (1) reduced average grain size within a growing band, (2) a tighter packing geometry, (3) increased grain angularity, and consequently increased shear resistance (Mair *et al*. 2002*b*). These factors inhibit shearing displacements within a cataclastic band, resulting in strain hardening. However, because grain crushing depends on several factors including mineralogy, grain size, packing geometry, grain composition, cementation (lithification) and shape (Wong *et al*. 1997; Wong & Baud 1999), the same stress state may produce compaction with or without cataclasis in different sandstones. Strain hardening moves the rock off the yield cap and upward until the stress state associated with frictional sliding along some sections of the band array is achieved (Fig. 14b, path D–E). At this point, slip surfaces nucleate and grow throughout the band network (Aydin & Johnson 1978; Shipton & Cowie 2001, 2003), and the band 'fails' unstably (Aydin & Johnson 1983), forming a faulted compactional shear band array (Schultz & Siddharthan 2005).

Scaling relationships

Displacement–length relationships

Deformation band displacement profiles are qualitatively similar to those of faults; for example, both develop an along-strike displacement profile with a central maximum (e.g. Fossen & Hesthammer 1997, 1998). However, some distinct differences between faults and deformation bands are worth examining.

Faults with length in excess of *c*. 100 m commonly exhibit a power law relationship between length (L) and maximum displacement (D) that can be expressed as (e.g. Cowie & Scholz 1992; Clark & Cox 1996)

$$D = cL^n$$

where c is a constant. The exponent (n) has a value of approximately one across nine orders of magnitude for faults as small as tens of centimetres long (Clark & Cox 1996; Schlische *et al*. 1996) but was found to be *c*. 0.5 for a population of cataclastic deformation bands in Utah (Fig. 6). Wibberley *et al*. (2000) also reported unusually small *D*/*L* ratios for cataclastic deformation bands in porous Cretaceous sandstone in Orange, France, indicating that cataclastic deformation bands have less displacement per unit length than faults.

Wibberley *et al*. (2000) explained their *D*/*L* observations by invoking the high frictional strength of cataclastic deformation bands in porous sandstones; they estimated that porous sandstones have lower shear moduli than other coarse-grained sedimentary rocks. Although this can certainly account for the smaller values of *D*/*L*, it cannot alone account for a reduction in slope from 1.0 to 0.5 observed for cataclastic deformation bands.

Schultz & Fossen (2002) attributed the anomalously low *D*/*L* ratios for Entrada Sandstone deformation bands to lithological layering, and suggested that the deformation bands that nucleate in the sandstone layer have difficulties propagating into the adjacent low-porosity, silty layers. They suggested that once a growing deformation band spans the thickness of the sandstone layer, it keeps growing horizontally until a through-going slip surface forms and cuts through the stratigraphy. The abundant field evidence indicating that deformation bands are sensitive to lithology and grow selectively in layers of high porosity supports this hypothesis.

Disaggregation bands exhibit *D*/*L* scaling similar to ordinary faults and slip surfaces (Fig. 6). Wibberley *et al*. (1999) related this observation to ‘slow’ tip propagation that exploits the reduced shear modulus in unlithified sand compared with sandstone. In this hypothesis, ‘slow’ implies that strain hardening in cataclastic bands may promote tip propagation rather than continued accumulation of offset in the central portion of the band. The result would be long cataclastic bands with small *D*/*L* ratios. For unlithified sand, strain hardening appears to be subordinate or absent, and strain accumulates more readily through continued shearing along the band (large *D*/*L* ratios).

Conclusions

The distinct style of deformation band faulting observed in porous rocks, where faulting is preceded by a history of deformation band formation and accumulation, has several implications. First, the resulting fault is contained in a volume of rock containing deformation bands (the damage zone) that add to the flow-reducing property of the fault. The thickness of this deformation band zone, and therefore their effect on fluid flow, scales positively with fault displacement. Second, because the damage zone extends beyond the tip line of the fault by up to several hundred metres, this potentially lengthens the region that may perturb fluid flow. A third implication is that fault orientation (strike and dip) is influenced by the orientation of the deformation band zone in which it grows (Johnson 1995; Johansen *et al*. 2005; Okubo & Schultz 2006).

The effectiveness of deformation bands as fluid flow barriers or baffles depends only partly on how their internal petrophysical properties are altered relative to the surrounding rock, which is primarily dictated by deformation mechanisms and mineralogy. Other important factors are the number of bands (collective thickness), their orientation and their continuity, and the variation in permeability and porosity along strike and dip.

The specific properties of deformation bands compared with ordinary fractures (joints and slip surfaces) makes it important to investigate the controls on when deformation bands will form. Porosity seems to be an important factor controlling whether deformation bands or slip surfaces form as the first mesoscopic structures during rock failure. Few experimental or field data are available to pinpoint such a critical porosity limit, but it may be of order 10–15% for many sandstones (Wong *et al*. 1997). Other rock properties (cementation, clay content, grain size, grain shape), burial depth and fluid pressure will contribute to this limit, as may the local and remote state of stress. Field studies indicate that unusually thin deformation bands (a half to a quarter the size of normal bands) may form in the borderland between the two regimes of classical fracturing and deformation banding. Johansen *et al*. (2005) related a sequence of deformation band (first), thin deformation bands and jointing–shear fractures (last) to progressive quartz cementation and the corresponding decrease in porosity. Clearly, detailed field observations coupled with laboratory tests and physical and numerical modelling will yield a better understanding of these structures and a better prediction of their effect on fluid flow in hydrocarbon reservoirs and groundwater aquifers.

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