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A geometric model for the formation of deformation band clusters

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ABSTRACT

Arrays of closely-spaced (approximately <70 mm) sub-parallel cataclastic deformation bands are common structures in deformed, high-porosity (~10–35%) sandstones. The distribution of strain onto many small-displacement deformation bands is thought by some to result from strain-hardening of the cataclasite within individual bands. Examination of both normal and strike-slip faults with displacements ≤7 m from southeastern Utah, USA, and the North Island of New Zealand suggests, however, that clusters of deformation bands systematically develop at fault geometric irregularities (e.g., fault bends, steps, relays, intersections and zones of normal drag). The strain-hardening model does not account for clustering of deformation bands at fault geometric irregularities or the associated widespread coalescence of bands, and is not unequivocally demonstrated by post-peak macroscopic mechanical responses in laboratory rock deformation experiments. A geometric model is proposed in which individual bands within clusters develop sequentially due to migration of incremental shear strains at fault geometric irregularities as part of a slip localisation, asperity removal and strain weakening process. The geometric model, which does not require strain hardening of the fault rock, applies for the duration of faulting and a range of rock types in the brittle upper crust.

1. Introduction

Cataclastic deformation bands are low-strain structures widely observed in high porosity (~10–35%) sands and sandstones deformed at shallow depths of <3 km and low confining pressures of <40 MPa (e.g., Aydin, 1978; Aydin and Johnson, 1978, 1983; Underhill and Woodcock, 1987; Fowles and Burley, 1994; Antonellini and Aydin, 1994, 1995; Antonellini et al., 1994; Fossen and Hesthammer, 1997; Davis, 1999; Cashman and Cashman, 2000; Wibberley et al., 2000; Shipton and Cowie, 2001; Davatzes and Aydin, 2003; Okubo and Schulz, 2005, 2006; Schultz and Balasko, 2003; Schultz and Siddharthan, 2005; Aydin et al., 2006; Fossen et al., 2007; Fossen and Bale, 2007; Eichhubl et al., 2009; Rotevatn and Fossen, 2011). They are a type of shear band generally less than ~2 mm wide in which mechanical fracturing of grains is induced by shear displacements of up to several centimetres (e.g., Aydin and Johnson, 1978; Davatzes and Aydin, 2003; Rawling and Goodwin, 2003; Fossen et al., 2007 and references there in). Because these bands are narrow discontinuities comprising fault rock across which shear displacement has accrued, here they are considered to be a type of fault, consistent with fault definitions (e.g., Hobbs et al., 1976; Sibson, 1977; Davis, 1984; Kearey, 1996; Peacock et al., 2000).

Deformation bands commonly occur in arrays comprising many (10–100 s) sub-parallel and closely-spaced (<70 mm) bands (Fig. 1) referred to here as clusters. Deformation band clusters may be the sole component of a fault zone (Figs. 1d, 2a,b and 3a,b) or occur in association with a slip surface(s) where they are often referred to as defining a damage zone (Fig. 2c and e). Clustered bands bifurcate and anastomose so that the number of individual bands within a cluster can vary significantly in both dip and strike directions (Figs. 2a,b, 3 and 4). The systematics of these variations in the locations and numbers of bands constrains the origin of clusters of deformation bands.

Deformation band clustering, the focus of this paper, appears to record millimetre-scale migration of incremental shear strain which is generally considered to be evidence for strain hardening within each band (Aydin, 1978; Aydin and Johnson, 1983; Underhill and Woodcock, 1987; Fowles and Burley, 1994; Antonellini and Aydin, 1995; Mair et al., 2000; Schultz and Balasko, 2003; Fossen et al., 2007). The increased shear-resistance (strain-hardening) within deformation bands is attributed to elevated grain-contact friction, arising from grain fracturing and comminution, which, in turn, is inferred to cause a new band to form in the adjacent

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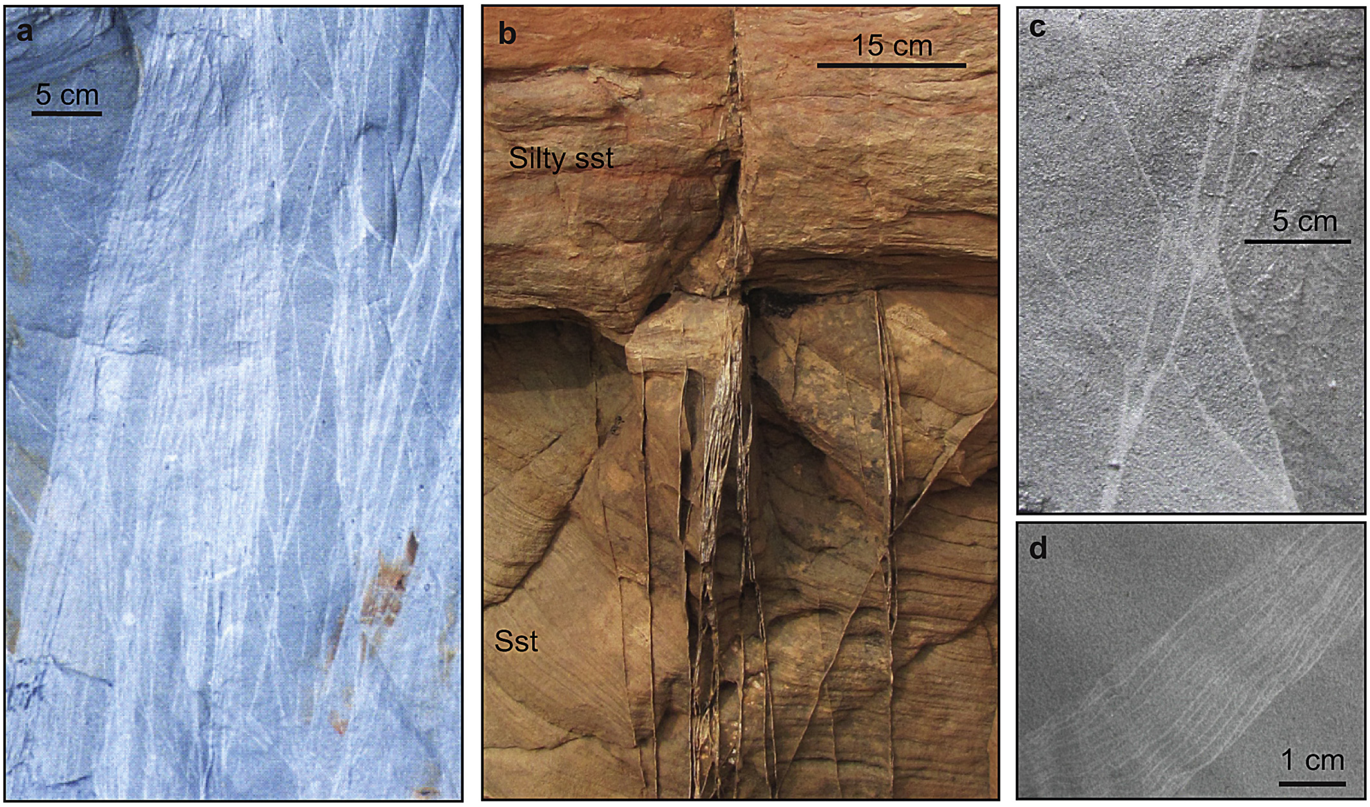

Fig. 1. Photographs of deformation bands in outcrop cross sections. (a) Sub-parallel deformation bands within a fault zone with a total throw of ∼5.5 m from coastal cliffs of Mount Messenger Formation, Tongaporutu, Taranaki Basin, New Zealand. (b) Deformation bands within a high-porosity sandstone which decrease upwards in number as the fault passes into a silty sandstone bed. Cross-sectional view from Navajo sandstone at Bridger Jack Mesa, southeast Utah, USA. (c) Cross sectional view of cross-cutting normal faults in sandstone of the Mount Messenger Formation at Tongaporutu, Taranaki, New Zealand. (d) Sub-parallel deformation bands from a small fault zone (throw 3 cm) within sandstone of the Mount Messenger Formation from Tongaporutu, Taranaki, New Zealand.

unfaulted wallrock rather than re-shearing the earlier-formed relatively stronger band (Aydin, 1978; Aydin and Johnson, 1978, 1983). Field observations and laboratory tests have suggested that many cataclastic deformation bands are presently stronger (and more resistant to erosion) than the adjacent host sandstone (e.g., Aydin and Johnson, 1983; Kaproth et al., 2010), however, evidence for strain hardening during band formation is equivocal and in some cases its applicability has been questioned (e.g., Herrin, 2001; Schafer, 2002). Therefore, the role of strain hardening in the formation of clusters of deformation bands in high-porosity sandstones remains a topic worthy of further consideration.

In this paper, we re-examine the outcrop and experimental data for cataclastic deformation bands and consider whether clusters of these bands could form without strain hardening of the fault rock. Using small normal and strike-slip faults (≤7 m displacement) from outcrops of high-porosity sandstone (∼10–35%) in southeastern Utah (Moab area), USA, and the North Island of New Zealand (Table 1) we conclude that clusters of deformation bands preferentially form at geometrical irregularities in fault zones (e.g., bends, steps, relays and zones of normal drag). A geometric model is proposed where the addition of new bands reflects progressive fault localisation during the strain weakening process generally thought to accompany faulting in the brittle upper crust (Sibson, 1977; Power et al., 1988; Sagy et al., 2007; Childs et al., 2009). The geometric model accounts for the formation of clusters of deformation bands without strain hardening and is applicable to models for the development of brittle fault zones in different rock types at shallow burial depths (<3 km depth), including high-porosity sandstones.

2. Deformation band data and observations

To address the question of whether the strain hardening or geometric model best accounts for the formation of cataclastic deformation-band outcrop patterns, small-displacement faults were examined from Tongaporutu and Whakataki in the North Island of New Zealand and near Moab in southeastern Utah, USA (Table 1). New Zealand normal (Tongaporutu) and strike-slip (Whakataki) faults displace sandstones in turbidites (beds ∼0.5–>10 m thick) buried to depths of 1.2–1.5 km (Tongaporutu) and 2–3 km (Whakataki), with average porosities of 30–35% (Tongaporutu, Browne et al., 2005) and 12–16% (Whakataki, Pollock et al., 2005). Normal and strike-slip faults from southeastern Utah, which have been widely studied (e.g., Aydin, 1978; Aydin and Johnson, 1983; Antonellini and Aydin, 1995; Foxford et al., 1998; Shipton and Cowie, 2001; Schafer, 2002; Davatzes and Aydin, 2003; Fossen et al., 2007; Eichhubl et al., 2009), displace massive to finely bedded sandstones of the Entrada Sandstone Formation, Navajo Sandstone Formation and Moab Tongue Member at burial depths of 1.5–3 km (Foxford et al., 1998; Shipton and Cowie, 2001) with typical porosities of ∼10–25% (e.g., Aydin, 1978; Antonellini et al., 1994; Antonellini and Aydin, 1995; Shipton and Cowie, 2001; Davatzes and Aydin, 2003). These Utah sandstones commonly contain >90% quartz (Foxford et al., 1996; Shipton and Cowie, 2001), which is in contrast to the significantly lesser proportions of quartz grains in the Mount Messenger Formation at Tongaporutu (quartz ∼30%, lithic ∼50% and feldspar ∼20%; Browne et al., 2005) and the Whakataki Formation at Whakataki (quartz 30–48%, lithics 13–23%, feldspar 3–18%; Pollock et al., 2005). Despite differences in sandstone