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Polygonal deformation bands

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ABSTRACT

We report for the first time the occurrence of polygonal faults in sandstone, which is compelling given that layer-bound polygonal fault systems have been observed so far only in fine-grained sediments such as clay and chalk. The polygonal faults are shear deformation bands that developed under shallow burial conditions via strain hardening in dm-wide zones. The edges of the polygons are 1–5 m long. The shear deformation bands are organized as conjugate faults along each edge of the polygon and form characteristic horst-like structures. The individual deformation bands have slip magnitudes ranging from a few mm to 1.5 cm; the cumulative average slip magnitude in a zone is up to 10 cm. The deformation bands heaves, in aggregate form, accommodate a small isotropic horizontal extension (strain <0.005). The individual shear deformation bands show abutting T-junctions, veering, curving, and merging where they mechanically interact. Crosscutting relationships are rare. The interactions of the deformation bands are similar to those of mode I opening fractures. The documented fault networks have important implications for evaluating the geometry of km-scale polygonal fault systems in the subsurface, top seal integrity, as well as constraining paleo-tectonic stress regimes.

1. Introduction

Polygonal fault systems are intriguing features that have important implications in structural and petroleum geology. First, because of their implications for compaction and fluid expulsion in sedimentary basins (Tewksbury et al., 2014) and second, because they may influence overpressures, top seal integrity of hydrocarbon reservoirs, and secondary migration along faults and fractures (Cartwright et al., 2007; Goulty, 2008). Recently, they have been used as analogs in planetary geology to interpret the polygonal terrains of Mars (Cooke et al., 2011; Moscardelli et al., 2012).

Polygonal faults were recognized for the first time in high-resolution reflection seismic lines and horizontal time slices datasets in Neogene (Paleogene) sediments of the North Sea slope and the North Atlantic Margin (Cartwright, 1996; Cartwright and Lonergan, 1997; Lonergan and Cartwright, 1999) and subsequently in many other basins throughout the world (Cartwright and Dewhurst, 1998; Watterson et al., 2000; Gay et al., 2004).

The major characteristics of geophysically-detected polygonal fault systems are the following: (1) The faults are arranged in a polygonal pattern without any preferential directional trend; (2) The faults are all normal in type and their throw varies from 10 to 100 m (10 m is the typical lower resolution limit for seismic surveys); (3) They have been observed only in deformed marine depositional sediments within passive tectonic settings showing negligible tectonic extension or compression (Cartwright and Dewhurst, 1998; Goulty, 2008); (4) They are layer-bound with a 100–1000 m horizontal spacing; (5) Their polarity may switch from synthetic to antithetic; (6) The sediments in which they develop are predominantly ultrafine-grained smectitic claystones or carbonate chalks (Hibsch et al., 2003; Hansen et al., 2004; Tewksbury et al., 2014) with high porosity (>0.5) and low permeability (<10^{-17} m^2).

A common trait of the sedimentary units containing polygonal fault networks is that they are deformed by uniform extensional strain (Cartwright and Dewhurst, 1998). In particular, Lonergan et al. (1998) and Goulty et al. (2002) suggest that the polygonal geometries result from local variations in horizontal stress around the polygonal fault segments as they develop.

Different explanations have been proposed for the genesis of these structures. One of the most accepted involves a volumetric contraction (including horizontal shortening of the bed layers) during the early stages of compaction and dewatering where syneresis, a colloidal mechanism common in smectite clays, allows for

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pore fluid expulsion from sedimentary gels under osmotic or electrochemical forces (Cartwright and Lonergan, 1996; Cartwright and Dewhurst, 1998). Other mechanisms that have been proposed include gravitational loading, density inversion (Cartwright et al., 2003), overpressure development (Cartwright and Lonergan, 1996; Tewksbury et al., 2014), and low friction coefficients on the faults (Goulty, 2001, 2008).

For the first time, we describe polygonal fault systems that develop in porous sandstone at the outcrop scale (tens to hundreds of m) and by means of strain localization and development of deformation band faults. Strain localization due to failure in porous granular rocks is a well-documented process in siliciclastic sedimentary rocks (Aydin et al., 2006; Fossen et al., 2007 and references therein), in porous carbonate rocks (Micarelli et al., 2006; Tondi et al., 2006), as well as in hyaloclastites and hyalotuffs (Barnes and Kattenhorn, 2010; Tewksbury, 2010). The process of strain localization results in tabular structures called deformation bands (*sensu* Aydin, 1978; Aydin and Johnson, 1978), which can accommodate shear and compaction or even dilatancy (Antonellini et al., 1994; Mollema and Antonellini, 1996; Aydin et al., 2006; Fossen et al., 2007). Depending on rock properties (i.e., grain size, grain sorting, porosity, and clay content) and on the stress state at the time of their formation, deformation bands may be associated with porosity reduction accomplished by grain sliding and rotation, grain crushing and pore collapse, or by dilatancy and granular flow (Antonellini et al., 1994; Antonellini and Mollema, 2002; Fossen et al., 2007).

The development of deformation bands in porous rocks has been a matter of interest in the past two decades for the effects that they may have on fluid flow (Nelson, 2001; Faulkner et al., 2010). Deformation bands influence sealing, flow buffering, and reservoir compartmentalization during production (Antonellini and Aydin, 1994, 1995; Matthäi et al., 1998; Manzocchi et al., 1999; Jourde et al., 2002; Sternlof et al., 2006). The literature characterizing the petrophysical properties of deformation bands in siliciclastic rocks is extensive and includes the early work of Antonellini and Aydin (1994) to the more recent works of Ballas et al. (2012, 2013), Braathen et al. (2013), and Deng et al. (2015).

The normal faults belonging to the polygonal systems described in this paper are compactive shear bands (*sensu* Aydin et al., 2006) or disaggregation bands (*sensu* Fossen et al., 2007) with no or poorly developed cataclasis. Single compactive shear bands are small shear zones (slip magnitudes 1–20 mm) and represent the incipient stage of faulting in porous rock (Aydin, 1978; Aydin and Johnson, 1978; Aydin et al., 2006). With continuing deformation, these structures evolve into zones of compactive shear bands (10–50 bands in close proximity), which have cumulative slip magnitudes of <0.5 m. As deformation progresses, at a slip magnitude larger than 0.5 m, depending on fault type, lithologic characteristics, and stress state (Aydin et al., 2006), zones of compactive shear bands become well-developed discontinuities – faults (*sensu strictu*) – containing discrete slip surfaces.

The importance of our work is the characterization of polygonal faults at the outcrop scale and in a different rock type and lithology, namely sandstone, than so far reported. This opportunity allows us to extend the hierarchy of polygonal structures below their observed seismic limit and in coarse-grained rock types. The observations of polygonal fault systems in sandstone imply that other genetic mechanisms and fault localization processes may be at work with respect to those so far described.

The polygonal fault networks described here also have implications for the interpretation of the regional geology of SE Utah in relation to the paleo-stress, the fluid flow, and the deformation characteristics within the upper portion of the Jurassic Carmel Fm and lower portion of the Jurassic Entrada Sandstone.

2. Study area

The polygonal deformation band systems described in this paper are observed at Arches National Park in southeastern Utah (USA) about 350 m northeast of Ring Arch (Fig. 1). The structures are contained in an aeolian cross-bedded sandstone layer (with an estimated thickness of 1–2 m) within the lower part of the Jurassic Entrada Sandstone (Doelling, 2002) (Fig. 2). This section of the Entrada Sandstone was previously named the Slickrock Member of the Entrada Sandstone (Doelling, 1985). The outcrops are exposed on a sandstone pavement 5 m above layered siltstones and mudstones belonging to the Middle Jurassic Carmel Fm. The pavement is 150 m by 170 m across. The Ring Arch area is located on the southwest limb of the NW-plunging Courthouse Wash syncline; no large slip (>1 m) faults are present and the strata gently dip < 5° towards the syncline axis (Fig. 3).

The post-Triassic tectonic history of Arches National Park includes the growth and subsequent collapse of a NW–SE-trending salt anticline (Salt Valley anticline). The deformation related to these events is recorded by faulting and folding in the exposed Mesozoic formations (Doelling, 1985, 2001). Units younger than the Entrada Sandstone have a uniform thickness over the area and, therefore, their deposition does not seem to have been affected by salt diapirism that occurred from Permian to Jurassic time. Salt dissolution and collapse processes leading to normal faulting along the edges of Salt Valley and Cache Valley started later and are probably still ongoing (Doelling, 1985, 1988; 2001).

The Entrada Sandstone and Carmel Fm are the two lithologic units of interest. The Entrada Sandstone is made up of fine-grained cross-bedded aeolian sandstone with a reddish color due to the presence of hematite. The layers in the upper part of the formation are massive beds more than 5 m thick, while the beds in the lower part are thinner (1–2 m). The thickness of the Entrada Sandstone is around 80 m (Doelling, 2002). The bedding of the Carmel Fm is thin (0.1–0.3 m), convolute, and contorted with syn-depositional folds that sometimes affect the bottom section of the overlying Entrada Sandstone. The Carmel Fm has a thickness ranging from 27 to 34 m (Doelling, 2002) and was deposited in a sabkha environment. It is composed of dark, reddish, fine-grained, silty sandstone (Blakey et al., 1988; Doelling and Morgan, 2000; Doelling, 2001).

3. Methods

The polygonal fault systems have been characterized by mapping at scales from 1:1 to 1:500. The polygons were mapped from an aerial photograph to identify the distribution of the structures on the pavement exposure at a scale of 1:500 (Fig. 3). The zones of deformation bands making up individual polygons were mapped in the field by means of tape rulers and photo-mosaics (at a scale of 1:1 to 1:50). Details of the intersections among zones of deformation bands at the vertices of polygons were mapped directly onto acetates laid on the structures (scale of 1:1 then reduced to 1:50). The traces on the acetates were reduced and digitized in Photoshop™.

The slip magnitudes across the deformation bands were measured on the outcrops using the top of the sandstone layer as an offset marker where the deformation bands were freshly exposed and fine lineations were present on the planes of the band surfaces. Crosscutting relationships between faults and cross-beds, or between multiple faults, were also used to determine slip solutions.

The cumulative average slip magnitude distributions along zones of deformation bands that form the edges of the polygons were also determined. This was done by considering each polygon edge as a single structure (see Fig. 6b for reference frame). The number of single deformation bands dipping towards the inner