Fault zone architecture and its scaling laws: where does the damage zone start and stop?

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Abstract: Damage zones of different fault types are investigated in siliciclastics (Utah, USA), carbonates (Majella Mountain, Italy) and metamorphic rocks (western Norway). The study was conducted taking measurements of deformation features such as fractures and deformation bands on multiple 1D scanlines along fault walls. The resulting datasets are used to plot the frequency distribution of deformation features and to constrain the geometrical width of the damage zone for the studied faults. The damage-zone width of a single fault is constrained by identifying the changes in the slope of cumulative plots made on the frequency data. The cumulative plot further shows high deformation frequency by a steep slope (inner damage zone) and less deformation as a gentle slope (outer damage zone). Statistical distributions of displacement and damage-zone width and their relationship are improved, and show two-slope power-law distributions with a break point at c. 100 m displacement. Bleached sandstones in the studied siliciclastic rocks of Utah are associated with a higher frequency of deformation bands and a wider damage zone compared to the unbleached zone of similar lithology. Fault damage zones in the carbonate rocks of Majella are often host to open fractures (karst), demonstrating that they can also be conductive to fluid flow.

Fault architecture, geometry and properties have formed key topics of interest across a range of disciplines through several decades (e.g. Caine et al. 1996; Wibberley et al. 2008; Bastesen & Rotevatn 2012; Torabi et al. 2013; Gabrielsien et al. 2016; Choi et al. 2016). Faults are of fundamental and societal importance, since they may trigger earthquakes (Martel & Pollard 1989), influence fluid flow in water aquifers (Bense et al. 2013) and in hydrocarbon and geothermal reservoirs (Gabrielsen et al. 1990; Loveless et al. 2014), and affect reservoir capacity and seal integrity for CO₂ and waste storage (Cappa & Rutqvist 2011; Rohmer et al. 2015). Faults can be described as comprising a core and an enveloping damage zone (e.g. Caine et al. 1996), with the fault core accommodating the majority of displacement and deformation. The fault core (Fig. 1) may comprise one or several slip surfaces, assemblages of fault rock (fault gouge, cataclasites, breccia, clay smear), fractures (veins, extension and shear fractures), diagenetic structures (cement, mineralization), and lenses of deformed and undeformed rocks trapped between slip surfaces (Wibberley et al. 2008; Bastesen et al. 2009; Braathen et al. 2009). The deformation intensity decreases away from the fault core into the damage zone and host rock (Fig. 1). A fault damage zone may include smaller faults, fault-related folds, deformation bands and/or fractures, depending on the initial porosity of the rock under deformation (Billi et al. 2003; Berg & Skar 2005; Faulkner et al. 2010; Torabi & Berg 2011; Rotevatn et al. 2016). A deformation band is a tabular strain-localization structure that forms in highly porous rocks and sediments.
Fig. 1. (a) Schematic illustration of a normal fault. Note the highly deformed fault core (including fault lenses) and the surrounding wall damage zones; black double-arrows indicate the location of scanlines along wall damage zones in the hanging wall and footwall. The scanlines were made 2 m apart along the fault height, defined as baseline (first scanline) and second scanline. (b) The damage zone of a single fault can be divided into the wall damage zone (Peacock et al. 2017) (the cross-fault and along-fault damage zones) and the around-tip damage zone (Choi et al. 2016). (b) is modified after Shipton & Cowie (2003). (c) A frequency plot of fractures/deformation bands within the damage zone of a fault combined with cumulative distribution, which has been used to constrain the damage-zone (DZ) width. (d) A photograph showing a mapped fault core in siliciclastic rocks in Utah. The photograph shows a cliff face of the outcrop. The white lines show the boundaries of fault core, while the red lines indicate a shale layer introduced into the fault core as a shale smear.
as a result of grain reorganization or breakage (e.g. Aydin 1978; Torabi 2007). The damage zone of a single fault is spatially classified into wall damage zone (cross-fault and along-fault damage zones) and tip damage zone, depending on its location around the fault (Fig. 1a, b) (Choi et al. 2016; Peacock et al. 2017) for a 3D fault exposure. Inner (close to the fault core) and outer (close to host rock) damage zones have also been identified in some cases, referring to the zones with higher and lower deformation frequency, respectively (Berg & Skar 2005).

Faults can be characterized by their geometrical attributes, such as fault length (fault dimension parallel to fault strike), fault height (fault dimension parallel to fault dip), displacement, damage-zone width and fault-core thickness (e.g. Barnett et al. 1987; Walsh & Watterson 1988; Nicol et al. 1996; Torabi & Berg 2011 and the references therein) (Fig. 1). The statistical distribution of these fault attributes and the relationship between them are used to predict and constrain the fault dimensions in the subsurface. Researchers have worked on the spacing and distribution of deformation bands and fractures around faults to constrain damage zone boundaries (width) using different criteria (e.g. Berg & Skar 2005; Schueller et al. 2013; Choi et al. 2016). Furthermore, the scaling relationship (mainly power law) between fault displacement and damage-zone width (Scholz et al. 1993; Knott et al. 1996; Beach et al. 1999; Fossen & Hesthammer 2000; Shipton et al. 2006; Mitchell & Faulkner 2009; Savage & Brodsky 2011; Torabi & Berg 2011), in addition to other relationships such as displacement v. fault length (e.g. Cowie & Scholz 1992; Dawers et al. 1993; Kim & Sanderson 2005; Kolyukhin & Torabi 2012), have been used to predict fault growth and evolution. However, these statistical relationships are based on limited data that do not cover all ranges of fault sizes. In addition, the data are scattered with a few points for some of the fault sizes. Furthermore, each static data point represents the end state for one fault and contains no dynamic information on how the fault evolved to get to that end point. Even though fault attributes and their scaling laws have been widely studied, there are still several issues that need to be addressed (e.g. Choi et al. 2016). These include: (i) the inconsistent and non-uniqueness of definitions of fault geometrical attributes such as damage zone and fault core (Fig. 1); (ii) a lack of a proper constraint on fault dimensions (e.g. damage-zone width) that can be globally applied; (iii) inaccessibility of the fault structure including its damage zone in 3D; and (iv) gaps in the data and the scaling relationships between fault displacement and damage-zone width. The main aim of this study is to focus on two of the above presented issues (i.e. issues ii and iv) by:

- improving the geometrical constraint on the fault damage-zone width;
- improving the scaling relationships between the fault damage-zone width and displacement.

This is fulfilled using a larger dataset that covers different fault types and lithologies. To address the above stated study aims, we investigate and compare faulted rocks of three different lithologies in different tectonic settings:

- Faulted siliciclastic rocks acquired from faults in Moab and the San Rafael Swell in Utah (USA), where Carboniferous–Cretaceous rocks were deformed in an extensional regime.
- Faulted carbonate deposits of the Majella Mountain (central Italy), where Cretaceous rocks are displaced by normal, reverse and strike-slip faults.
- Faulted basement rocks in the Øygarden Gneiss Complex on Sotra Island on the west coast of Norway that are displaced by normal faults with a strike-slip component.

The studied faults vary in type and size, yielding a good representation of possible variations in damage-zone width and internal structures for different fault sizes and types. Our work is performed by studying damage zones through multiple 1D scanlines covering the outcrops along fault walls where all fault-related structures are recorded (Fig. 1a). In order to constrain the damage-zone width, we define the statistical damage zone boundaries using changes in the slope of cumulative frequency of fracture/deformation bands as a function of the distance from the fault core in both the hanging wall and footwall (Fig. 1c) (e.g. Berg & Skar 2005; Choi et al. 2016).

Methodology

Fieldwork

In the present study, only measurements of fault-wall damage zones (or cross-fault damage zone in Fig. 1a, b) are presented. We differentiate between fault core and damage zone following the definitions provided in the introduction of this chapter (Fig. 1a, b, d). This means, by first identifying the fault core and its boundaries with the damage zone (Fig. 1d). The fault-core thickness is defined as the total thickness of fault rocks or fault breccias, crushed material and lenses incorporated between slip surfaces in the fault core (Fig. 1d). For most of the studied outcrops, 1D scanlines were acquired in the wall damage zone (cross-fault damage zone) at least at two levels (2 m apart) along the fault height. These are called baseline (at the ground level) and second scanline (at 2 m elevation along the fault height) (Fig. 1). Whenever the outcrop was not accessible, only a baseline was acquired. This method was chosen in
order to study the changes in damage zone structure and width along a fault in the accessible parts of the damage zones. Along each scanline, the orientation and distance from the master fault for deformation features, such as small faults, deformation bands, stylolites and fractures, were registered using a compass clinometer and measuring tape. The resulting dataset is then used to plot the frequency distribution of deformation features around the faults and to constrain the geometrical width of damage zone for the studied faults (Fig. 1c). It is worth mentioning that we do not differentiate between deformation bands and fractures or between different types of fractures (such as veins, tension and shear fractures) in the plots, although we have identified them in the field and discuss their effects on the damage-zone width. Displacement was measured both in the field and on the scale-calibrated photographs of the outcrops wherever it was not accessible in the field. The field and photograph displacement measurements show a linear correlation with a very high coefficient of determination ($R^2$ of 0.98).

**Statistical analysis**

In this study, frequency plots are used to document deformation frequency along the scanlines, which are separated into 1 m bins. The bin value is the number of deformation features located within that specific metre (Fig. 1c). For single faults (non-overlapping damage zones), the scanline data are further studied using cumulative frequency where the bin values across the frequency plot are summed (Fig. 1c). The change in the slope of the cumulative plot is used to constrain the width of both inner and outer damage zones wherever it is possible (Berg & Skar 2005; Choi et al. 2016; Ellingsen 2017) (see Fig. 1c). This plot further shows a high frequency of deformation by a steep slope and less intense deformation as a gentle slope, which could be considered as the inner and outer damage zones, respectively. The inner and outer damage zones represent different degrees of strain distribution around the master faults. The inner damage zone is closer to the fault core, and includes higher strain and deformation intensity in comparison to the outer damage zone. Local variations are also observed in the frequency plots that could be related to the presence of minor faults or changes in the mechanical properties of layered rocks, as well as to the interaction of damage zones from separate faults.

Sometimes, damage zones are interacting, which could result in either overlapping or linking faults. Where damage zones are closely spaced and interacting, the resulting cumulative plot yields a fluctuating pattern, not optimal for constraining the damage-zone width. In this case, we use a criterion to constrain the host rock or background fracture frequency first and, hence, indirectly find the width of the damage zone. The background fracture frequency represents the frequency of pre-existing deformation features that are not assumed to be directly linked to the faulting itself. Background fracturing is recognized by a low and stable fracture frequency without any severe fluctuation in frequency, such as a rapid rise in deformation intensity attributed to faulting. Using background fracture frequency of the outcrop, we are thereby able to constrain the width of the damage zone.

In order to investigate the distribution of univariate parameters such as damage-zone width and displacement, exceedance frequency plots are applied:

$$EF_{x_i} = \frac{n - n_i}{n}. \quad (1)$$

Here, EF is the exceedance frequency for the value $x$ of a measured variable, defined as the number of data with values greater than that $x$ (numerator) divided by the total number of the data ($n$, or the denominator), see equation (1). $i$ is the data sample number. The $x$ values in this study are either damage-zone width or displacement measurements, plotted on the $x$-axis. These plots (see later in the discussion) further illustrate how the gathered data points are distributed. The distribution of the points could be normal (Gaussian distribution), power law (hyperbolic), log-normal or exponential (Poissonian), in which they are distinguished from each other by their characteristic patterns (Nemec 2011; Seifried 2012). A power-law distribution is often associated with these fault geometrical attributes, displayed as a linear fit to the data on logarithmic axes, indicating a scale independency (Torabi & Berg 2011) as opposed to exponential and log-normal distributions (Bonnet et al. 2001). In order to exclude the resolution effect, the highest and lowest values referred to as the higher and lower bounds of the dataset, respectively, are removed when using power-law distribution, as these values are at risk of being outliers.

Furthermore, the relationship between damage-zone width and displacement is investigated for a wide range of fault sizes and compared with the existing data from the literature. This means studying the statistical dependence of damage-zone width and displacement, enabling the prediction of the former based on the knowledge of the latter.

**Geological framework**

Data for this work were acquired from several fieldwork seasons conducted in three key study areas (Figs 2, 3 & 4): (i) the Colorado Plateau (Paradox Basin and San Rafael Swell, Utah, USA: Fig. 2); (ii) the Majella Mountain (Central Apennines,
Fig. 2. A map showing the studied localities in Utah. The following abbreviations have been used: Loc. 1, Hidden Canyon; Loc. 2, ANP (R-191); Loc. 3 for Cache Valley; Loc. 4, Humbug Flats locality. Modified from Doelling (2001).

Fig. 3. A geological map over the Majella Mountain. The geological evolution is illustrated through the stratigraphic column. Modified from Di Cuia et al. (2009), Aydin et al. (2010) and Masini et al. (2011).
Italy: Fig. 3); and (iii) on Sotra Island (western Norway: Fig. 4).

Utah Field localities, USA – Moab Fault, Cache Valley and Humbug Flats

Four localities were studied in Utah; three of these are in the Paradox Basin. Two of the Paradox Basin localities (along the Hidden Canyon and Highway R-191, respectively) are segments of the Moab Fault (Foxford et al. 1996). The third Paradox Basin locality is in Cache Valley outside Arches National Park (Fig. 2). The fourth outcrop studied in Utah is in the Humbug Flats at the northern flank of the San Rafael Swell (Fig. 2).

The Paradox Basin developed in middle Pennsylvanian times as one of a series of intra-cratonic basins across what now comprises the western USA (e.g. Trudgill 2011). It developed as a foreland basin, the result of flexural subsidence in the footwall of the growing Uncompahgre Ancestral Rocky Mountain thick-skinned uplift (Baars & Doelling 1987; Doelling 1988; Draut 2005; Trudgill 2011). The Moab Fault follows the Moab Valley, which in turn defines the collapsed crest of the Moab Valley salt wall (Fig. 2). The Moab Valley salt wall formed as a result of salt mobilization during Permian–Triassic times, whereas the Moab Fault itself formed later (c. 60–43 Ma: Pevear et al. 1997), clearly unrelated to the initiation and growth of the salt wall. The distinctive topography of the area is likely to have developed as the fault accumulated further slip in recent (<5 Ma) times, driven by dissolution and collapse of the salt wall (Trudgill 2011).

The Moab Fault is a 45 km-long NW-striking normal fault system, with a maximum surface throw of c. 950 m (Foxford et al. 1996), which offsets rocks of Carboniferous–Cretaceous age (e.g. Berg & Skar 2005) and detaches within the evaporitic Pennsylvanian (Carboniferous) Paradox Formation (Fig. 2). To the north, the Bartlett, Courthouse and Thusher faults (Johansen & Fossen 2008) are...
splay faults of the Moab Fault, which branch westwards to northwesterns of the Moab Fault at the Courthouse branch point (Fig. 2). The Bartlett Fault exhibits an offset in the range of 170–300 m, and is studied at the Hidden Canyon locality (Fig. 2).

A salt-crestal fault system with a similar history is found in the Cache Valley outside Arches National Park (ANP), where normal faulting was initiated by the collapse and dissolution of salt walls (Trudgill 2011; Alikarami et al. 2013). The studied normal fault in Cache Valley forms part of this salt-crestal fault system, and exhibits a displacement of c. 30 m (Alikarami et al. 2013), juxtaposing the Dewey Bridge and Slick Rock members (Entrada Formation) in the hanging wall against the Navajo Formation in the footwall (Fig. 5).

The San Rafael Swell (SRS) is an asymmetrical, doubly NNE–SSW-plunging domal uplift structure, interpreted to have formed during Late Cretaceous Eocene Laramide contraction (Bump & Davis 2003; Fischer & Christensen 2004; Cross 2009). A gently dipping western limb characterizes the SRS, whereas the eastern flank forms a monocline (the San Rafael Monocline (SRM)) with a steep forelimb. The SRS has been interpreted as a forced fold (Stearns & Jamison 1977) or a fault-propagation fold (Bump & Davis 2003; Davis & Bump 2009) controlled by a blind thrust or reverse fault. The studied localities in the SRS area are located in the NE part of the SRS (Loc. 4 in Fig. 2). Faulting in the Humbug Flats is considered to be related to the formation of the SRS; the uplift resulted in normal faulting of part of the Entrada Sandstone, which is equivalent to the Dewey Bridge Member in the Paradox Basin in the Moab area (e.g. Shipton & Cowie 2001). The normal faulting accommodates a displacement gradient along the plunging anticline.

### Majella Mountain, Italy

The Majella Mountain (Central Apennines, Italy) is a thrust-related anticline plunging both northwards and southwards (Patacca et al. 2008). It is composed of Upper Jurassic–Upper Miocene limestones and dolostones pertaining to the Apulia Carbonate Platform (Crescenti et al. 1969) developed in the southern margin of the Tethys Ocean (Fig. 3) (Eberli et al. 1993, 2004). The deformation history of the carbonate successions exposed in the Majella area can be classified into four main episodes (Di Cuia et al. 2009):

1. **Middle Cretaceous ENE–WSW extension due to hinterland subsidence and volcanic activity** has led to NNW–SSE-striking normal faults in advance of the tilting of the carbonates (Aydin et al. 2010). Late Cretaceous extension related to the closure of the Ligurian–Piemontese Basin resulted in the tilting of the first set of normal faults.

2. A second episode of extension during the late Miocene is responsible for the post-tilting development of a new set of normal faults. NE–SW extension and shearing of pre-existing pressure-solution seams parallel to bedding, and further reactivation and linking of older normal faults led to additional faulting in the area (Graham et al. 2003; Aydin et al. 2010).

3. Pliocene folding and thrusting is related to the convergence between the European and the African plates (Cosentino et al. 2010). In particular, the Middle–Late Pliocene is characterized by the formation of the Majella anticline and reverse faults (Di Lazio et al. 2004).

4. Strike-slip tectonics were caused by regional uplift and thrusting from the Pliocene to the Pleistocene. Left- and right-lateral strike-slip faults were initiated by the shearing of bed-parallel pressure-solution seams, where these two sets cross each other at a 60°–70° angle.

The studied area in the Vallone Santo Spirito (outside Fara San Martino: Fig. 3) consists of the Lower Cretaceous Morrone di Pacentro Formation (Fig. 3), whereas the outer forelimb of the Majella anticline is represented by the Upper Cretaceous Cima delle Murelle Formation (Agosta et al. 2010a; Festa et al. 2014). These formations are located adjacent to the central part of the Apennine fold-and-thrust belt (Agosta et al. 2010b) in the eastern forelimb (Aydin et al. 2010).

### Sotra Island, western Norway

The Øygarden Gneiss Complex exposed along Lislaska in Sotra Island is situated in the west of the Bergen Arc System, a group of arcuate tectonic units along the west coast of Norway (e.g. Fossen 1988; the green area in Fig. 4). This gneiss complex consists of felsic gneisses, a parautochthonous block forming part of the Baltic Shield during the Precambrian–Caledonian (Fossen & Dunlap 1998; Larsen et al. 2003). The Øygarden Gneiss is reported to be the basal unit of the allochthonous rocks in the Bergen Arc System, bounded by a mylonitic contact (Fossen & Rykkveld 1990). The rocks were exposed to extensive deformation of lower-amphibolite-facies metamorphism, both before and after the Scandian thrusting. Even though older deformation was overprinted by more recent tectonic events, some evidence of the pre-Caledonian deformations prevails in the rock record (Fossen & Rykkveld 1990).

From the Ordovician to the Early Devonian, Laurentia collided with Baltica and formed what is called the Caledonian orogeny. This episode of deformation subjected the Øygarden Gneiss Complex to
Fig. 5. (a) Outcrop photograph of the Hidden Canyon, displaying a segment of the Moab Fault. The red dashed lines show the approximate location of fault lenses within the fault core. The black lines show the approximate locations of scanlines in the footwall. Frequency plots (in columns) combined with cumulative distributions (in dots) from the Hidden Canyon scanlines data (b)–(e). The grey lines show the general slope of the cumulative distributions. (b), (c) & (d) are scanlines along the Slick Rock Member and (e) is the scanline taken along the Moab Tongue Member of the Entrada Sandstone. (f), (g) & (h) Stereoplots showing the orientations of faults, deformation bands and fractures at this locality. Different colours indicate orientation data from different scanlines: pink, scanline 1; blue, scanline 2; green, scanline 3.
amphibolite-facies metamorphism (Fossen & Rykkylid 1990). Mylonitic foliation and a large fold displaying parasitic buckle folds are evidence of a complex semi-ductile evolution of the area (Fossen & Rykkylid 1990).

Most of the brittle deformation observed in the Øygarden Gneiss Complex is related to the post-Caledonian deformation as brittle deformation started to dominate. The Devonian fracture and fault population can be divided into set I (NE–SW orientation) and set II (NW–SE orientation) (e.g. Larsen et al. 2003, Fig. 4). Set I is found in the form of fractures and faults, which are normal faults with strike-slip components; the faults formed during NW–SE extension and are filled with epidote. The set II fractures are filled with calcite and breccia formed during later east–west extension (Larsen et al. 2003).

**Results**

**Utah outcrop data**

**Hidden Canyon.** The Moab Fault segment in Hidden Canyon (Bartlett Fault, Loc. 1 in Fig. 2) has a throw of c. 200 m at the surface, juxtaposing the Jurassic Slick Rock Member and Moab Tongue Member of the Entrada Sandstone in the footwall against the Cretaceous Cedar Mountain Formation in the hanging wall (e.g. Foxford et al. 1996; Berg & Skar 2005; Johansen & Fossen 2008) (Fig. 5). The footwall damage zone was investigated in detail in four scanlines. The data were gathered on the Slick Rock Member on three scanlines spaced 2 m apart, orientated perpendicular to the fault (Fig. 5b, c, d). Each scanline covers a 50 m-long section in the footwall. In addition, a fourth footwall scanline (78 m in length) was recorded in the Moab Tongue Member (Fig. 5e). Orientation data gathered from fractures and deformation bands along the scanlines (Fig. 5) show a dominant NW–SE trend dipping towards either the NE or SW in synthetic or antithetic orientations relative to the fault (on average, N325, 75).

The frequency data (Fig. 5) display a rapid decrease in deformation intensity away from the fault characterizing the extent of the damage zone. In all of the three scanlines recorded in the Slick Rock Member, deformation bands are the dominant structures close to the fault, whereas fractures dominate the damage zone further from the fault. Scanlines 2 and 4 were performed in bleached layers, characterized by their brighter brownish colour compared to the red sandstone in scanlines 1 and 3. Both scanlines 2 and 4 show relatively higher frequency of deformation bands, and wider inner and outer damage zones in comparison to the other two scanlines (Fig. 5). Inner and outer damage zones are estimated to be approximately at 8.5 and 31.5 m for scanline 2, and 13 and 61.5 m for scanline 4 (Fig. 5c, e).

Whereas, inner and outer damage zones for scanlines 1 and 3 are 4 and 5.5 m, and 2.5 and 27.5 m wide, respectively (Fig. 5b, d). The damage zone in the hanging wall includes fractured rocks and a drag fold (syncline) in the Cedar Mountain Formation. Since most of the rocks in the hanging wall were crushed due to intense fracturing observed in the upper part of the outcrop, only a rough estimation of damage-zone width (c.169 m wide) was made along a scanline in Cedar Mountain Formation based on the background fracture frequency.

**ANP (Highway R-191).** The studied outcrop lies near the entrance to the Arches National Park (Loc. 2 in Fig. 2) and features a footwall damage zone at a segment of Moab Fault affecting the Carboniferous Honaker Trail Formation (Fig. 6). Therefore, the outcrop is included to sample smaller faults and their damage zones within the broader Moab Fault damage zone. The Honaker Trail Formation can be separated into four distinct stratigraphic layers (Nuccio & Condon 1996; Seifried 2012). The deformation in this damage zone is represented by small faults, deformation bands and fractures. The orientation data from deformation bands and fractures show a clear similarity to the fault orientation (NW–SE), with the majority showing a synthetic or antithetic orientation relative to the Moab Fault (Fig. 6).

Two 181 m-long scanlines spaced 2 m apart were recorded (Fig. 6b, c). In total, 26 normal faults were measured. Most faults show displacement of less than 1 m, but larger displacement up to 7.2 m occurs. Faults 1, 2 and 3 are not observed in the second scanline, while the rest of the faults are cut by this scanline (Fig. 6b, c). The frequency of fractures around the faults is slightly higher in the second scanline (Fig. 6b, c). The small spacing between the faults resulted in overlapping damage zones, which makes it hard to distinguish the individual damage-zone widths for each fault using cumulative plots. The frequency of fractures and deformation bands tends to increase close to faults; exceeding two or more deformation features per metre (Fig. 6). The background frequency is one deformation band or fracture per metre, which is used to constrain the damage-zone width of the small faults in this outcrop.

**Cache Valley.** The main fault in the studied locality is a normal fault, which has a displacement of around 30 m and is situated outside the Arches National Park (Loc. 3 in Fig. 2). In addition, we have measured five other faults (numbered in Fig. 7) with displacement ranging from a few centimetres to several metres in this outcrop. The deformation in the footwall and hanging wall were recorded using 1D scanlines. Since the outcrop was steep and not accessible at higher levels, only baselines were made on the ground level. The footwall is made of Navajo
Fig. 6. (a) Studied outcrop along ANP (R-191), representing a section of the footwall damage zone of a Moab Fault segment. The Moab Fault is exposed in the Arches National Park entrance. (b) & (c) Frequency plots (in columns) for the data gathered along scanlines. (d)–(f) Stereoplots showing the orientation of fractures, deformation bands and faults in this locality.
Fig. 7. (a) Outcrop photographs showing the studied faults and the stratigraphy in Cache Valley. Note the bleached zones in Navajo Sandstone. (b) A close-up of the hanging wall showing the bleaching and small faults. (c) & (d) Frequency plots (in columns) with cumulative distributions (in dots) from the Cache Valley scanlines data for the hanging wall and footwall, respectively. The grey lines show the general slope of the cumulative distributions. (e) Orientations of faults. (f) & (g) Orientation of deformation bands and fractures in the hanging wall (HW) of the master fault (fault 1). (h) & (i) Orientations of deformation bands and fractures in the footwall (FW) of fault 1.
Sandstone, whereas the hanging wall includes different rock units (i.e. Navajo Sandstone at its base overlain by the Dewey Bridge Member and Slick Rock Member of the Entrada Sandstone) (Fig. 7). Like the Hidden Canyon locality, a large part of the sandstone is bleached. Bleaching is most prominent in the footwall, where deformation bands occur more frequently in this fault block compared to the hanging wall. In the hanging wall, fractures and subsidiary faults are dominant and bleaching occurs only close to the fault plane, where there are some clusters of deformation bands. The main fault is orientated NW–SE, dipping towards the south to SW. Deformation bands and fractures are orientated parallel to sub-parallel to the main fault (fault 1: Fig. 7).

The frequency and cumulative plots (Fig. 7c, d) demonstrate local variations in the frequency of deformation bands in the footwall damage zones. The high frequencies are in most cases associated with deformation band clusters. These clusters are up to 25 cm thick and occur more frequently in the extensive bleached footwall damage zone up to 30 m away from the main fault (fault 1: Fig. 7). This has resulted in a 10 m inner and a 20 m outer damage zone in the footwall of fault 1. In the hanging wall, where the fractures are dominant in the Dewey Bridge and Slick Rock members, the inner and outer damage zones of fault 1 are wider, up to 16 and 78 m, respectively (Fig. 7).

**Humbug Flats.** The Humbug Flats are in the northeastern flank of the San Rafael Swell (Loc. 4 in Fig. 2), where studied faults affect the Entrada Sandstone. Here, data were acquired along three 55-m-long scanlines to study the hanging wall of a master fault. Three subsidiary faults were measured along the scanline, of which faults 1 and 2 are in the damage zone of the master fault (Fig. 8). The higher frequencies are related to the presence of deformation-band clusters along the scanlines. The damage zone in the bleached rocks in scanline 3 shows a higher frequency of deformation bands, and wider inner and outer damage zones (12 and 35 m, respectively) compared to those in the unbleached rocks (Fig. 8). The inner and outer damage zones are approximately 7 and 22.5 m wide in scanline 1, and 10 and 33.5 m in scanline 2, respectively (Fig. 8).

Most fractures and deformation bands strike parallel to the studied faults trending NW–SE, dipping either NNE or SSW (see Fig. A1 in Appendix A). Fractures, however, show a larger variation in orientation between different scanlines, in which they strike both synthetic and antithetic to the faults.

**Majella outcrop data**

Data were gathered along scanlines outside the Val lone Santo Spirito on the eastern section of the Majella Monocline forelimb and was continued on both sides of the valley as long as the outcrops were accessible. Twenty-four scanlines were measured at each selected location in the valley, comprising 12 sets (see Appendix A), each with a baseline and a second scanline at 2 m elevation. All data were collected from the Morrone di Pacentro Formation, which represents the lower unit of the carbonate platform. We measured and documented deformation features such as fractures, pressure-solution seams and five fault sets (see Fig. A2 in Appendix A) in the carbonate deposits. Fault classification is based on the examination of slip surfaces and displaced marker beds, as well as their orientations (see Fig. A2 in Appendix A), which are consistent with previous work from the area and can be used to determine the fault types (Di Cuia et al. 2009; Aydin et al. 2010). The five fault sets studied in the area include two sets of normal faults, one set of reverse faults and two sets of strike-slip faults in right-lateral (purple great circles in Fig. A2 in Appendix A) and left-lateral orientations (yellow great circles in Fig. A2 in Appendix A). Forty-two faults were captured in these scanlines, from which 19 were normal faults (six pre-tilting and 13 post-tilting), nine were reverse faults and 14 were strike-slip faults (seven right-lateral and seven left-lateral) (see Fig. A2 in Appendix A). The numbering of faults differs from scanline to scanline. A scatter in the orientation of the post-tilting normal faults is also observed, even though the general trend show that most normal faults are strike-parallel with the bedding (Aydin et al. 2010). The relative age of different fault types is also backed up by cross-cutting faults, where normal faults are cut by younger vertical strike-slip faults. The reverse faults have opposing dips to both bedding and the post-tilting normal faults (see Fig. A2 in Appendix A).

Typically, large scatter in the fracture orientations corresponds to the diversity of fault classes within the outcrop. Pressure-solution seams are observed throughout the area (Graham et al. 2003; Di Cuia et al. 2009). The structural complexity varies between different stratigraphic layers, and has probably been influenced by the variation in thickness and lithology (Di Cuia et al. 2009). The carbonates have been dissolved in several cases, leaving behind large cavities (karsts) in the outcrops (Fig. 9). Such features complicated the data gathering process and resulted in large gaps in the scanline dataset. Karsts tend to follow stratigraphic bedding planes and continue in the fault damage zones (Fig. 9).

In several cases, the frequency of fractures increases outside the damage zone and in the distal parts of the fault zones (Fig. 10c, d). This can be attributed to brecciation of the deposit and the occurrence of stratigraphic layers more prone to
fracturing. This leads to an apparently intense deformation independent of faulting. Since most of the fault damage zones interact, we mainly use background fracturing to find the damage-zone width and only used the cumulative plots in a few scanlines to constrain the damage-zone width. Two examples of these are indicated in Figure 10e and f, where the inner and outer damage-zone width in the

**Fig. 8.** (a) A photograph of the outcrop in Humbug Flats showing the three scanlines measured in the damage zone of a master fault. Frequency plots (in columns) with cumulative distributions (in dots) from the Humbug Flats scanline 1 (b), scanline 2 (c) and scanline 3 (d), as explained in the text. The grey lines show the general slope of the cumulative distributions.
baseline and second scanline is estimated to be 5 and 6 m (Fig. 10e), and 7 and 3 m (Fig. 10f), respectively.

Sotra Island outcrop data

In total, nine normal faults were studied (Figs 4 & 11) with dip-slip displacement ranging from a few centimetres to several metres (Torabi et al. 2018). The main set of fractures and faults, which are predominantly filled with epidote, are captured in scanlines 1 and 2 (Fig. 11), orientated ENE–WSW like the set I fractures of Larsen et al. (2003). Fault 2 is the largest fault (with c. 3 m dip-slip displacement) in this scanline, with a strike component of displacement too great to be measured. Fracture frequency increases close to fault 2 (Fig. 11) but drops rapidly at 23 m (the boundary of the footwall damage zone), and again increases as the scanline crosses faults 3–6 (Fig. 11b). Due to these irregularities and the larger number of small faults, the footwall exhibits a greater damage zone (around 8 m wide) than the hanging wall, which is estimated to be of around 3 m wide (Fig. 11b).

Discussion and utilization of the results

Implication for fault modelling

In order to predict the fluid-flow behaviour of a fault, data and knowledge of fault geometrical attributes and fault rock properties are needed. Currently, fault 3D models (ellipsoids) are dependent on the 1D and 2D data of fault cores and damage zones obtained from outcrops, mines or subsurface data from core and image logs allied to seismic data to sample faults without erosion and across the scale range (e.g. Harris et al. 2003; Qu et al. 2015). This is because the scale of detail included in the fault damage zone and core is below seismic resolution.
**Fig. 10.** Frequency plots (in columns) from scanlines 3 (a) & (b), 7 (c) & (d) and 8 (e) & (f) (with cumulative distributions shown in dots) for the Majella outcrop. PSS indicates the location of pressure-solution seams in the scanlines. Note that the locations of faults are identified with F and their numbers as measured in the outcrop. N stands for normal fault; RL and LL stand for strike-slip faults with right-lateral and left-lateral motions. The grey lines show the general slope of the cumulative distributions.
Fig. 11. (a) Overview map of the outcrop studied in the Sotra Island. Fault 3–6 were too small to mark on the map. (b) Frequency plot of cross-fault damage zones from scanlines 1 and 2 perpendicular to the faults. Cumulative distributions (in dots) were used to constrain the width of damage zones around fault 2. The grey lines show the general slope of the cumulative distributions. The rest of faults have an overlapping damage zone. (c) & (d) Orientation data of the faults and fractures at this locality. Modified after Torabi et al. (2018).
although there have been studies on capturing damage zones on seismic data using seismic attributes (e.g. Alaei & Torabi 2017; Botter et al. 2017). Commonly, only limited segments of a fault are present in outcrops. This provides only a snapshot of the geometry of the entire fault. In most cases, it would be hard to study the variation in displacement and damage-zone width of a fault as large parts of them are usually eroded and/or not accessible.

Our data gathering is limited to the lowermost 2 m (baseline and the second scanline) of the vertical outcrops in most cases. Although the results could not be representative of the entire fault height, it can show the details of damage zone structure and the variations in both damage-zone width and their corresponding displacements at the level of measurements. These variations are related to different deformation mechanisms in rocks, changes in the geometry of the fault during fault propagation and different stages in the faulting process.

The variability in the damage zone affects its fluid-flow properties and needs to be considered when modelling faults (Rohmer et al. 2015). The details of the damage zone internal structure and width through the statistical analysis conducted in this work would help to constrain the size of the fault ellipse in the models. This would result in a more data-driven approach in the future stochastic fault modelling.

The type and frequency of deformation features, and the width of the damage zone, will affect the permeability of rocks, introducing both heterogeneity and anisotropy to the faulted rocks. The across- and along-fault flow paths will be dependent on the geometries and permeability characteristics of the damage adjacent to the fault core. In this study, only measurements of deformation frequency across the fault are provided, which can reflect the permeability changes across the fault depending on the connectivity of the deformation features. The anisotropy in permeability can be considered in the fault ellipse when simulating fluid flow in fault models (e.g. Caine et al. 1996) by applying different across-fault permeability (the minor axis of the ellipse) and fault-parallel permeability (the major axis of the ellipse). The across-fault permeability is usually lower than the fault-parallel permeability (by up to two orders of magnitude) and will significantly impact the fluid flow across the fault (Caine et al. 1996; Jourde et al. 2002).

Effect of lithology, rock properties and fault geometry

The presence of different deformation mechanisms within damage zones can be attributed to the lithology and mechanical properties (such as elasticity or Young’s modulus), as well as the petrophysical properties (such as porosity and permeability), of the host rock under deformation. In Utah, several studies report variations in fault damage-zone width along different segments of the Moab Fault as a result of changes in the type and frequency distribution of deformation features (fractures and deformation bands) in layers of different mineralogy, grain size and porosity and along fault bends, overlapping and linking segments (Shipton & Cowie 2001; Davatzes et al. 2005; Johansen & Fossen 2008). A porous sandstone with lower Young’s modulus, in comparison to a low-porosity rock, responds to stress by grain reorganization and breakage, resulting in different types of deformation bands. A low-porosity rock, such as a low-porosity carbonate rock, usually responds to stress by fracturing (e.g. Torabi & Zarif 2014). This may result in different deformation patterns and asymmetrical damage zones in the footwall and hanging wall of faults (Berg & Skar 2005). Fault damage zones tend to widen in fault bends and relay ramps (Davatzes et al. 2005).

The higher frequency of discrete small faults accommodating distributed displacement in the damage zone of the Moab Fault segment at ANP could be related to fault bending. In the Hidden Canyon locality, a large fault segment with displacement of up to 200 m has cut through the rocks, and in the footwall a set of deformation bands are developed with millimetre to centimetre displacement that extends sub-parallel to the fault core with an intensity that decreases away from the core. While in the hanging wall, fractures representing the dominant deformation features and the syncline structure (i.e. drag fold) have accommodated most of the deformation (strain), resulting in an asymmetrical deformation pattern and damage zone. In the footwall of this fault (Entrada Sandstones), the Moab Tongue Member is more porous than the Slick Rock Member, which causes changes in the mechanical properties of the rocks (mechanical stratigraphy). This has resulted in the higher frequency of deformation bands (up to 50) in the Moab Tongue Member (Fig. 5c). In addition, even within different layers of the Slick Rock Member, there is a variation in the type and frequency of deformation bands depending on the porosity and grain size of the rock (Johansen & Fossen 2008).

The widest damage zone measured in the Entrada Sandstone is estimated to be in the bleached rocks of the Moab Tongue (75 m for the inner and outer damage zones together), whereas the hanging-wall damage-zone width is approximately 169 m. In addition, the presence of a drag fold (syncline) in the hanging wall of the fault in the Hidden Canyon could have caused the asymmetrical strain distribution within the damage zones in this locality (Berg & Skar 2005).
Another example of the lithology effect is observed in the Cache Valley (Loc. 3 in Figs 2 and Fig. 7), where hanging-wall rocks of reddish Dewey Bridge and Slick Rock members of the Entrada Sandstone are dominated by fractures, while the porous Navajo Sandstone in the footwall is dominated by deformation bands. The damage zone in the fractured hanging wall is wider than the damage zone dominated by deformation bands in the footwall. A similar observation was made in the Majella Mountain where different faulted sedimentary successions show variations in deformation intensity. When several fault types and layers with variations in mechanical properties and thickness are crossed over a single scanline, the resulting dataset can show large fluctuations in the frequency of deformation features. This aspect needs to be considered when interpreting the data.

In the carbonate rocks of Majella (Figs 3 & 9), the dissolution of carbonates is dominant in areas where water flows through the rock, such as carbonates with high primary and secondary porosity (fractures) which is typical of damage zones (Ford & Williams 2013). Planes and surfaces such as bedding, pressure-solution seams or faults affect the flow direction and, in some cases, work as barriers or conduits and channel fluids along specific paths. In the studied outcrops of Majella (Fig. 9), large karsts normally occur in the damage zone (Micarelli et al. 2006; Agosta et al. 2010a). Therefore, karst topography provides proof of fluid-flow processes in the rocks and can give information on the petrophysical properties of the faulted rocks. The brecciation in the damage zone would also contribute to fluid flow in carbonate rocks unless the breccia is cemented.

Based on our measurements of the studied faults in the Majella area, the damage zones of normal faults are the most symmetrical ones, in which the frequency of fractures decreases away from the fault in a similar and symmetrical manner in both the hanging wall and footwall. In Majella, the thickness of host-rock layers intersected by the fault also influences the frequency of deformation features. Thin beds tend to host a higher number of fractures than thicker beds in carbonates (Micarelli et al. 2006; Di Cuia et al. 2009). The occurrence of such thin and highly deformed layers often leads to complex frequency plots.

**Bleaching**

Bleaching of sandstone, a feature frequently encountered in outcrops studied in Utah (Hidden Canyon, Cache Valley and the Humbug Flats), correlates to the type and intensity of deformation features found in the outcrop. The bleaching is observed in both host rocks and the fault rocks within the fault core and the damage zone. Bleaching can be a result of the removal of the red hematite coating of the sand grains, leading to a change in colour from red brownish to light brownish or sometimes white. Reducing fluids containing CO₂ or oil might originate either from salt structures in the Paradox Basin or a monocline related to the Laramide uplift in the San Rafael Swell (Parry et al. 2004; Kampman et al. 2014). Fluid flow favours permeable paths and while the sandstone may be permeable, the reducing fluids sweep the permeable sandstones and flow along conduits, such as permeable fault planes and damage zones, washing out the iron oxide coating (Chan et al. 2000; Parry et al. 2004; Kampman et al. 2014). The bleaching patterns may indicate the ancient fluid-flow path and is evidence of hydrocarbon migration (Garden et al. 2001; Parry et al. 2004).

Therefore, the bleached zones usually have a higher permeability than the unbleached zones of similar rocks. Accordingly, the bleached areas host zones of a greater number of deformation bands and their clusters in the Entrada and Navajo sandstones studied in the Cache Valley and Humbug Flats compared to unbleached zones in a similar lithology (Figs 5, 7 & 8). Parry et al. (2004) discussed the presence of bleached (white) deformation bands in bleached sandstone and interpreted them as having formed after the bleaching of the sandstone. This could be the case for some of our case studies since the deformation bands are mostly bleached in the studied bleached sandstones. On the other hand, in the areas where bleaching is concentrated around the faults, one can argue that bleaching has been associated with breached traps. Hence, the bleaching could have occurred at the same time as faulting and the formation of deformation bands.

**Univariate and bivariate distributions of fault damage-zone width and displacement**

The distribution of displacement and damage-zone width has been individually studied by exceedance frequency plots. Data from three different localities (Utah, Majella and Sotra Island) are plotted together with data from the literature (for the references, see the caption to Fig. 12), which mainly includes normal faults in siliciclastic rocks and some tight rocks with low porosity from Choi et al. (2016) and references therein. When examining the exceedance frequency (EF) on log–log plots, the power law is the best fit to the data with a high coefficient of determination (Fig. 12a, b). The tails (highest and lowest values) are excluded due to resolution effects (Fig. 12). The EF plot of the displacement data shows a two-slope power-law distribution (change point at c. 100 m displacement) with a high coefficient of determination (Fig. 12a). This is also true for the damage-zone width data, where the change
in the slope of the power law occurs at around a 7 m-wide damage zone. Kolyukhin & Torabi (2013) investigated the distribution of damage-zone width data for normal faults in different lithologies and found that a truncated power law with two slopes fits best to the data. The changes in the slope of the power-law distributions for both displacement and damage-zone width indicate a hierarchy in the fault systems (Torabi & Berg 2011). This would imply a change in the fault-growth mechanism as accumulated displacement increases (greater than c. 100 m) in larger and more mature faults with linking segments.

The relationship between displacement and damage-zone width is investigated and compared to the previous data (Fig. 13). Such large datasets incorporating different case studies will reduce the bias and errors related to local restrictions at specific localities. This plot can therefore give a more representative relationship between the fault attributes over a wide range of scales. Despite the large number of points, there is still a lack of data from large faults (>100 m displacement). A power-law relationship between displacement and damage-zone width would fit these data and support a hypothesis (model) for a mutual evolution of both parameters – it can be used to predict the size of one attribute based on the knowledge of the other. Choi et al. (2016) reported a power-law relationship between displacement and damage-zone width when damage zones are dominated by either fractures ($R^2 = 0.70$) or deformation bands ($R^2 = 0.76$). We have, however, mixed the data from different fault types and lithologies, which might affect the statistical relationship between damage-zone width and displacement. Our compiled data can be divided into two parts: a power law ($R^2 = 0.73$) fits well to plotted data up to around 100 m displacement (Fig. 13); however, there is a change in the distribution of data, where the slope of a possible power-law relationship is expected to decrease in the second part of the data (>100 m displacement) (Fig. 13).

There is a need for enough data points for larger faults in the second part of this plot in order to find a
statistically valid relationship. Seismic studies that provide such data from seismic-scale faults could help in assessing the relationship between damage-zone width and displacement for larger faults (e.g. Alaei & Torabi 2017). The change in the slope of this plot indicates a hierarchy and stepwise growth of faults for all fault types forming in different lithologies (Torabi & Berg 2011; Kolyukhin & Torabi 2012). The change in the relationship between fault displacement and damage-zone width can be attributed to the timing of deformation and the faulting process. In siliciclastic rocks of Utah, deformation bands and their clusters are precursor structures to faulting, in which they show an increased frequency in the fault relays (Davatzes et al. 2005). With increasing strain, patches of slip surfaces form adjacent to the clusters and eventually link up to create a through-going fault (Torabi & Zarifi 2014). In addition, joints and shear fractures occur at intersections and extensional relays in Utah, which eventually contribute to the formation of a through-going fault. In carbonate rocks of Majella, faults in several cases form by shearing of pre-existing pressure-solution seams or shearing along the bedding (Aydin et al. 2010), which can reflect the stepwise growth of damage zones. Similar implications can be applied to fractures and fracture zones in metamorphic rocks of Sotra that result in faulting.

The stepwise nature of fault growth in all three types of the lithologies (siliciclastic, carbonate and basement rocks) can be related to the stepwise distribution of fault displacement and width, as well as the relationship between these two fault geometrical attributes, as shown in Figures 12 and 13. It could be that damage zones of faults with displacement larger than 100 m cannot get wider indefinitely because the faults are localized and have the planarity required to accumulate throw most efficiently. Some of the damage observed in the walls of the fault may relate to processes operating ahead of the propagating fault tip. Hence, these are related to the mechanical properties at the time of faulting within individual parts of the mechanical stratigraphy and are not necessarily related to the later wall damage as further strain is accumulated on the fault.

Conclusions

The outcomes of this study are:

- We constrained damage-zone width by applying cumulative distributions on the frequency plots for isolated faults and background frequency in complex fault zones and overlapping damage zones.
- Bleached zones in the Utah siliciclastic rocks are associated with a greater frequency of deformation

![Fig. 13. A stepwise relationship between displacement and damage-zone width with a power law as the best fit to the first part of data up to around 100 m displacement. For previous data, see the references cited in the caption to Figure 12.](image-url)
bands compared to the unbleached zones. This causes a wider damage zone in these zones compared to the unbleached rocks of similar lithology.

- Fault damage zones in the carbonate rocks of Majella are often host to open fractures (karst), demonstrating that they can also be conductive to fluid flow.
- Variable deformation mechanism type and asymmetrical damage zones of normal faults in siliciclastic rocks inferred from the extent of the damage zone are common where rocks of different lithologies and properties are exposed in the footwall and hanging-wall damage zones.
- A univariate distribution of damage-zone width and displacement data combined from different lithologies (siliciclastics, carbonates and gneiss) and fault types (normal, reverse and strike-slip faults) follow two-slope power-law distributions with change points at c. 100 m displacement and 7 m damage-zone width.
- Integrating data from this research with previously presented results in Figures 12 and 13 yields a stepwise power-law relationship between displacement and damage-zone width. The result implies a relationship over a wide range of scales and lithology that can be used to predict the dimensions of one of the attributes based on knowledge of the other.

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Appendix A

Figure A1 shows orientation data for the Humbug Flats locality in Utah, USA and Figure A2 shows the orientation data for the Majella locality in Italy.

Fig. A1. Orientation of fault, fractures and deformation bands for the Humbug Flats locality in Utah, USA.
Fig. A2. (a) Google Earth image of Vallone Santo Spirito, Majella, Italy with marked scanline positions and stratigraphic layers. Michelin Traffic ©2006–2015 TomTom ©Bing. (b) Orientation data for different faults based on fault type in Majella, Italy. The pre-tilting normal faults are easily distinguished from the post-tilting normal faults by their orientation. The two sets of strike-slip faults show, in some cases, larger similarities in orientation but the left-lateral (yellow lines) strike-slip faults show a more continuous pattern.

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