Journal of Structural Geology 31 (2009) 117–127

Contents lists available at ScienceDirect

Journal of Structural Geology

journal homepage: www.elsevier.com/locate/jsg

A geometric model of fault zone and fault rock thickness variations

Conrad Childs [a,*], Tom Manzocchi [a], John J. Walsh [a], Christopher G. Bonson [b], Andrew Nicol [c], Martin P.J. Schöpfer [a]

[a] Fault Analysis Group, UCD School of Geological Sciences, University College Dublin, Belfield, Dublin 4, Ireland
[b] SRK Consulting (UK) Limited, Fifth Floor, Churchill House, Churchill Way, Cardiff, CF10 2HH, UK
[c] GNS Sciences, PO Box 30368, Lower Hutt, New Zealand

ARTICLE INFO

Article history:
Received 5 March 2007
Received in revised form 25 August 2008
Accepted 28 August 2008
Available online 27 September 2008

Keywords:
Fault zone
Fault rock thickness
Fault damage zone
Segmentation

ABSTRACT

The thicknesses of fault rock and fault zones and the fault normal separations for breached and intact relay zones each show a positive correlation with fault displacement. The displacement to thickness ratio, or average shear strain, varies for the different structures increasing from intact relay zones (median value = 0.27) to fault rocks (median value = 50). The correlation for fault rocks is widely interpreted as a growth trend controlled by fault rock rheology, but the progression of displacement to thickness ratios for the different structures suggests an alternative model. In this alternative model a fault initiates as an array of irregular fault segments. As displacement increases, relay zones separating fault segments are breached and fault surface irregularities are sheared off, to form fault zones containing lenses of fault-bounded rock. With further displacement these lenses are progressively comminuted, and ultimately converted to zones of thickened fault rock. The final fault rock thickness is therefore influenced strongly by fault structure inherited from the geometry of the initial fault array. The large scale range on which fault segmentation and irregularities occur provides the basis for application of this model over a scale range of at least 7 orders of magnitude.

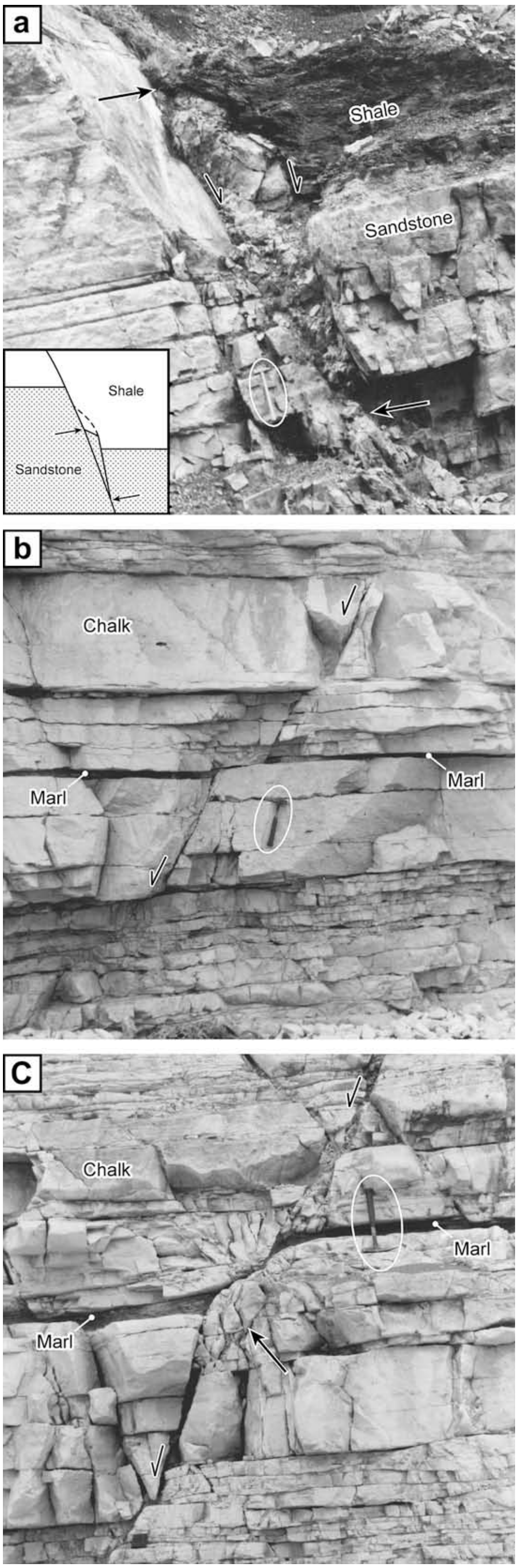
1. Introduction

Faults are zones of extreme internal complexity and heterogeneous strain distribution over a wide range of scales. Although this complexity does not lend itself to a simple description to which all faults conform, a simplified and generalised description of faults is required to achieve a better understanding of fault evolution and for many practical applications, such as the production of oil from faulted reservoirs and earthquake hazard assessment. Although the stages in development of a fault are reasonably well established, the fault geometric components characteristic of these stages (e.g. relay zones, damage zones, fault zones, fault rocks, etc.) are not generally combined in outcrop descriptions of fault architecture. In this paper we collate thickness measurements of fault rocks, fault zones, relay zones and damage zones and relate these to different stages of fault growth. We use the collated data to propose a semi-quantitative description of fault structural evolution indexed to ranges of average shear strain intensity associated with these different geometric components.

The geometric component of faults to have received most attention in the published literature is fault rock thickness. A broad positive correlation between fault displacement and fault rock thickness is well established (Robertson, 1983; Hull, 1988; Marrett and Allmendinger, 1990), with particularly well defined trends sometimes being used to infer fault displacements from measurements of fault rock thickness (Marrett and Allmendinger, 1990; Little, 1995). Given that fault rocks are, with rare exceptions, weaker than the surrounding wall rocks, the cause of a continual increase in fault rock thickness with increasing displacement is not self evident and a number of fault rock widening models have been proposed. These widening models typically rely upon shearing and attrition of fault wall rock (Hundley-Goff and Moody, 1980; Robertson, 1982; Scholz, 1987; Power et al., 1988) or strain hardening of fault rocks (Hull, 1988; Faulkner et al., 2003). These models are based on the premise that the positive correlation between fault displacement and thickness represents a growth trend and that a low displacement fault will progressively widen with increasing displacement (Robertson, 1982; Scholz, 1987; Hull, 1988). With the exception of the model of Power et al. (1988), fault rock thickness within these models is directly controlled by the rheological properties of the wall rocks and fault rocks without considering the extent to which fault geometry may control fault rock development. In this paper we propose an alternative model in which the thickness and distribution of fault rock are largely controlled by fault geometry. While fault geometry may to a significant extent be controlled by the lithological and rheological stratigraphy of the wall rock, for

* Corresponding author.
E-mail address: conrad@fag.ucd.ie (C. Childs).

0191-8141/$ – see front matter © 2008 Elsevier Ltd. All rights reserved.
doi:10.1016/j.jsg.2008.08.009



example, due to fault dip changes (refraction) across bedding planes, we suggest that in other respects the rheological properties of the fault rock and the wall rock *lithologies* are of secondary importance. We argue that internal fault zone geometry and the distribution of fault rock thickness over the fault surface are strongly influenced by the locations and dimensions of steps or bends of the fault surface produced during fault propagation.

Our model for the development of fault rock is underpinned by a model for the internal geometry of faults and well documented processes of fault zone evolution. The geometric model adopted here recognises that faults are not simple planar features but are highly complex zones within which displacement and strain are concentrated onto one or several discrete slip surfaces or zones of intense shearing, enclosing variably strained rock volumes (Wallace and Morris, 1979; Cox and Scholz, 1988; Childs et al., 1997). This complexity of structure is largely the result of modification of fault surface geometric irregularities inherited from the initial propagation geometry of a fault (e.g., Childs et al., 1996b). When a fault propagates through a rock volume it does so as an irregular and segmented surface (Naylor et al., 1986; Mandl, 1987; Cox and Scholz, 1988; Huggins et al., 1995; Marchal et al., 2003; Walsh et al., 2003). Fault segmentation is observed both in map view and in cross-section on a wide range of scales (Tchalenko, 1970; Larsen, 1988; Morley et al., 1990; Peacock and Sanderson, 1991; Walsh et al., 1999) and for each mode of faulting (Dahlstrom, 1969; Aydin, 1988; Larsen, 1988; Woodcock and Fischer, 1986). The rock volumes between adjacent kinematically related fault segments are zones of high strain (Chadwick, 1986; Walsh and Watterson, 1991; Childs et al., 1995) that are variously referred to as transfer zones, relay ramps, fault bridges and steps. Here we follow Walsh et al. (1999) and use the term 'relay zone' to refer to any such rock volume between kinematically related fault segments irrespective of the mode of faulting or the nature of the strain (contractional, extensional or constant volume, i.e. neutral).

As displacement increases on a segmented fault, the strains at relay zones increase, eventually causing failure of the relay zone by the formation of a linking fault to form a 'breached relay zone' (Peacock and Sanderson, 1994; Childs et al., 1995; Ferrill et al., 1999; Ferrill and Morris, 2001; Soliva and Benedicto, 2004; van der Zee and Urai, 2005). The site of the breached relay zone will remain as an irregularity or asperity on the new through-going continuous fault surface, and both the asperity and the continuous fault may be bypassed at a later stage. The former relay zone may become incorporated into the fault, initially as a fault-bounded lens and ultimately as a zone of thickened fault rock, which may be entrained along the fault surface in the displacement direction. Similarly, irregularities or asperities on a continuous surface will be sheared off, initially forming fault-bounded lenses but ultimately becoming comminuted to fault rock (Fig. 1). The simple model of

Fig. 1. Outcrop photographs of fault-related structures associated with asperity removal and fault segmentation. (a) Photograph of a 9 m throw normal fault contained within the Upper Carboniferous sandstones and shales of Round O Quarry, Lancashire, UK. The fault dips to the right and the contact between the sandstone and overlying shale is visible in the fault hanging wall. A fault-bounded triangular wedge of sandstone at the centre of the photograph is interpreted to have been derived from the uppermost part of the footwall sandstone unit by removal of an asperity caused by a change in fault dip across the sandstone-shale interface (see inset). The motion of the wedge down the fault has been accompanied by intense fracturing of the component sandstones and breccia formation where the lens tapers downwards. (b and c) Segmented faults from the chalk and marl sequences of Flamborough, Yorkshire, UK. (b) A segmented fault steps across a thin marl unit (above and to the left of the hammer) giving rise to an irregularity which represents an asperity to fault movement. (c) A region of intense fracturing adjacent to a bend in a fault. This fault is interpreted to have had an initial geometry similar to (b) with increased displacement causing fracturing due to concentration of strain at an initial asperity centred on the offset marl unit. With further increase in displacement the fault bend may be by passed by continued movement on the slip-surface arrowed.

fault architectural evolution presented here is one of progressive strain localisation resulting in a reduction in the area of active fault surface by bypassing areas of overlapping fault surfaces and fault surface irregularities (Ferrill et al., 2001; Walsh et al., 2001).

In this paper we collate thickness data for the structures characteristic of the stages of fault zone growth (i.e. fault rock, fault zone, relay zone and damage zone). The data, which are from brittle faults offsetting a range of rock types with displacements ranging over 7 orders of magnitude, provide a semi-quantitative description of fault zone geometry and geometric evolution that is consistent with systematics of fault rock scaling and distribution.

2. Data and terminology

Study of the evolution of fault thickness has been hampered by terminology problems. Terms such as fault rock, fault zone and damage zone are not uniquely defined and there are no unambiguous rules for measuring the thicknesses of these features. It has become common in recent years to describe fault zones as comprising a fault core, containing slip surfaces, gouge, cataclasites and breccias, and a damage zone comprising subsidiary minor structures e.g. veins and small faults (Caine et al., 1996). In this paper we do not use the two component core/damage zone description but have extended our fault descriptions to include other data categories. The correspondence between the core/damage zone terms and terms used here is shown in Fig. 2. The terms used in this paper are each described in this section.

For the purposes of data collation we have used the term *fault rock* to refer to fault gouge, breccia and cataclasite. Field measurement of breccia thickness can be subjective as breccias have a broad continuum of clast sizes and there is no rigidly defined size cutoff distinguishing breccia clasts from fault-bounded rock volumes (Marrett and Allmendinger, 1990). Problems comparing breccia/gouge thickness data from different sources are therefore inevitable (Evans, 1990). We have, where possible, minimised these problems by referring to descriptions in the source articles. Data from Knott et al. (1996), for example, have been classified here as fault rock thickness, although in the original article, they are referred to as fault zone thickness. The definition provided by Knott et al. (1996) in their article is "the zone where most fault slip has occurred and usually includes the slip surfaces and the band of fault gouge and … cataclasis" and, since it does "not include undeformed blocks entrained in the fault zone", corresponds to the definition of fault rock used here. For the most part we have not discriminated fault rocks according to the wall rocks from which they are derived, we have however, assigned deformation bands, which are characteristic of faults in porous sandstones, to a separate fault rock category for reasons discussed below.

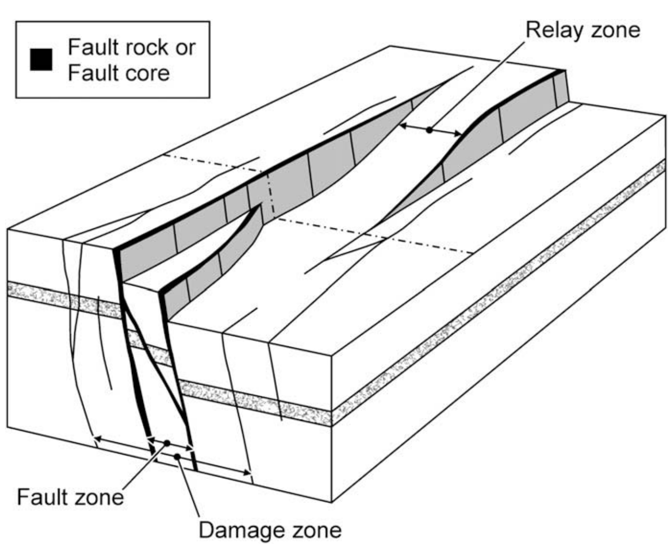


Fig. 2. Schematic diagram comparing the terms fault rock, fault zone and relay zone used in this paper with the fault core/damage zone description. Fault rock is synonymous with fault core *sensu* Caine et al. (1996). At the centre of the block diagram (along the dashed line) the bulk of the displacement is accommodated on a single slip-surface and therefore the fault zone thickness is equal to the thickness of the fault rock/fault core. Thin lines indicate faults with minor displacements.

Fault zone thickness measurements, with the exception of data from Wolf (1985) and van der Zee (2002), have been collected by the authors (Childs et al., 1996a; Childs, 2000; unpublished Fault Analysis Group data) from outcrop and principally from cross-sections through normal faults. The term *fault zone* has recently been defined as a system of related fault segments that interact and link and are restricted to a relatively narrow band or volume (Peacock et al., 2000). Adapting this definition to one which can be used as a guide to thickness measurement in the field, we define fault zone thickness as the distance between synthetic slip-surfaces (same dip direction and sense of offset) that can be demonstrated at outcrop to be kinematically related and which each accommodate at least a few percent of the total offset. Slip-surfaces have been judged to be kinematically related when they are connected at a branchpoint or when they bound regions of elevated strain (e.g., fault rock, zones of bed rotation, clusters of antithetic faults). The displacement required to discriminate fault zone bounding slip surfaces from minor faults within a damage zone cannot be rigidly defined since it depends on a number of factors such as fault style, fault density or the extent of outcrop; perhaps the most common manifestation of a fault zone is an anastomosing network of through-going synthetic slip surfaces and associated fault rock. Examples of structures which have been measured as fault zones are shown in Fig. 3a and b.

Relay zone measurements are derived exclusively from normal faults and comprise measures of fault displacement and relay zone separation, i.e. the fault normal distance measured between a pair of relay zone bounding faults. The data are derived from the map view geometries of relay zones from outcrop studies and interpreted 3D seismic reflection surveys, and include both intact relay zones and breached relay zones in which one, or both, relay bounding faults has propagated to intersect its partner fault.

Since particular outcrop conditions are required for unambiguous identification of a relay zone, many relay zones will instead be characterised as fault zones. For example, the structures in Fig. 3a and b may represent cross-sections through relay zones, but this cannot be established from the available outcrop data; instead the structures are taken to be fault zones. By contrast, the fault in Fig. 3c displays a similar internal structure to those in Fig. 3a and b, but because it is bounded by two slip-surfaces which both tip-out within the outcrop, the structure can clearly be identified as a relay zone. The structure shown in Fig. 3c is a contractional relay zone, i.e. transfer of displacement between the relay zone bounding faults requires contractional strains. Unlike the structure in Fig 3c, it is seldom possible to identify relay zones on normal faults in cross-section so that the available data for contractional and extensional relay zones are limited (but see Walsh et al., 1999; Kattenhorn and Pollard, 2001; Kristensen et al., 2008) and these data (other than Fig. 3c) have not been included in this study.

Measurements of damage zones are derived from the literature. A *damage zone* is generally taken as the volume of deformed wall rocks around a fault surface that results from the initiation, propagation and build-up of slip along faults (Kim et al., 2004). Measurement of damage zone thickness, which excludes the core thickness (Caine, et al., 1996), relies on defining a zone within which fault density is higher than the background density, something which, depending on the degree of strain heterogeneity, may

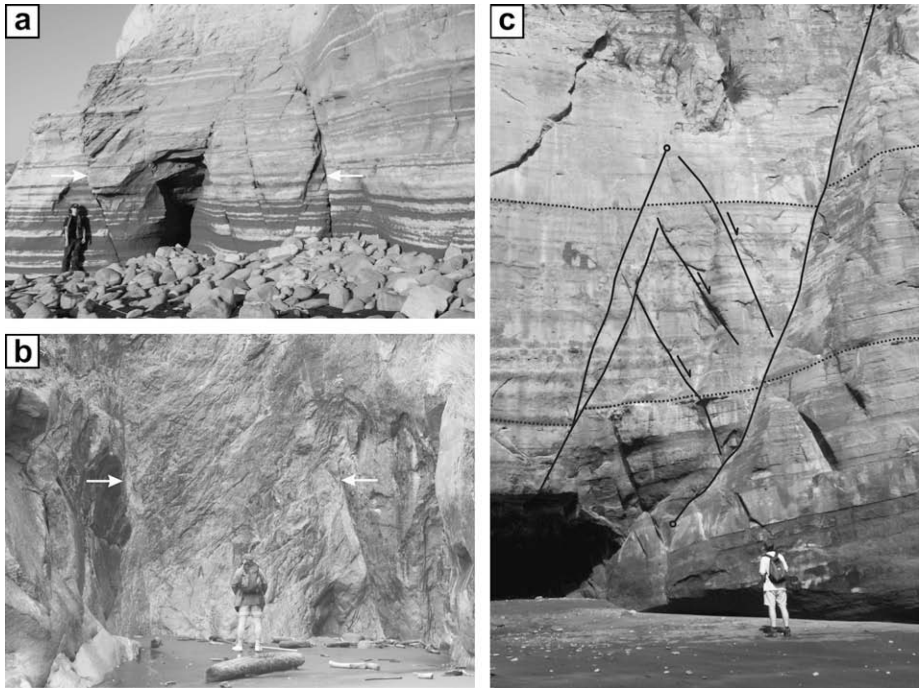


Fig. 3. Outcrop photographs of normal faults offsetting a weakly lithified mixed clastic sequence at Taranaki, New Zealand. The main slip-surfaces bounding the fault zones in (a) and (b) are arrowed and downthrow to the right. The total throw across the fault zone in (a) is 1 m and across (b) is some 10s of metres. In (c) two slip-surfaces bound a zone of bed rotation and antithetic faulting. The upward decrease in throw on the hanging wall slip-surface and the complementary upward increase in throw on the footwall slip-surface identify this structure as an intact relay zone. Individual fault segments are shown in black and two bedding horizons are shown with dotted lines. The total throw on the two relay zone bounding fault segments is 0.4 m.

be difficult to establish. The published measurements of damage zone thickness often include features which, by our scheme would be classified as fault zones or relay zones, and are predominantly for faults within porous sandstones.

The fault components defined above do not incorporate the full range of thickness measurements required to define fault geometry. For example, the width of the zone of normal drag adjacent to a fault is not included, principally because these data are not yet available in the literature and neither are we aware of any published study which investigates the relationship between drag wavelength and displacement. Another and more important measure in the context of this paper, would be the amplitudes of fault surface asperities (i.e. irregularities) and their variation with displacement and scale. Although measurements of fault surface geometry have been published (Power et al., 1987, 1988; Sagy et al., 2007), these data are relatively sparse and are not, in any case, readily converted to asperity wavelength to displacement ratios (see Section 5).

It is clear from the above that the definitions of the various fault components for which we have collated data are often ambiguous and measurements of their thicknesses can be subjective. Although the semi-quantitative model of fault evolution described below recognises the spectrum of structures which occurs (from intact relays through to fault gouge), because of the subjective nature of fault component definition and measurement, the model defines the progression of structures in terms of strain intensity rather than attempting to define precise ranges of strain magnitude for the different fault components.

3. Fault component thicknesses

3.1. Fault rock

The fault rock thickness data (Fig. 4a) demonstrate the well established correlation between thickness and displacement over seven orders of magnitude. The dataset has a median displacement:thickness ratio of 50:1 (Fig. 5). The range of fault rock thicknesses measured on a single fault trace is as much as 2.5 orders of magnitude for a given displacement. There is a decrease in the spread of thickness data from 3 orders of magnitude at low displacements to 2 orders of magnitude at high displacements (>10 m). This reduction in spread is, at least partly, a sampling artefact due to the decrease in the proportion of the fault surface observed in outcrop for large faults compared with small faults.

Fault rock data derived from high porosity sandstones are presented separately in Figs. 4b and 5, as these data tend to obscure the displacement:thickness correlation for fault rocks developed in other lithologies. At small displacements, faults in high porosity sandstones are zones of porosity loss with or without grain size reduction. Individual deformation bands typically have offsets less than 0.5 cm and are up to 0.5 cm thick. Increase in fault displacement is preferentially accommodated on newly formed deformation bands rather than by reactivation of existing structures. This behaviour has been taken as evidence of strain hardening in deformation bands (Aydin and Johnson, 1983; Antonellini and Aydin, 1995), although laboratory experiments have replicated the development of multiple deformation bands without any evidence of strain hardening (Mair et al., 2000). Due to the non-reactivation of deformation bands, the displacement:thickness ratios (D:T) for low displacement faults (<10 cm) in high porosity sandstones are significantly lower than for faults in other rock types, and the median D:T ratio is closer to that measured for fault zones than fault rocks (Fig. 5). This difference becomes less marked at higher displacements when deformation band formation is superseded by the formation of a through-going slip-surface on which subsequent movement is focused (Jamison and Stearns, 1982; Antonellini and Aydin, 1995).

Fault rock thickness data are derived from a wide variety of sources and include the three modes of faulting in lithologies ranging from relatively unconsolidated mixed clastic sequences to

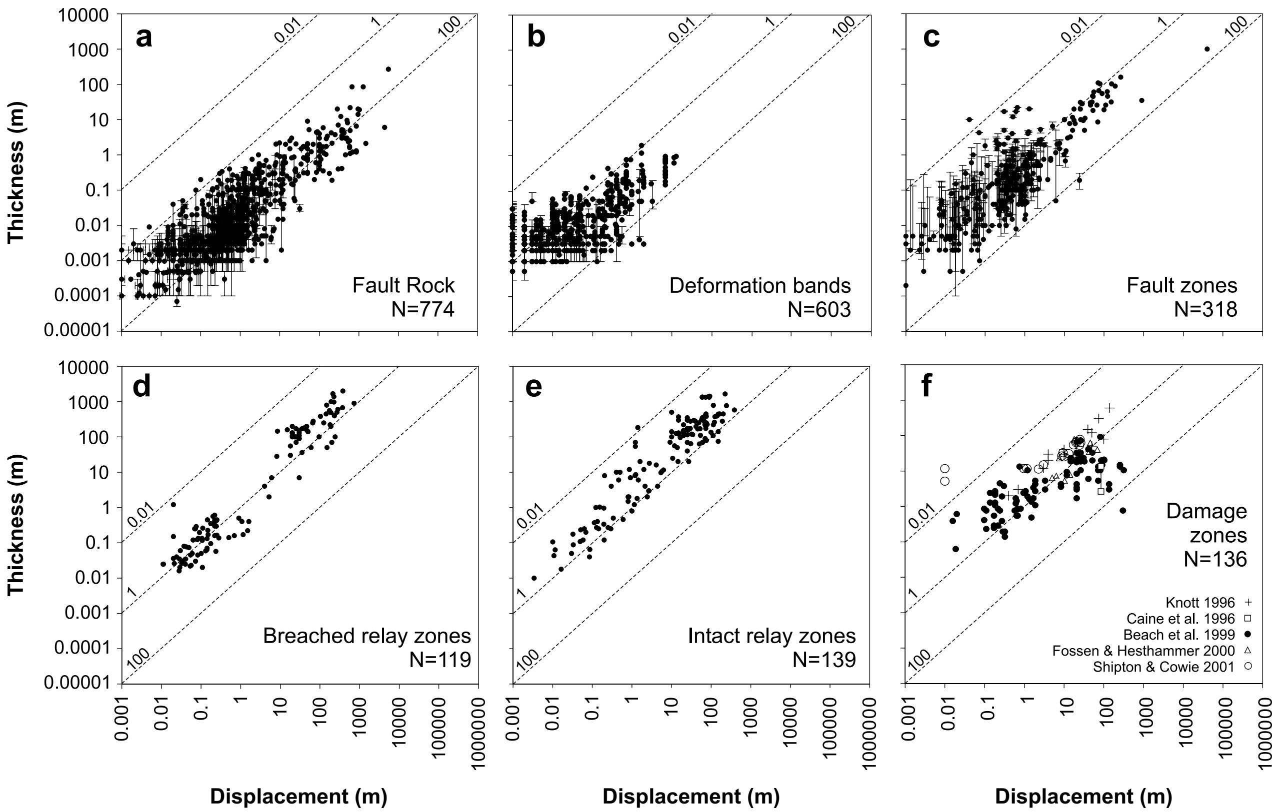


Fig. 4. Displacement versus thickness data for the different fault geometric components identified. Vertical bars are thickness ranges measured on individual faults on decametre scales. The data are derived from the following sources. (a) Fault rock data; Otsuki (1978), Robertson (1983), Hull (1988), Blenkinsop (1989), Marrett and Allmendinger (1990), Ameen (1995), Little (1995), Caine et al. (1996), Knott et al. (1996), Childs et al. (1996a), Foxford et al. (1998) and Childs et al. (2007). (b) Deformation bands; Knott (1994) and Manzocchi (1997). (c) Fault zones; Wolf (1985), Childs et al. (1996a), Childs (2000), Faulkner and Rutter (2001) and van der Zee (2002). Relay zones; breached (d) and intact (e), Imber et al. (2004), Soliva and Benedicto (2004) and Worthington (2006). (f) Damage zones; Caine et al. (1996), Knott et al. (1996), Beach et al. (1999), Fossen and Hesthammer (2000), Shipton and Cowie (2001). Knott et al. (1996) and Beach et al. (1999) record damage zone thicknesses for one side of the fault only. These have been doubled in this compilation. The fault rock, fault zone and deformation band data also include previously unpublished data for normal faults from a range of lithologies. Note that the data in Fig. 4b, and to a lesser extent data shown in Fig. 4a, are influenced by a tendency to round up thickness or displacements less than 1 mm.

granites. The data demonstrate a positive correlation for all fault modes (Fig. 6a) and for all lithologies, with similar D:T ratios in rocks of very different rheologies (Otsuki, 1978; Robertson, 1983). For example D:T ratios for crystalline rocks fall within the range of values recorded for poorly lithified sediments (Fig. 6b). Since there are many factors which may affect D:T ratios, including those related to deformation history and conditions, it is perhaps not surprising that different fault types in different host rocks appear to have similar D:T scaling. Some apparently distinctive characteristics are however revealed by dividing the dataset into subsets. For example fault rock thicknesses derived from limestones are greater than for clastic sediments (not shown). Previous workers have conducted detailed studies on limited, more focused datasets, to distil the effects of individual controls on D:T ratios (Marrett and Allmendinger, 1990; Knott, 1994; Little, 1995; Childs et al., 2007), however given the large numbers of factors which can control fault rock thickness, and the relative paucity of data once they are subdivided according to each factor, we believe that it is probably premature to derive mechanistic conclusions from the global dataset collated here.

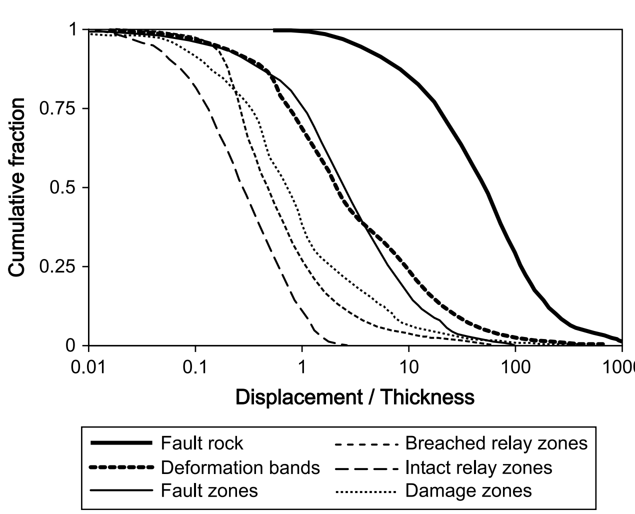


Fig. 5. Probability distributions of displacement/thickness ratios for different fault geometric components.

3.2. Fault zone

The fault zone thickness data display a broad correlation with fault displacement. The median D:T ratio is 2.5 (Figs. 4c and 5) with a spread in thickness measurements of up to 3.5 orders of magnitude at low displacements (<10 cm). Again there is a significant drop in the data spread at higher displacements ($\gtrsim$10 m). The reduction in the number of measured low D:T ratios (<1) (and a corresponding increase in the minimum D:T ratio) at higher displacement is expected, as the slip-surfaces of a wide, large displacement fault zone are less likely to be identifiable as part of the same zone at outcrop, than are the slip-surfaces of a thinner and

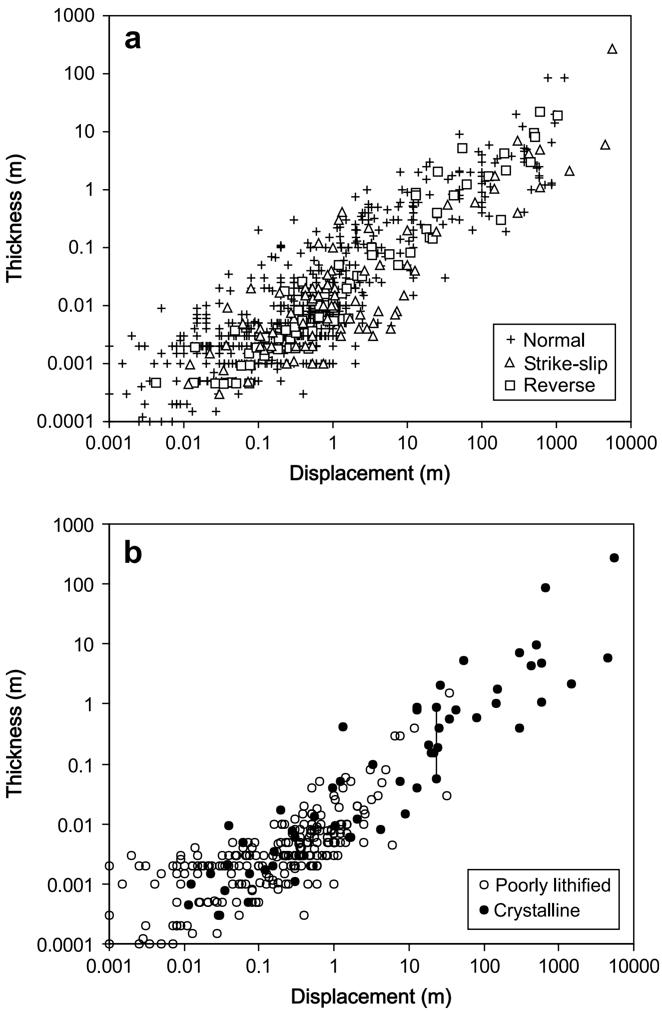


Fig. 6. Plots of fault rock thickness versus displacement for (a) the different modes of faulting and (b) different lithologies. The data in (b) are for normal faults offsetting poorly lithified clastic sequences (data from Otsuki, 1978; Childs et al., 2007), and for reverse and strike slip faults in crystalline rocks (data from Robertson, 1983; Hull, 1988; Blenkinsop, 1989).

lower displacement fault. The individual slip-surfaces of a large fault are therefore more likely to be recorded as several separate faults. In addition, the distinction between fault rock and lenses within a fault zone becomes more ambiguous on larger faults so that large displacement, narrow fault zones are more likely to be recorded as a single zone of breccia or gouge, as opposed to a fault zone.

3.3. Relay zones

Both the intact (Fig. 4d) and breached (Fig. 4e) relay zone data define positive correlations between displacement and thickness (i.e. separation), and have median D:T ratios of 0.27 and 0.46 respectively (Fig. 5). The higher ratio for breached relay zones reflects the tendency for relay zones to become breached with increasing shear strain. There is a significant overlap in D:T ratios for intact and breached relay zones as would be expected for a dataset collated from a wide range of rock types and sequences and from disparate fault systems. The degree of overlap between breached and intact relays can be greatly reduced if the data are derived from individual areas (e.g. Soliva and Benedicto, 2004), but is unlikely to be removed completely because faulted sequences are inherently heterogeneous.

The lower limit of the relay zone data occurs at D:T ratios ~0.01 (Figs. 4e and 5), because it becomes impossible to demonstrate displacement transfer between adjacent fault segments at lower ratios. For example, it would be difficult to demonstrate the presence of a relay zone between two faults of 1 m displacement with 100 m separation, however if the faults grow to a displacement of 10 m then the presence of a relay zone and related structures (e.g., a relay ramp) is more easily recognised.

3.4. Damage zone

Unlike the other categories, the compiled damage zone data do not define a linear correlation between thickness and displacement (Fig. 4f). The majority of the data lie within a trend which shows a relatively subdued increase in thickness (~2 orders of magnitude) over the ~ 4.5 orders of magnitude range of sampled displacements. The dataset as a whole suggests that damage zone thickness is established at low displacements but does not grow in direct proportion to the fault displacement. Such a growth trend might be expected, as the majority of the data are derived from normal faults in porous sandstones and, as for deformation bands in these rocks, damage zone growth may be retarded or stopped once a through-going slip-surface is developed. The data of Beach et al. (1999) define a slope of ~0.5 (Fig. 4f) and are consistent with this interpretation. However, the data from Knott et al. (1996) define a linear correlation (slope of 1), whilst the data of Shipton and Cowie (2001), define a linear relationship between displacement and damage zone thickness only for fault displacements greater than ca. 2 m (Fig. 4f). The reasons for the differences in the trends defined by the different datasets are unclear. Because damage zones in porous sandstones are thought to be bypassed when a through-going slip surface forms at a particular displacement, the correlation between damage zone thickness will break down above that displacement. Therefore, unlike the other fault components, damage zones cannot accommodate all, or the majority, of a fault displacement over the full scale range. For this reason we do not discuss further damage zones in the model for the evolution of fault zone structure, although we do acknowledge that zones of minor faults and fractures may be important fault components, but not necessarily in terms of the shear strain they accommodate.

4. Model for fault zone growth

The various fault components plotted in Fig. 4, with the exception of the fault damage zone data, each display a broadly linear correlation between fault displacement and thickness, or separation in the case of relay zones. While there is significant overlap in the distributions of each component, there is also a clear difference in their median D:T ratios (Figs. 5 and 7). As the D:T ratio is a proxy for average shear strain intensity, each of the various fault components (i.e., fault rock, fault zone and relay zone) can be regarded as representing structures which occur within a shear strain range. It is apparent, both from the difficulty in defining the differences between the various fault components, and the extent of data overlap between them, that there is a continuous spectrum of strain intensities within which the various categories of measurement are defined; difficulties of defining precise cut-offs between categories reflects both the variety of controlling factors and the sampling issues discussed above. The continuous spectrum of strain intensities reflects the progressive nature of related deformation in which an individual relay zone evolves through breaching and fault zone stages, and with sufficient displacement may ultimately yield a fault rock thickness which approximates the initial relay zone separation. A conceptual model of fault evolution based on these considerations is illustrated in Fig. 8. At the outset (very low displacements at Time

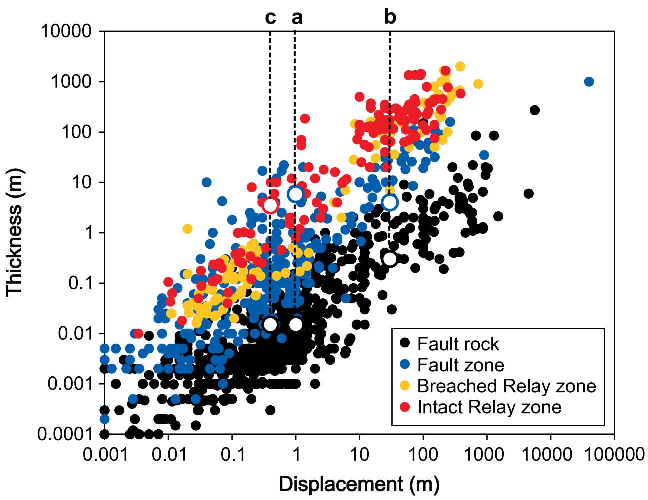


Fig. 7. Thickness versus displacement plot for the four fault geometric components used in defining the fault evolution model. The large circles show the measurements made on the three faults shown in Fig. 3a–c.

1) the fault comprises a series of segments each with surface irregularities on a range of scales: the latter will include asperities arising from refraction through heterogeneous sequences and rock volumes. We assume faults to be initially segmented on a wide range of scales but the kinematic coherence between fault segments at a given displacement will only be apparent over a limited scale which will increase with increasing displacement. For example at Time 1 only the structure at E is easily identified as a relay zone while the fault separations for other structures (e.g. A and H) may be too large to allow them to be identified as components of a single fault. Similarly the wavelength of fault surface irregularities may be too large in relation to fault offset and length to constitute a fault asperity; for example, a 100 m wavelength, 1 m amplitude irregularity is unlikely to be the focus for high strain on a fault with a displacement of 10 cm. As displacement increases all structures experience an increase in average shear strain and develop along growth lines which parallel the displacement axis of the idealised displacement versus thickness plot in Fig. 8b. Progressively larger relay zones are breached and fault surface asperities bypassed, each structure becoming incorporated into the fault zone and ultimately converted to fault rock which may be entrained along the fault in the displacement direction. In this scheme the thickness of fault rock which will form at a point on the fault surface is strongly influenced from the outset by the scale of fault segmentation and the amplitude of fault surface irregularities.

While the separations of individual relay zones are fixed from the outset, thereby placing an upper limit on fault rock thickness derived from their breaching and comminution, segmentation can occur on a wide range of scales so that several fault widening events caused by relay zone destruction may occur at the same point on a fault (e.g. structures C and H). This point is illustrated in Fig. 9 which shows an initial fault array comprising four overlapping fault segments with variable fault separations. As fault displacement increases, so does the fault rock thickness measured across the zone as successively larger zones of overlap are converted to fault rock. This figure is highly stylised, ignoring fault rock thicknesses other than those formed in the zones of overlap and assuming that each relay zone will be incorporated into the fault to become fault rock, but serves to illustrate how the thickness measured on a fault can increase progressively although the individual fault geometric components, as shown in Fig. 8, do not widen with increasing displacement. Unlike relay zone separations, asperity amplitudes are not defined solely at the propagation stage but asperities may also form as a result of deformation of a fault surface (e.g. by bedding parallel slip in the wall rock, Watterson et al., 1998) and by fault segment linkages. For example, breaching of D at Time 3 (Fig. 8) results in a large asperity (I) in the new through-going fault which is bypassed at Time 4. Increased displacement beyond Time 4 could lead to preferred displacement of the later footwall splay, with associated fault rock developed along this surface and possibly associated with deformation of the hanging wall fault-bounded blocks. Alternatively, because of changes in the footwall geometry it is just as likely that either a new slip surface will be generated or that any one of the hanging wall splays again becomes the principal slip surface, with the footwall surface, associated fault rock and immediate hanging wall, becoming the new footwall. Repeated movement on slip surfaces and generation of new slip-surfaces means that the fault zone structure and fault rock content can become very complex despite the operation of a single relatively simple process, i.e. asperity removal upon an already irregular and segmented surface.

Generation of new slip surfaces due to bypassing of fault surface asperities and movement on new or existing internal slip surfaces provide a means of locally increasing or decreasing fault rock, or fault zone, thicknesses. Depending on the orientation of the new slip surface, it may completely excise or duplicate the layer of fault rock so that the preserved fault rock at that point records only the

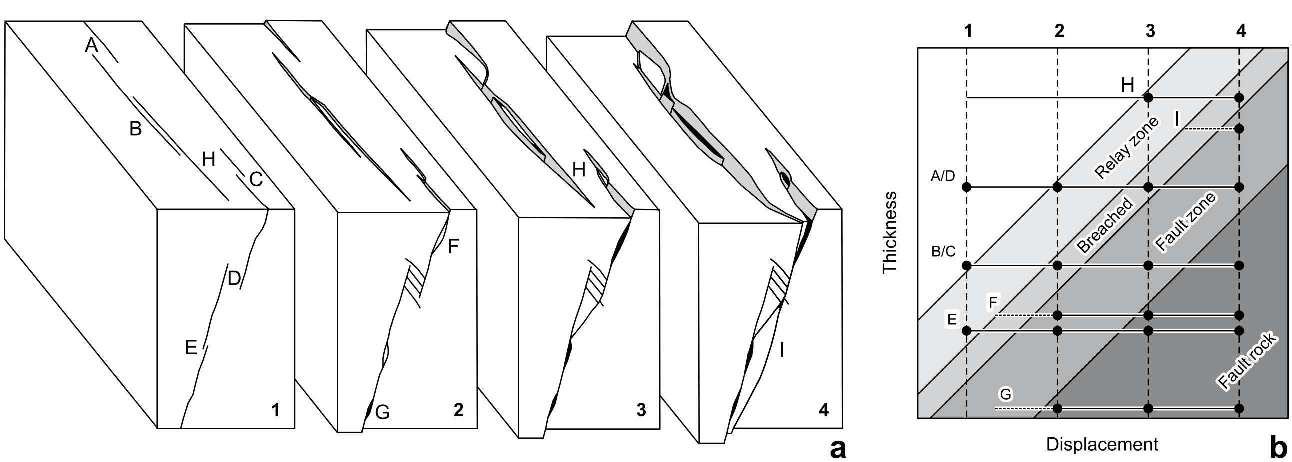


Fig. 8. (a) Block diagram illustrating the evolution of a fault array with increasing displacement at times 1 to 4. Individual fault irregularities, either asperities or segment boundaries are labelled A to I. The change in structure with fault displacement (and shear strain) at these irregularities is tracked on the displacement-thickness plot in (b). Vertical dashed lines in (b) are drawn at displacements at times 1 to 4 and the horizontal lines are growth curves for the individual fault geometric irregularities. Horizontal dashed lines extend to the left of asperities incorporated into the fault. The areas filled in black in (a) indicate fault rock.

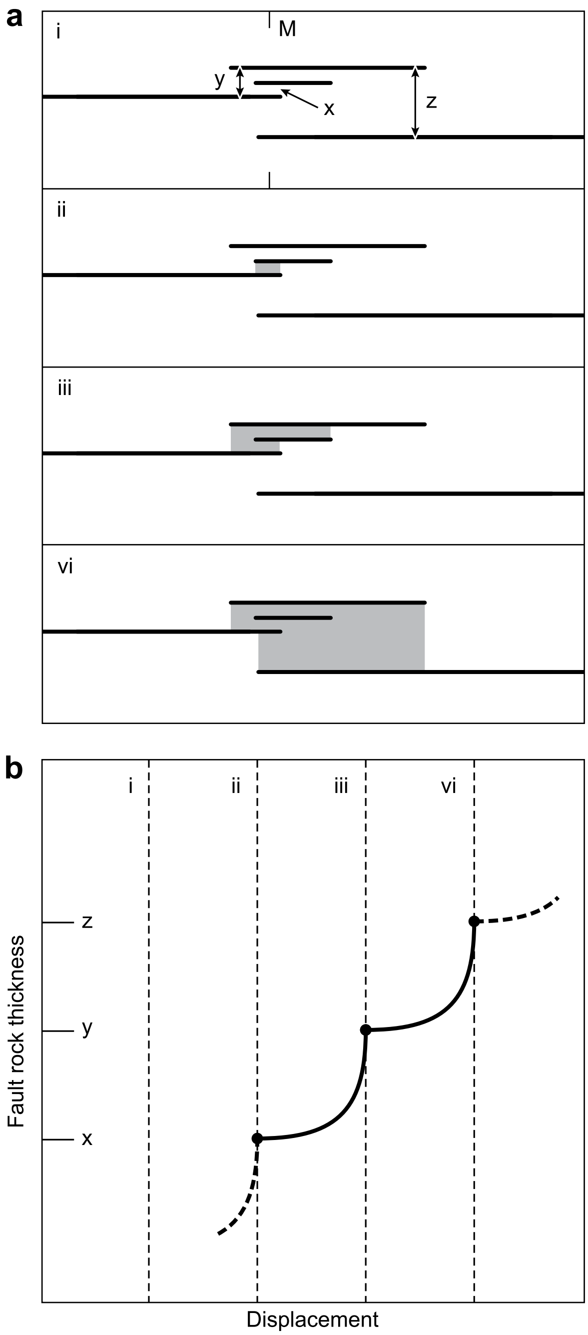


Fig. 9. (a) Schematic illustration of the increase in fault rock thickness at a point on a fault due to the conversion of rock volumes within fault relay zones into fault rock. Three relay zones with separations *x*, *y* and *z* are converted to fault rock (shaded) with increasing displacement (i to iv). (b) Log-log plot lot of fault rock thickness versus displacement showing the change in fault rock thickness measured along a section through M in (a) at displacements i through to iv. This figure is simplified in several ways (see text) for example, the fault rock thickness derived from a relay zone is shown to be equivalent to the initial relay zone separation.

displacement on the new slip surface or twice the original fault rock thickness; in that respect, the preserved fault rock and associated slip surfaces may represent an aggregate displacement which is quite different from the displacement across the zone. Detailed analysis of fault rock content in otherwise relatively simple normal faults reveals a level of complexity which is best explained by the repetitive nature of slip surface generation and displacement (Foxford et al., 1998; Berg and Skar, 2005).

Figure 8b shows separate ranges for each of the fault components although the different ranges have significant overlap (Figs 5 and 7) and the passage of a point on a fault from, for example, the fault zone to fault rock field represents a gradual rather than abrupt increase in the proportion of the incorporated rock volume converted into fault rock. In accordance with field observation (e.g. Bonson et al., 2007) Fig. 8a indicates the preferred site for fault rock generation to be fault branchlines, whether at linkages between segments (A at Time 4) or at the limits of fault-bounded lenses (B at Time 3).

Fig. 8 illustrates fault evolution at a single scale, although the processes illustrated will occur simultaneously over a wide range of scales. At Time 1 the structures labelled in Fig. 8a have a limited range of D:T ratios, the magnitude of D:T ratios associated with each structure increasing as the fault grows. However, at scales of observation smaller than shown in Fig. 8, higher D:T ratios would be encountered so that fault rock would be seen on the fault segments at Time 1. The structure shown at Time 4 and the associated D:T ratios could therefore represent a portion of one of the fault segments at Time 1. Therefore at any particular time the range of D:T ratios associated with a fault may range from infinite (at a discrete slip-surface) to a lower limit determined by the largest scale of irregularity/segmentation associated with the fault. The range of D:T ratios and fault geometric components encountered is therefore a function of both the displacement and the scale of observation. Differences in fault geometry and mathematical schemes for describing faults at different scales of observation have been reviewed by Ben-Zion and Sammis (2004).

Fig. 3 shows examples of cross-sections through three normal faults offsetting a poorly lithified sequence of sandstones and siltstones which serve to illustrate the model. Two thickness measurements are made for each structure, a fault rock measurement and a fault zone or relay zone measurement and these are plotted on Fig. 7. The structures of the faults in Fig. 3a and c are very similar, with two sub-parallel slip surfaces bounding a zone of low strain accommodated by movement on a small number of antithetic and, in the case of Fig. 3a, synthetic faults. In Fig. 3c, the upward termination of the hanging wall bounding fault and the downward termination of the footwall fault demonstrate that displacement is transferred between them by an anticlockwise rotation of the relay zone which is accommodated by movement on antithetic faults; this structure is identified as an intact relay zone on Fig. 7. The bounding faults of the structures in Fig. 3a do not tip out within the outcrop and the appropriate classification is therefore as a fault zone. The fault in Fig. 3b has the same overall structure and thickness as those in Fig. 3a and c, but has a displacement of some 10s of metres. The higher shear strain across this structure (a factor of ~50 larger) has resulted in a zone of intense minor faulting, along antithetic and synthetic faults. The fault rock thickness is ~20 times higher than that on the lower displacement structures due to both an increase in the number of internal minor faults and thickening of fault rocks associated with the two bounding slip surfaces. The structure in Fig. 3c is a contractional relay zone while the remaining relay zone data presented, as described above, are for neutral relay zones i.e. map view relay zones on normal faults. However the suite of structures in Fig. 3 serve to illustrate the fundamental feature of the proposed model, namely the evolution of fault structure by intensification of strain within a zone whose thickness is established at very low displacement.

The model of fault evolution described is one of progressive strain localisation resulting ultimately in a continuous fault containing variable fault rock thickness. The heterogeneity of fault rock thickness variations and fault rock type reflect the fact that both fault segmentation and fault surface asperities occur on a wide range of scales. A model of fault architectural evolution based on segmentation and the repetitive nature of asperity removal provides a basis for explaining not only the continued widening of fault rock with increased displacement but also the complexity of fault zone content.

5. Discussion

5.1. Fault architecture

In recent years it has become common to describe fault internal structure in terms of a high strain fault core surrounded by a damage zone, in which the fault core is the part of a fault where most of the displacement is accommodated and comprises fault gouge, breccia or cataclasite (Caine et al., 1996; Rawling et al., 2001; Billi et al., 2003). While there are obvious parallels between the core/damage zone model and the model described in this paper, with our fault rock being synonymous with fault core, and our other components (fault zones, breached and intact relay zones and damage zones) all included in the damage zone component, the differences between the two models are not simply semantic since they reflect differences in implied strain distribution. In the fault core/damage zone description, shear strain is a maximum at the core of the fault and decreases to a background strain at the damage zone margin. In the model described here, strain is distributed heterogeneously within a fault zone and strain distributions on transects across a fault vary with location on the fault. Although certain transects across a fault may conform to the core/damage description, highest strains (and regions of thickest fault rock) are generally at the margin of the fault zone with relatively little or no strain in the wall rock. These two conceptual models will tend to promote different models for the evolution of fault internal structure. An increase in core/damage zone thickness with displacement (Knott et al., 1996; Beach et al., 1999; Shipton and Cowie, 2001), together with the strain distribution implied in the core/damage zone description, will lead to a model of fault evolution in which the zone of deformation progressively widens through time (Shipton and Cowie, 2003). Our model envisages a more complex and often punctuated growth scheme, in which fault rock thicknesses do not display a monotonic increase but may even show decreases, arising from the repeated activity of cross-cutting internal slip surfaces.

The term fault damage zone incorporates many different types of structures including horse-tails at fault tips, relay ramps and normal drag: the variety of different structures which can be classified as damage zone components has been described by Kim et al. (2004). The scaling properties of these different structures will not be the same. In our opinion, improved understanding of fault evolution can be achieved by studying separately the scaling properties of each component rather than by grouping all components into a single measure. Similarly, the different structures are associated with different three-dimensional arrangements of fault rock (which may have permeabilities either higher or lower than both the wall rocks they separate and their parent wall rocks) and of juxtaposed wall rocks. These arrangements control flow across and within faults in either crystalline or clastic reservoirs or aquifers. Therefore, as we have demonstrated elsewhere (Manzocchi et al., 2008), we consider that outcrop studies and conceptual models aiming to characterise the permeability structure of faults are more useful if the older, more specific, terms are used rather than the generic terms (fault core, damage zone) which have become prevalent in the last decade.

5.2. Fault rock thickness

The model of fault evolution described here incorporates the generation of fault internal complexity both by segmentation leading to relay zone failure and by asperity removal. Asperity models of fault rock widening have been presented previously (Scholz, 1987; Power et al., 1988 and others). The model described here is slightly different to previous asperity models in that the emphasis is placed on the shearing-off of asperities as blocks of fault-bounded rock which are subsequently comminuted to fault rock and entrained along the fault, rather than the progressive incorporation of fault rock by continuous migration of the wall rock/fault rock interface and associated destruction of asperities. While the end result in terms of volume of fault rock generated may ultimately be the same, the inclusion of large wall rock blocks within fault zones is more consistent with field observations (e.g. Faulkner et al., 2003; Bonson et al., 2007).

Power et al. (1988) present a wear model based on observations of power-law scaling of fracture/fault surface roughness over scales from 5 μm to 40 m. In their model, fault rock volume generated by slip across the fracture is calculated as the volume of interpenetration of the two initially mated surfaces. This model does not incorporate the shearing off of discrete asperities. Definition of a model of fault development from an initial power-law fracture surface, which does incorporate shearing of asperities as lenses of relatively low strain rock, requires numerous additional assumptions to define what constitutes a discrete asperity at different scales and fault displacements. We have not attempted to devise such a model here as these assumptions cannot, as yet, be supported by field evidence. The measurements of Power et al. (1988) cannot therefore be easily converted to amplitude:displacement data and plotted on Fig. 7 and, to our knowledge, there are no published data on asperity amplitudes at different displacements. Furthermore, there is no objective means of determining whether a fault-bounded lens was formed by asperity removal or fault segmentation. Therefore, while the model we describe acknowledges the importance of fault surface irregularity, there is at present no means of estimating the relative fault rock volumes formed due to segmentation and asperity removal.

Strain hardening has previously been invoked as a possible explanation for the increase in fault rock thickness with increased displacement (Hull, 1988; Faulkner et al., 2003). However, positive correlations appear to be observed for all faulted lithologies (e.g. Fig. 6b), even though it is relatively rare for fault rocks to be stronger than their parent wall rocks (one exception is deformation bands in porous sandstones; Aydin and Johnson, 1983). Strain hardening does not therefore provide a general explanation for fault rock widening. Instead we present a geometric, as opposed to rheologic, model, in which fault rock weakening and progressive strain concentration accompany the expansion of fault zones and the accretion of fault rock. The fact that the D:T ratios for fault rocks derived from widely different lithologies, e.g. poorly lithified sediments and crystalline rocks (Fig. 6b), are relatively similar, even though the deformation mechanisms and associated fault rock products may be very different, suggests that neither lithology nor deformation mechanism have, on their own, a pre-eminent control. Instead, we suggest that geometric effects arising from fault surface irregularities and segmentation may be the main factor. Nevertheless, because initial fault geometry is likely to be controlled by the mechanical heterogeneity of the wall rock (e.g. bedding; Peacock and Sanderson, 1992; Childs et al., 1996a; Wilkins and Gross, 2002; Ferrill and Morris, 2003; Schöpfer et al., 2006), it is likely that rheology is, at least indirectly, an important factor.

6. Summary and conclusions

A model for the evolution of fault internal structure has been presented. In its simplest form, the model is one of progressive strain concentration in a zone, within which the active fault surface progressively approaches, albeit along a potentially complex path, a more planar geometry. The width of this zone is influenced strongly by the scale of initial fault segmentation and fault surface irregularity, but increasing strain intensity with increasing displacement results in an apparent widening of the various fault components. The by-passing of fault surface irregularities and the linkage of fault segments on a broad range of scales and displacements, provides for a model that relates different fault structures to differing degrees of strain intensity, as measured by increasing displacement to thickness ratios. This model provides a basis for the well known positive correlation between fault rock thickness and fault displacement. The principal features of this model are:

- Irregular fault surfaces and fault segmentation arising from the propagation of faults through rock volumes result in fault surface complexities which are bypassed during subsequent increases in displacement.
- The scale and distribution of initial fault surface complexities provide a critical control on the thickness and distribution of fault rock.
- Displacement accumulation causes the removal of asperities and the breaching and deformation of relay zones as a fault becomes progressively more planar with time.
- Various fault components, such as relay zones, fault zones and fault rock, can be assigned to overlapping shear strain ranges which together define a continuous spectrum of strain.
- Fault zone complexity and ultimately fault rock thickness are more directly controlled by fault geometry and segmentation than by the local wall rock and fault rock rheologies. The length-scales and orientations of rheological heterogeneity (including the layering of the wall rock sequence) may, however, influence strongly the initial fault surface geometry and segmentation.
- This model of fault rock development is consistent with the wide range of fault rock thicknesses on individual faults and with the lack of a clear lithological dependence, because it acknowledges the complex nature of fault surface irregularities and segmentation, combined with the temporal and spatial variations in fault displacements.

Acknowledgements

We thank former members of the Fault Analysis Group, in particular Juan Watterson and Julian Strand, for discussions on fault zone structure. Alan Morris and David Peacock are thanked for their helpful reviews of the manuscript. We thank the following companies and their representatives for supporting our work on faults in Taranaki, New Zealand, carried out as apart of the ITF brokered FIFT research project: Amerada Hess, BG Energy Holdings, BP Exploration, ConocoPhillips, Kerr-McGee North Sea, Shell, Statoil and Total.

References

Ameen, M.S., 1995. Fracture characterisation in the Chalk and the evolution of the Thanet monocline, Kent, southern England. In: Ameen, M.S. (Ed.), Fractography: Fracture Topography as a Tool in Fracture Mechanics and Stress Analysis. Special Publication, 92. Geological Society, pp. 149–174.
Antonellini, M., Aydin, A., 1995. Effect of faulting on fluid flow in porous sandstones: geometry and spatial distribution. American Association of Petroleum Geologists Bulletin 79, 642–671.
Aydin, A., 1988. Discontinuities along Thrust Faults and the Cleavage Duplexes. Geological Society of America, Special Paper 222, pp. 223-233.
Aydin, A., Johnson, A.M., 1983. Analysis of faulting in porous sandstones. Journal of Structural Geology 5, 19–31.
Beach, A., Welbon, A.I., Brockbank, P.J., McCallum, J.E., 1999. Reservoir damage around faults: outcrop examples from the Suez rift. Petroleum Geoscience 5, 109–116.
Ben-Zion, Y., Sammis, C.G., 2004. Characterization of fault zones. Pure and Applied Geophysics 160, 677–715.
Berg, S.S., Skar, T., 2005. Controls on damage zone asymmetry of a normal fault zone: outcrop analyses of a segment of the Moab fault, SE Utah. Journal of Structural Geology 27, 1803–1822.
Billi, A., Salvini, F., Storti, F., 2003. The damage zone-fault core transition in carbonate rocks: implications for fault growth, structure and permeability. Journal of Structural Geology 25, 1779–1794.
Blenkinsop, T.G., 1989. Thickness-displacement relationships for deformation zones: discussion. Journal of Structural Geology 11, 1051–1054.
Bonson, C.G., Childs, C., Walsh, J.J., Schöpfer, M.P.J., Carboni, V., 2007. Geometric and kinematic controls on the internal structure of a large normal fault in massive limestones: the Maghlaq Fault, Malta. Journal of Structural Geology 29, 336–354.
Caine, J.S., Evans, J.P., Forster, C.B., 1996. Fault zone architecture and permeability structure. Geology 24, 1025–1028.
Chadwick, R.A., 1986. Extension tectonics in the Wessex Basin, southern England. Journal of the Geological Society 143, 465–488.
Childs, C., 2000. The structure and hydraulic properties of fault zones. Unpublished PhD thesis, University of Liverpool.
Childs, C., Watterson, J., Walsh, J.J., 1995. Fault overlap zones within developing normal fault systems. Journal of the Geological Society 152, 535–549.
Childs, C., Nicol, A., Walsh, J.J., Watterson, J., 1996a. Growth of vertically segmented normal faults Journal of Structural Geology 18, 1389–1397.
Childs, C., Watterson, J., Walsh, J.J., 1996b. A model for the structure and development of fault zones. Journal of the Geological Society 153, 337–340.
Childs, C., Walsh, J.J., Watterson, J., 1997. Complexity in fault zone structure and implications for fault seal prediction. In: Møller-Pedersen, P., Koestler, A.G. (Eds.), Hydrocarbon Seals – Importance for Exploration and Production. Special Publication, 7. Norwegian Petroleum Society (NPF), pp. 61–72.
Childs, C., Walsh, J.J., Manzocchi, T., Strand, J., Nicol, A., Tomasso, M., Schöpfer, M.P.J., Aplin, A., 2007. Definition of a fault permeability predictor from outcrop studies of a faulted turbidite sequence, Taranaki, New Zealand. In: Jolley (Ed.), Structurally Complex Reservoirs. Special Publication, 292. Geological Society, London, pp. 235–258.
Cox, S.D.J., Scholz, C.H., 1988. On the formation and growth of faults: an experimental study. Journal of Structural Geology 10, 413–430.
Dahlstrom, C.D.A., 1969. Balanced cross sections. Canadian Journal of Earth Sciences 6, 743–757.
Evans, J.P., 1990. Thickness-displacement relationships for fault zones. Journal of Structural Geology 12, 1061–1065.
Faulkner, D.R., Rutter, E.H., 2001. Can the maintenance of overpressured fluids in large strike-slip fault zones explain their apparent weakness? Geology 29, 503–506.
Faulkner, D.R., Lewis, A.C., Rutter, E.H., 2003. On the internal architecture and mechanics of large strike-slip fault zones: field observations of the Carboneras fault in southeastern Spain. Tectonophysics 367, 235–251.
Ferrill, D.A., Morris, A.P., 2001. Displacement gradient and deformation in normal fault systems. Journal of Structural Geology 23, 619–638.
Ferrill, D.A., Morris, A.P., 2003. Dilational normal faults. Journal of Structural Geology 25, 183–196.
Ferrill, D.A., Stamatakos, J.A., Sims, D., 1999. Normal fault corrugation: implications for growth and seismicity of active normal faults. Journal of Structural Geology 21, 1027–1038.
Ferrill, D.A., Morris, A.P., Sims, D.W., Stamatakos, J.A., 2001. The geometric strength of fault systems. http://aapg.confex.com/aapg/de2001/techprogram/paper_8288.htm.
Fossen, H., Hesthammer, J., 2000. Possible absence of small faults in the Gullfaks Field, northern North Sea: implications for downscaling of faults in some porous sandstones. Journal of Structural Geology 22, 851–863.
Foxford, K.A., Walsh, J.J., Watterson, J., Garden, I.R., Guscott, S.C., Burley, S.D., 1998. Structure and content of the Moab Fault Zone, Utah, USA, and its implications for fault seal prediction. In: Jones, G., Fisher, Q.J., Knipe, R.J. (Eds.), Faulting, Fault Sealing and Fluid Flow in Hydrocarbon Reservoirs. Special Publication, 147. Geological Society, pp. 87–103.
Huggins, P., Watterson, J., Walsh, J.J., Childs, C., 1995. Relay zone geometry and displacement transfer between normal faults recorded in coal-mine plans. Journal of Structural Geology 17, 1741–1755.
Hull, J., 1988. Thickness-displacement relationships for deformation zones. Journal of Structural Geology 10, 431–435.
Hundley-Goff, E.M., Moody, J.B., 1980. Microscopic characteristics of orthoquartzite from sliding friction experiments, II. Gouge. Tectonophysics 62, 301–319.
Imber, J., Tuckwell, G.W., Childs, C., Walsh, J.J., Manzocch, T., Heath, A.E., Bonson, C.G., Strand, J., 2004. Three-dimensional distinct element method simulations of relay growth and breaching along normal faults. Journal of Structural Geology 26, 1897–1911.
Kattenhorn, S.A., Pollard, D.D., 2001. Integrating 3-D seismic data, field analogs, and mechanical models in the analysis of segmented normal faults in the Wytch Farm oil field, southern England, United Kingdom. American Association of Petroleum Geologists Bulletin 85, 1183–1210.
Jamison, W.R., Stearns, D.W., 1982. Tectonic deformation of Wingate sandstone, Colorado National Monument. American Association of Petroleum Geologists Bulletin 62, 2584–2608.
Kim, Y.S., Peacock, D.C.P., Sanderson, D.J., 2004. Fault damage zones. Journal of Structural Geology 26, 503–517.
Knott, S.D., 1994. Fault zone thickness versus displacement in the Permo-Triassic sandstones of NW England. Journal of the Geological Society 151, 17–25.
Knott, S.D., Beach, A., Brockbank, P.J., Lawson Brown, J., McCallum, J.E., Weldon, A.I., 1996. Spatial and mechanical controls on normal fault populations. Journal of Structural Geology 18, 359–372.
Kristensen, M.B., Childs, C.J., Korstgård, J.A., 2008. The 3D geometry of small-scale relay zones between normal faults in soft sediments. Journal of Structural Geology 30, 257–272.

Larsen, P.-H., 1988. Relay structures in a Lower Permian basement-involved extension system, East Greenland. Journal of Structural Geology 10, 3–8.

Little, T.A., 1995. Brittle deformation adjacent to the Awatere strike-slip fault in New Zealand: faulting patterns, scaling relationships, and displacement partitioning. Bulletin of the Geological Society of America 107 11, 1255–1271.

Mair, K., Main, I., Elphick, S., 2000. Sequential growth of deformation bands in the laboratory. Journal of Structural Geology 22, 25–42.

Mandl, G., 1987. Discontinuous fault zones. Journal of Structural Geology 9, 105–110.

Manzocchi, T., 1997. Quantification of flow impairment in faulted sandstone reservoirs. Unpublished PhD thesis, Heriot-Watt University.

Manzocchi, T., Heath, A.E., Palananthakumar, B., Childs, C., Walsh, J.J., 2008. Faults in conventional flow simulation models: a consideration of representational assumptions and geological uncertainties. Petroleum Geoscience 14, 91–110.

Marchal, D., Guiraud, M., Rives, T., 2003. Geometric and morphological evolution of normal fault planes and traces from 2D to 4D data. Journal of Structural Geology 2, 135–158.

Marrett, R., Allmendinger, R.W., 1990. Kinematic analysis of fault-slip data. Journal of Structural Geology 12, 973–986.

Morley, C.K., Nelson, R.A., Patton, T.L., Munn, S.G., 1990. Transfer zones in the East African Rift System and their relevance to hydrocarbon exploration in rifts. American Association of Petroleum Geologist Bulletin 74, 1234–1253.

Naylor, M.A., Mandl, G., Sijpesteijn, C.H.K., 1986. Fault geometries in basement-induced wrench faulting under different initial stress states. Journal of Structural Geology 8, 737–752.

Otsuki, K., 1978. On the relationship between the width of shear zone and the displacement along fault. Journal of the Geological Society of Japan 84, 661–669.

Peacock, D.C.P., Sanderson, D.J., 1991. Displacements, segment linkage and relay ramps in normal fault zones. Journal of Structural Geology 13, 721–733.

Peacock, D.C.P., Sanderson, D.J., 1992. Effects of layering and anisotropy on fault geometry. Journal of the Geological Society 149, 793–802.

Peacock, D.C.P., Sanderson, D.J., 1994. Geometry and development of relay ramps in normal fault systems. American Association of Petroleum Geologists Bulletin 78, 147–165.

Peacock, D.C.P., Knipe, R.J., Sanderson, D.J., 2000. Glossary of normal faults. Journal of Structural Geology 22, 291–305.

Power, W.L., Tullis, T.E., Brown, S.R., Boitnott, G.N., Scholz, C.H., 1987. Roughness of natural fault surfaces. Geophysical Research Letters 14, 29–32.

Power, W.L., Tullis, T.E., Weeks, J.D., 1988. Roughness and wear during brittle faulting. Journal of Geophysical Research 93 (B12), 15268–15278.

Rawling, G.C., Goodwin, L.B., Wilson, J.L., 2001. Internal architecture, permeability structure, and hydrologic significance of contrasting fault-zone types. Geology 29, 43–46.

Robertson, E.C., 1982. Continuous formation of gouge and breccia during fault displacement. In: Goodman, R.E., Heuze, F.E. (Eds.), Issues in Rock Mechanics, Proceedings of the 23rd Symposium on Rock Mechanics. American Institute of Mining Engineers, pp. 379–404.

Robertson, E.C., 1983. Relationship of fault displacement to gouge and breccia thickness. American Institute of Mining Engineers Transactions 35, 1426–1432.

Sagy, A., Brodsky, E.E., Axen, G.J., 2007. Evolution of fault-surface roughness with slip. Geology 35, 283–286.

Scholz, C., 1987. Wear and gouge formation in brittle faulting. Geology 15, 493–495.

Schöpfer, M.P.J., Childs, C., Walsh, J.J., 2006. Localisation of normal faults in multilayer sequences. Journal of Structural Geology 28, 816–833.

Shipton, Z.K., Cowie, P.A., 2001. Damage zone slip-surface evolution over μm to km scales in high-porosity Navajo sandstone, Utah. Journal of Structural Geology 23, 1825–1844.

Shipton, Z.K., Cowie, P.A., 2003. A conceptual model for the origin of fault damage zone structures in high-porosity sandstone. Journal of Structural Geology 25, 333–344.

Soliva, R., Benedicto, A., 2004. A linkage criterion for segmented normal faults. Journal of Structural Geology 26, 2251–2267.

Tchalenko, J.S., 1970. Similarities between shear zones of different magnitudes. Bulletin of the Geological Society of America 81, 1625–1640.

van der Zee, W., 2002. Dynamics of fault gouge development in layered sand-clay sequences. PhD thesis, RWTH Aachen. Shaker Verlag, ISBN3-8322-0585-3, 155 pp.

van der Zee, W., Urai, J.L., 2005. Processes of normal fault evolution in a siliciclastic sequence: a case study from Miri, Sarawak, Malaysia. Journal of Structural Geology 27, 2281–2300.

Wallace, R.E., Morris, H.T., 1979. Characteristics of faults and shear zones as seen in mines at depths as much as 2.5 km below the surface. United States Geological Survey Open File Report 79-1239, 79–100.

Walsh, J.J., Watterson, J., 1991. Geometric and kinematic coherence and scale effects in normal fault systems. In: Roberts, A.M., Yielding, G., Freeman, B. (Eds.), The Geometry of Normal Faults. Special Publication, 56. Geological Society, London, pp. 193–203.

Walsh, J.J., Watterson, J., Bailey, W., Childs, C., 1999. Fault Relays, Bends and Branchlines. Journal of Structural Geology 21, 1019–1026.

Walsh, J.J., Childs, C., Meyer, V., Manzocchi, T., Imber, J., Nicol, A., Tuckwell, G., Bailey, W.R., Bonson, C.G., Watterson, J., Nell, P.A.R., Strand, J.A., 2001. Geometrical controls on the evolution of normal fault systems. In: Holdsworth, R.E., Strachan, R.A., Magloughlin, J.F., Knipe, R.J. (Eds.), The Nature of the Tectonic Significance of Fault Zone Weakening. Special Publication, 186. Geological Society, London, pp. 157–170.

Walsh, J.J., Bailey, W.R., Childs, C., Nicol, A., Bonson, C.G., 2003. Formation of segmented normal faults: a 3-D perspective. Journal of Structural Geology 25, 1251–1262.

Watterson, J., Childs, C., Walsh, J.J., 1998. Widening of fault zones by erosion of asperities formed by bed-parallel slip. Geology 26, 71–74.

Wilkins, S.J., Gross, M.R., 2002. Normal fault growth in layered rocks at Split Mountain, Utah: influence of mechanical stratigraphy on dip linkage, fault restriction and fault scaling. Journal of Structural Geology 24, 1413–1429.

Wolf, R., 1985. Tiefentektonik des linksneiderrheinischen Steinkohlengebietes. In: Drozdzewski, G., Engel, H., Wolf, R., Wrede, V. (Eds.), Beiträge zur Tiefentektonik Westdeutscher Steinkolhlenlagerstätten. Geologisches Landesamt Nordrhein-Westfalen, Krefeld.

Woodcock, N.H., Fischer, M., 1986. Strike-slip duplexes. Journal of Structural Geology 8, 725–735.

Worthington, R., 2006. Geometric and kinematic analysis of Cenozoic faulting onshore and offshore west of Ireland. Unpublished PhD thesis, University College Dublin.