The influence of layering and pre-existing joints on the development of internal structure in normal fault zones: the Lodève basin, France

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Abstract: This paper examines the role of mechanical stratigraphy on the evolution of normal fault geometry and fault zone internal structure, using a well-exposed normal fault system from the Permian Lodève Basin, southern France. Faults formed early during the syn-deformation tilting history of the basin tend to have steeper segments in the competent sandstone layers due to refraction, assisted by pre-existing early bedding-perpendicular joints, where displacement remained on the order of bed thickness. Faults which continued to slip during tilting have a more complex structure of splays due both to the space incompatibility problem of slip at fault bends of this irregular geometry, and because tilting favours the generation of new splays at a different angle to the earlier faults experiencing rotation. Continued deformation between faults and their splays often causes both distributed deformation in between the two, and reconnection of splays to the main fault forming isolated lenses. Thus, fault zone complexity increases greatly as slip exceeds competent bed thickness, owing both to the presence of the mechanical layering, and the fact that this layering is being tilted.

Faults in the upper crust evolve in relation to changes in the structure of the effective stress field and the mechanical properties of the rocks. For example, basin-wide tilting during ongoing deformation can rotate faults to an unfavourable orientation for further slip, and cause faults to splay or even initiate new faults (cf. Jaeger & Cook 1976; Sibson 1985; Buck 1993; Agnon & Reches 1995; Wibberley et al. 2007). Mechanical heterogeneity (e.g., sedimentary layering, pre-existing faults or joints) has a strong influence on fault propagation through the system (Peacock & Sanderson 1992; Bürgmann & Pollard 1994; Childs et al. 1996; Gross et al. 1997; Martel 1999; Wilkins & Gross 2002; Soliva & Benedicto 2005). The interplay of all these processes can be expected to produce a variety of fault geometric irregularities at all length scales. During slip on a non-planar fault additional deformation of the wall rock must occur. The deformation of asperities is thought to cause widening of the fault zone (such as those caused by the interaction with bedding-parallel slip planes (Watterson et al. 1998), leading to the suggestion that there is a universal relationship between displacement on a fault and fault zone thickness (e.g., Scholz 1987; Hull 1988). The likely operation of these processes implies that the internal structure will evolve to become generally more complex through time during deformation. However, the influence of such mechanical heterogeneity and the generation of asperities on fault zone complexity is poorly understood. Furthermore, the feasibility of directly comparing data sets from different fault systems formed in different geological and tectonic settings has been questioned (e.g., Blenkinsop 1989; Evans 1990).

In this paper, we use outcrop data of faults in a sand–shale sequence to better understand the evolution of normal faults in a layered medium. We investigate the influence of layering and pre-existing joints on the fault zone width, fault throw, and the internal structure of normal fault
zones. The faults studied are related to the Mas d’Alary–Saint Jean fault zone in the Permian Lodève Basin in the South of France, and crop out in the Tréviers open-cast mine from which the observations presented in this paper are taken. The outcrop is situated near the hamlet of Mas d’Alary, 3 km SSE of Lodève in the eastern part of the Lodève Basin, southern France (Fig. 1).

Geological setting

The 15 × 25 km Lodève basin is situated 50 km west of Montpellier, south of the Massif Central (Lopez 1992). It is a half-graben which formed during the Permian. The Palaeozoic strata dip 15–30° southward and consist of Permian sediments concordantly deposited on a Cambrian basement, in turn covered by horizontal Mesozoic sediments (Fig. 1). The Mesozoic cover is almost completely eroded, exposing both the Permian sediments and basement. The Permian sediments comprise a combination of continental, detrital and bioclastic sediments representing a lacustrine to floodplain succession, and related to a progressive change from a humid-tropical to semi-arid climate.

The rocks exposed in the Tréviers open-cast mine comprise a layered shale-sandstone sequence of 30–35% sandstone (low-porosity, calcite-cemented arkose) and 65–70% shale. The bed thickness of the sandstone varies from 10 cm to 1 m, whereas the bed thickness of the shale ranges from very thin laminae (5 mm) to thick beds (2 m). The shale is hard and weathers to a flaky aggregate, which can be related to overconsolidation during burial to approximately 5 km maximum depth (pers. comm., J. P. Petit). Most of the geologic boundaries of the Permian Lodève basin are formed by faults. During the early- to middle-Permian, the regional Cambrian structure was eroded and the valleys filled with conglomerates, causing the non-uniform thickness of the conglomerate sequence. At the end of this period the major basement faults were reactivated.

During the late Permian the basin was tilted 15–20° to the south by reactivation of older basement faults, interpreted by Lopez (1992) as a roll-over anticline involving the Permian half-graben and its Cambrian basement above a late orogenic listric fault, the Aires fault. Also during this tectonic event, the Mas d’Alary–Saint Jean fault zone formed, presumably in the collapsed crest of the roll-over anticline. Thus the faults examined in this paper are thought to be early Saxonian in age, and active before the main burial by the remaining Upper Permian sediments. This is compatible with a recent interpretation that these faults started forming at the onset of tilting and continued to evolve during tilting of the Permian strata as the listric Lodève Basin evolved (Wibberley et al. 2007). Although the exact depth at which the faults formed is not known, it is therefore likely to be less than 1000 m based on correlated thicknesses of the Upper Autunian and Lower Saxonian (syn-rift) strata.

The opening of Neotethys during the Jurassic time caused marine sedimentation and syn-sedimentary deformation. Inside the basin this period is only marked by rotated blocks in the Mesozoic cover. The Pyrenean shortening (c. 40 Ma) caused strong deformation east of the Cévennes fault (northwards thrusting and east–west trending folds), but only minor deformation in the Lodève basin. The deformation styles found in the Lodève basin are slight inversions of normal faults, strike-slip reactivation of the Permian fault and some bedding-parallel slip at the base of the sedimentary sequences. The large extensional deformation in the Oligocene (opening of the Camargue trough) only had an influence east of the Cévennes fault, and caused a reactivation of the Cévennes fault itself (normal movement). Therefore, although the tectonic history of the basin is rather complex, the structures found in the Mas d’Alary–Saint Jean fault zone investigated in this study are essentially extensional and of Permian age with a small overprint of the Pyrenean compression (manifested at the outcrop scale by slight reverse-sense bedding-parallel slip).

Field study of the Mas d’Alary–Saint Jean fault zone

The Tréviers pit-mine exposes a large variety of normal fault structures in lower Permian rocks. The outcrop face is approximately 140 m long and 25 m high. The bedding is tilted 30° southwards and crosscut by mostly north-dipping normal faults and some nearly vertical and south-dipping faults (Fig. 2).

Methods

The total outcrop face was photographed in detail with a 300 mm objective to prevent distortion as much as possible. The overview angle from north to south is 40° and from top to bottom 15° (photos are taken level with the half-height of the outcrop). The interesting structures in the lower part of the outcrop (up to c. 6 m) were examined in more detail from ground level. Care was taken to take photographs parallel to fault strike where possible, to avoid distortion. Structures were investigated with the help of detailed sketches and interpretation of photographs. Orientation data and
structural observations were collected along a scan-
line at the bottom part of the outcrop (ground level).

Samples were collected after on-site impreg-
nation with a very low viscosity epoxy, which was
poured on the outcrop before extracting the
samples. After extraction from the outcrop, the
sample was impregnated from all sides, and stabil-
ized with a gypsum collar. For microstructural
analysis, the sample was cut perpendicular to the fault plane, to allow high resolution photography.

**Characteristics of individual faults**

The strike of the faults is perpendicular to the outcrop face. Fault offset ranges from 10 cm to 25 m. The structure as seen in the outcrop is shown in the profile of Figure 2. Although the faulting took place during the Permian, no evidence for syn-sedimentary faulting of Permian beds is found. Significant features are the larger number of normal faults which dip 40–60° to the north and the few larger normal faults which down-throw to the south (very steep, sometimes overturned). The measured fault orientations (Fig. 3) are biased towards the steeper faults, because measurable fault surfaces were mostly found in the sandstone beds due to outcrop quality, and these faults tend to be steeper. In the northern part of the outcrop, bedding-parallel slip is evidenced by bedding-parallel offset of the steep normal faults.

Low displacement (d < 1.5 m) north-dipping normal faults are bedding-oblique and have dips of 20–50°. Their orientation with respect to bedding is between 50 and 80°. Although they are irregular in nature, the faults are consistently steeper in the sandstone beds than in the shale beds (e.g., Fault 2, Fig. 4a; Fault 12, Fig. 4j), a point also borne out by separating the orientation statistics of sandstone fault contacts from shale-shale juxtaposition contacts (Wibberley et al. 2007). In the sandstone beds the fault surfaces follow pre-existing joints, which are perpendicular to bedding. Where these faults coincide with joint surfaces, the faults are often splayed and the area between the two splays is filled with shale (e.g., Fig. 4a). Step-over geometries often form restraining bends which either partly or fully transfer the offset (e.g., Fig. 4c & d).

Low-displacement (d < 1.5 m) south-downthrowing faults are vertical or slightly overturned, at angles of 65–80° to bedding. They are irregular, following bedding-perpendicular joints in the sandstone beds and have splays and/or conjugates which form small graben structures in the hanging walls of the faults. Irregularity is also expressed by the generation of lens structures.
bounded by splays (Figs 4h & 5a) or at step-overs (Fig. 4b).

A single high-displacement north-downthrowing fault, Fault 8 (Fig. 4g), has a large offset (10 m). This fault splays in the middle of the outcrop, which is the top part of a large lens. The full lens geometry of this structure was sketched by Bruel (1997) before the open-cast mine was recultivated and partially filled (Fig. 2c). The overall dip of the fault above the lens is 45°. Detailed observation showed a relatively undisturbed wall rock, and a severely faulted and folded interior of the lens (Fig. 4g). There is a remarkable absence of deformation in the hanging wall of this fault.

High-displacement south-downthrowing faults, Faults 11 and 13, are vertical and have offsets of 25 and 5 m respectively. Both fault zones have splays branching off the main fault, but the higher-displacement fault, Fault 11, shows a much more complicated internal structure with splays re-connecting to the main fault and consequent lenses inside the fault zone in which rotated bedding is still recognizable. Both fault zones contain shale-rich gouge, but Fault 11 also has cemented breccia within the gouge.

Bedding-perpendicular faults (Faults 6 & 7, Fig. 2b) are north-downthrowing normal faults interpreted to have formed as the system rotates during regional tilting of the basin (Wibberley et al. 2007). They dip 60° to the north and have offsets of approximately 5 m and 2 m respectively. They both consist of two parallel strands separated by weakly deformed rock in which bedding is still visible (Figs 4e, f & 5b). In the case of Fault 7, the strands anastomose and widening of the fault zone by divergence of the splays is observed where bedding-parallel slip surfaces interact with the fault during movement (Fig. 5b); here, fault gouge thickness varies from 1 to 15 cm. The edges of the fault zone cut the sandstone bed in a clean break identical in appearance to joint surfaces, and indeed bedding-perpendicular joints in the vicinity often have small (millimetre to centimetre) amounts of slip on them (Fig. 6a), suggesting that these bedding-perpendicular faults propagated along joints in the sandstone beds. Similar reactivated bedding-parallel joints are present elsewhere in the outcrop and suggest that this is a general phenomenon (e.g., Fig. 6b). In the case of Fault 6, one of the strands terminates halfway up the outcrop (Fault 6a in Fig. 4f) in a zone of continuous deformation (folding) of the wall rock so that total displacement \( a + b \) stays constant (Fig. 4e). Anomalous bedding rotation to the north in the overlap zone between strands a and b, facilitated by bedding-parallel slip, attests to the transfer of displacement from one strand to the other (Fig. 4e). Striations are visible on the exposed fault surfaces, showing a purely down-dip movement (Wibberley et al. 2007). A very thin coating of clay is present on parts of the fault surfaces.

A system of bedding-parallel faults makes up a complex structure in the northern part of the outcrop adjacent to Fault 13 (Fig. 4j). The tilted bedding functions as a slip plane for low-angle ‘normal-sense’ south-dipping faults. These bedding-parallel slip planes are amongst the youngest faults in the entire outcrop, offsetting the older north-dipping faults in this region. The evolution of this network is described in more detail in Wibberley et al. (2007).


**Joints**

Two joint sets are recognized in the outcrop. The joints are typically only visible in the sandstone layers. The joint surfaces cross cut the sand layers at high angles to bedding. The strikes of the two joint sets are perpendicular and parallel to that of the faults. The fault-parallel joint set is often reactivated (Fig. 7), as mentioned above, particularly in the generation of bedding-perpendicular faults. The combination of the lower angle faults in the shale and the slip-reactivated joints in the sandstone bed give the faults an irregular geometry (e.g., Fig. 8).

**Fault zone internal structure**

The fault zones consist mostly of clay with quartz and/or sandstone blocks present in places, as may
Fig. 4. (Continued).
Fig. 5. (a) Lens structure in the lower part of fault zone 10. Measuring tape = 2 m. (b) Detail of Fault 7d. Note the strong variation in fault gouge thickness.
be expected from the composition of the wall rock (65–70% shale and 30–35% sandstone). In all cases of sand–sand juxtapositions (and throw > bed thickness), a continuous clay layer is found on the fault surfaces. The thickness of the shale-rich zones ranges from a thin coating of clay on the fault surfaces of the sandstone beds, to a zone of clay gouge or shale-rich deformation bands 10–20 cm thick in the larger faults and in oblique pull-aparts. Microstructural analyses show that the sandstone layers are deformed in a brittle manner whereas the shale was 'ductile', resulting in angular sandstone clasts floating in a clay matrix (Fig. 6), and this style appears to be consistent for earlier (rotated) and later faults. The sandstone layers are cross-cut by little faults which separate the sandstone layer into blocks. The space between these blocks is filled with clay. This suggests that during deformation the sand was much stronger than the clay, indicating that the clay was very weak (probably water-rich) during the faulting. Supporting evidence for this is found in the scaly microfabrics of the clay-rich gouge (Fig. 9), which are identical to those reported from active overpressured décollements in mud-rich accretionary prisms (e.g., Agar et al. 1989). Evidence for the brittle deformation of the sandstone and the ductile deformation of the shale is also observed at larger scales. Fault 3 (Fig. 8) shows folded shale layers (although some localization is present), above a sharply faulted sandstone bed originating from a pre-existing joint. Thus the deformation style and fault zone structure are strongly influenced by the contrasting mechanical properties of the sandstone and shale.
Fig. 7. (a) Reactivated joints in a rigid sandstone bed. (b) Example of reactivated joints in Fault zone 14.
Where faults have formed splay and lens-shaped structures, we observe a higher degree of deformation inside the lens (or between the splays) than outside, implying a very wide overall zone of deformation. An example is the large lens of Fault 8, where folding and faulting inside the lens tip were observed (Figs 2c & 4g), and the right splay forming the lens contains lens-shaped bodies which are strongly deformed themselves. The phenomenon of lens generation and deformation in fault zones is analyzed in more detail in van der Zee & Urai (2005). Other examples are the faults with large offsets (e.g., Fault 11, ‘Faille Nord’ in Figs 2b & 4i) which consist of different (sub) parallel fault strands separating zones with different degrees of deformation. Subsequent

Fig. 8. Detail of the lower part of Fault 3 showing folded shale layers and a sandstone layer with discrete faults. The measuring tape is approximately 1 m.
continued movement incorporates the lesser-deformed zones into the fault zone.

Several authors have discussed the existence of a uniform displacement-thickness ratio for faults (Scholz 1987; Hull 1988; Blenkinsop 1989; Evans 1990; Knott et al. 1996). Yet fault gouge thickness can vary greatly along a fault. The thickness of a single fault strand in the Tréviels open-cast pit ranges from <1 mm in the reactivated joints in sandstone beds, which did not slip further than the sandstone bed thickness, to 30–40 cm in the shaly intervals. Fault 7, for example, shows a much thicker fault gouge in areas where shale is juxtaposed against shale than the areas where the sandstone is juxtaposed against sandstone. This variation in thickness of 1–1.5 orders of magnitude is often observed and is clearly a problem in defining a simple and accurate relationship. Large thickness variations along strike are also observed are reported by other authors (Foxford et al. 1998; van der Zee & Urai 2005).

The definitions and interpretations of fault thickness are problematic, because:

- a single fault has a variable thickness along strike and down-dip;
- the definition of the edges of a fault zone is very subjective.

To define the thickness of a fault zone, the edges of the fault have to be identified. This definition is mostly very subjective and will vary between different authors (Blenkinsop 1989). Some authors (e.g., Knott et al. 1996) prefer to sum the thicknesses of individual strands, thereby discarding the (supposedly undeformed) material between these strands. The problem of this method is that in sequences without marker horizons it is hard to determine if the material is deformed or not, which makes the thickness determination lithology-dependent, which is unwanted. In the case of the normal faults in the Lodève basin, lenses of relatively undeformed material were commonly imbricated into the fault zones by sequential displacement on successive fault strands. Thus, a practical solution was to measure the widths of highly deformed fault strands, and also of the overall fault zone width including lenses of less-deformed material and zones of rotated bedding, i.e., an overall fault zone width. Nevertheless, distinguishing ‘highly deformed’ state in the shales is often subjective because the destruction of bedding may not be recognized if a shaley foliation is generated in the shale-rich shear zone. Hence the distinction should be treated with caution.

The fault zone thickness of a single fault strand in the Tréviels pit-mine ranges from <1 mm in the slip-reactivated joints, which did not move further than a sandstone bed thickness, to 30–40 cm in the shaley intervals. Along with the two different types of fault zone width measurement, fault zone offset at the point of width measurement was also made. The measurements were performed at

![SEM image of shale gouge in Fault zone 14.](image)
different points along the fault trace because the fault gouge thickness can vary strongly along the fault. The displacement-thickness data are plotted together with data from the Airport road outcrop, Miri, Malaysia (van der Zee & Urai 2005) and literature data of Evans (1990) and Foxford et al. (1998) (Fig. 10). The data illustrate the large variation in thickness even along a single fault, highlighting the danger in using an average displacement–thickness relationship for predicting fault zone thicknesses, as well as the difference between highly deformed zone thickness and ‘total’ fault zone thickness. Nevertheless, such data can be used in probabilistic estimates of the uncertainties in estimating fault zone thickness from throw, for example.

Discussion

Fault orientations

We interpret our data to show that the faults in the sandstone beds originated as slip-reactivated joints. We conclude this because on many occasions the orientation of the fault planes in the sandstone is the same as the orientation of the joint surfaces in the same unit (Fig. 3). Further, these faults typically have little or no damage in the wall rock, and in the case of the smallest faults, segments generated as slip-reactivated joints have low displacement–length ratios in the sandstone beds in comparison with typical data in the literature, and in such cases, the sandstone beds have demonstrably lower displacement gradients along any one fault than the shale-rich portions of the stratigraphy (Wibberley et al. 2007).

There are several possible reasons for the range of normal fault orientations found in the outcrop. In a heterogeneous stress field, the fault could initiate in different orientations. Such a heterogeneous stress field may be caused by interaction between the regional stress field and a basement fault below the outcrop. Another possibility is that the lower-angle faults initiated while the bedding was close to horizontal, and they reached their present orientation after basin-wide tilting (Agnon & Reches 1995). After, or in the late stages of, the tilting the steeper faults initiate and therefore their dips equal those predicted by Anderson (1942). Such a fault distribution pre- and post-rotation is illustrated in Wibberley et al. (2007).

Although the data do not allow a definite test of the different models, we favour the model of faulting during progressive tilting (Wibberley et al. 2007). In the outcrop described here, the large normal faults with their southern block moving downwards are very steep or overturned, suggesting

![Fig. 10.](image-url)
that these formed before the tilting and reached their present orientation after tilting. In the northern part of the outcrop some low-angle normal faults are present. These faults are bedding parallel and are probably using the weak bedding interfaces as slip planes. These faults are interpreted to have initiated after tilting, because (i) they offset other faults, and (ii) the dip of these faults before tilting would be too low to act as slip planes.

Relationship between faults and bedding orientation

A schematic block diagram (Fig. 11) illustrates the relationships between the structures of different orientations. In the outcrop, the bedding mostly dips 30° to the south (case a in the block diagram Fig. 11). However, in areas close to faults some north-dipping bedding can be found. The two cases observed for this are:

- The bedding is rotated by simple shear between two faults, such as the bedding in the top of the lens of Fault 8, or between Faults 6a and 6b. This is case b in the block diagram (Fig. 11).
- A fault termination below a layer causes bending of the layer over the fault tip, such as above Fault 6a. This is represented by case c in the block diagram (Fig. 11).

Fault trace geometry

Most of the faults observed in the Treviels pit mine often have an irregular, strongly non-planar geometry. We interpret this appearance as being caused largely by the difference in orientation of the fault in the sandstone beds compared with the shale beds (e.g., Fault 3, Figs 4b & 8). In the sandstone beds, the fault is very often perpendicular to bedding, because it is a reactivated joint, whereas in the shale it is a link between these reactivated joints. This illustrates how the fault shape and orientation is controlled by the distribution and orientation of the initial discontinuities (the joints) in the sandstone and the contrasting material properties between the shale and the sandstone. The present day dip of the fault in the shale intervals can be lower than it was during faulting due to compaction of the shales during burial (Davison 1987; Wang 1995).

Fig. 11. Schematic block diagram not representing the reality in detail, but a model geometry to understand the observed structures and orientations. The letters are key features referred to in the text.
The movement of the reactivated joints parallel to the joint surface can cause problems for large fault displacements because a space problem occurs in the shale. We note that we did not observe separated joint surfaces as long as the throw was smaller than the sand layer thickness (cf. Ramsay & Huber 1987; Peacock & Sanderson 1992; McGrath & Davison 1995). The commonly observed splays of the fault inside the shale beds just above sandstone beds are probably in response to this space problem (e.g., Fault 2, Fig. 4a). A possible evolution of the fault structure around a reactivated joint is shown in Figure 12.

**Fault gouge thickness and composition**

The displacement–thickness data for the Tréviels outcrop show a similar scatter to the global trend taken from the literature (Fig. 10). It is noticeable that, even if the same definition of fault thickness is used across the whole outcrop, a wide spread in throw–thickness ratios is observed. This is due to the thickness variations along dip, and an abrupt step in the fault thickness related to the throw–bed thickness ratio between:

- small displacement faults (< bed thickness);
- medium-large displacement faults (≥ to > bed thickness).

**Small displacement.** Sandstone–sandstone contacts are very thin (c. 1 mm), and are often reactivated joints. Shale–sandstone contacts show a narrow deformation zone. The lithological difference and competence contrast between the sandstone and shale favours the deformation in the weak shale layer towards the contact. Shale–shale contacts are 1–2 cm thick, and are recognized by the lining up of clay flakes or folding of the shale layers causing a broader zone of deformation.

**Medium-large displacement.** Sandstone–sandstone contacts have the same range of thickness as the other contacts after slip further than the sandstone bed thickness. The contacts of sandstone–shale

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**Fig. 12.** Schematic cartoon showing the reactivation of a joint as observed in the Tréviels pit-mine. The observed foliation in the shale is indicated. An example of such a shale-filled pull-apart structure is observed in Fault 2 (Fig. 4a).
show a narrow zone of deformation along the sandstone, but also often a short-cutting structure, which will widen the fault zone. Shale–shale contacts merely consist of an anastomosing network of slip planes with variable width and degree of deformation (Fig. 4i).

In all cases of sand–sand juxtapositions and throw of more than the bed thickness, a consistent seal of clay is found on the fault surfaces. This coating resembles the clay smear structures as described by Lindsay et al. (1993) in which a thin veneer of clay coats the fault surface of the sandstone bed, and shale has often either been injected or imbricated into the fault zone between juxtaposed sandstone blocks. Places in the fault zone with very thick clay material are mostly related to multiple strands (‘pseudo gouge’, see van der Zee & Urai 2005). For example, where a fault offsets a sandstone bed more than its bed thickness, a pull-apart structure is formed (e.g., Fault 2; Fig. 4a). This opening is filled with shale, forming a locally thick, shale-rich fault zone. An example is the Faille Nord (Fig. 4i), in which the fault zone consists of shale and fault breccia. In the breccia parts the original bedding is recognizable, albeit with difficulty in places, suggesting that ongoing movement incorporated the lesser-deformed zones into the more highly deformed part of the fault zone.

In previous articles (e.g., Wojtal & Mitra 1986) the importance of strain hardening in widening or localizing the active zone of deformation and strain softening in the fault zone is discussed. We think that the laboratory-derived data on this behaviour are not very suitable to simulate a natural fault gouge, because laboratory fault surfaces are not rough enough. Our data on medium to large displacement faults show that the fault roughness plays an important role in the fault gouge development.

Fault gouge generation and coupled localization in a fault is described by Tullis (1999). He reports that movement on a rough fault plane causes asperities to be sheared off and that a localized (short-cutting) slip zone develops in the generated fault gouge. Waterson et al. (1998) also describes this process of fault gouge generation by fault roughness caused by bedding parallel slip, oblique to a normal fault. In our case the roughness is largely caused by pre-existing joints (see fault shape section) and splays, as illustrated in Figure 12, leading to a fault zone of variable thickness due to the generation of lenses of less-deformed material bounded by high-deformation fault strands.

It is argued by several authors (e.g., Hull 1989) that the fractal roughness of the fault would generate a constant throw-thickness ratio because, with further throw, an asperity with a larger amplitude and wavelength would be sheared off. This is not necessarily true, because fault planes can be self-affine instead of self-similar (Power & Tullis 1991); this means that, with increasing displacement, the wavelength–amplitude ratio does not stay constant (Develi & Babadagli 1998). The result of this is that the shearing of larger wavelength asperities does not automatically cause a linear increase in gouge thickness.

We can conclude that a global displacement–thickness relationship is not appropriate for the details of this outcrop. As reviewed by others (e.g., Evans 1990; Knott et al. 1996) this relationship is more an artefact of the log–log presentation than the result of a physical process. The thickness definition is very subjective, and is dependent on authors, lithology and resolution of observation. Down-dip thickness variations of 1–2 orders of magnitude will also cause a large scatter of the throw–thickness ratios.

Conclusions

The faults exposed in the Tréviels pit-mine have different orientations because of the different timings of the faulting relative to the tilting of the bedding. Many of the faults follow bedding-perpendicular joints in the sandstone beds. Hence those faults generated oblique to bedding, typically early on in the tilting history, have irregular stepped geometries due to different orientations in the sandstone and in the shale beds. However, those faults that propagated perpendicular to bedding after bedding tilt by 25–30° tend to be planar, and were assisted by the pre-existing joints. The operation of bedding-parallel slip during faulting can also increases the complexity of the fault zones and fault surface geometry.

The complicated fault shape (pre-existing joints and bedding parallel slip), particularly of the earlier faults formed oblique to bedding and rotated by regional tilting during their continued activity, causes extra deformation of the wall rock during ongoing deformation. The less competent shales between the rigid sandstone layers deform to overcome space problems by initiating splays at irregularities in the fault surface which re-connect to leave isolate lenses within the fault zone. These lenses may be of broken sandstone beds, or weak shale that has flowed into pull-aparts between separated sandstone blocks. Microstructural evidence suggests that during faulting the sandstone deformed by brittle fracturing whereas the shale deformed by ductile mechanisms, being much weaker than the sandstone and probably water-rich.

The global gouge thickness–throw relationship does not hold for this outcrop. For a single fault, the fault zone thickness often varies widely up- and down-dip. This variation in thickness is...
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