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Initiation of brittle faults in the upper crust: a review of field observations

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Abstract

A review of field observations reveals that faults in the upper crust initiate in one of three styles: linkage of pre-existing structures, linkage of precursory structures, or by localization of slip in a precursory shear zone. The first two styles can be characterized by three general stages. During stage 1, faults initiate by shear along pre-existing structures (formed during an earlier event), or by the initiation of precursor structures (formed earlier in the same deformation event). Stage 2 involves the pre-existing or precursor structures becoming linked by differently orientated structures, as stresses are perturbed within the developing fault zone. Linkage allows displacement to increase. A through-going fault develops during stage 3. The third style of fault initiation requires the development of a shear zone, which may change local mechanical properties and lead to faulting. These observations highlight the importance of mode I fracturing in the initial stages of faulting. Rock architecture and the orientation of layering with respect to the principal stresses exert a strong influence over the style of fault initiation. Most brittle faults in the upper crust initiate from precursory structures, and the mechanism for fault growth changes with scale. Thus we should not expect a simple, universal slip–length scaling relationship for faults.

Keywords: Fault initiation; Fault scaling; Joints; Veins; Pressure solution seams; Deformation bands

1. Introduction

A review of new and previously published field observations of exposure-scale brittle faults gives an insight into the initial stages of faulting in the upper crust. From the observations reported below, we recognize three styles of fault initiation: initiation from pre-existing structures, initiation with precursory structures, or initiation as continuous shear zones. These styles of initiation and subsequent growth influence fault damage zones and therefore ultimately control fault-rock characteristics and properties, including fluid flow along the faults (e.g. Knipe, 1997; Hesthammer and Fossen, 2000). Recognition of the styles of fault initiation and propagation is also important to understand the way in which strain accumulates in a region (e.g. Reches and Lockner, 1994; Cowie et al., 1995; Knott et al., 1996). Faults commonly form complex zones of interacting and linked segments, and this segmentation is, in part, a consequence of the manner in which the fault initiates and grows (e.g. Aydin and Schultz, 1990; Peacock and Sanderson, 1996). Understanding the styles of fault initiation and propagation is therefore of great importance in understanding fault-rock properties and fault populations.

Despite the importance of these observations to structural, resource and hazards analyses, and despite a vast literature describing fault structures, there is not, to date, a published review of the styles of fault initiation, and no published systematic attempt to determine the factors that control these styles. Different styles of fault initiation have been described from various areas, tectonic settings and lithologies, but such descriptions have not been synthesized. Here we review some important styles of brittle fault initiation from observations of rocks that were deformed at or near the Earth's surface. Our goals are to illustrate how detailed analysis of areas around fault terminations can give valuable information about how faults initiate, and to discuss the factors that influence the style of fault initiation.

2. Field observations of meso-scale faults

This contribution reviews and illustrates the most common styles of fault initiation, especially those that

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occur in sedimentary rocks in the upper crust. Our emphasis is by necessity on faults with small amounts of slip (meters or less), because these structures illustrate faults in their early stages. For the field examples we illustrate here, the termination zones are analyzed to determine the propagation styles and histories of the faults, and schematic models are presented for their evolution. A simple method is to study the area beyond the fault tip, and to describe the progressive development of a fault by moving towards the tip and along the fault. In this way we use space as a proxy for time: structures at and around the fault tip are presumed to represent the earliest stages of fault development, and structures behind the tip, toward the center of the fault, are presumed to represent later stages. Fieldwork involved qualitative analysis of the conditions of deformation, with the aim of determining the mechanical controls on the different styles of fault initiation and propagation. Cited field studies that we did not conduct were selected to follow the same methodology.

For the purposes of this study, we define *fault* broadly, to include a surface, composite surface, or zone of deformation across which there is a displacement discontinuity that includes (but is not limited to) slip parallel to the surface or boundaries of the zone. This definition may encompass so-called *shear fractures*, which are single surfaces of displacement discontinuity thought to initiate in shear. The definition is distinct from *shear zone*, which is a region of localized but continuous displacement.

Field studies show that many faults develop from earlier-formed meso-scale structures. These structures are most commonly joints (mode I fractures), veins (filled mode I fractures), or pressure solution surfaces (mode −I fractures or 'anticracks'; Fletcher and Pollard, 1981). We make the distinction between pre-existing and precursory structures. Pre-existing structures are those that have formed at an earlier time in a stress field apparently unrelated to the faulting. Precursory structures are those that form as an early stage of faulting, in the same stress field (these are the 'premonitory' structures of Johnson (1995)). Here, we review observations of fault initiation associated with joints, veins, pressure-solution surfaces and deformation bands.

2.1. Fault development from pre-existing structures

The strike–slip faults at Marsalforn, Gozo, Maltese Islands (latitude 36°5′N, longitude 14°11′E), are well exposed on gently dipping Miocene Lower Globigerina Limestone (Fig. 1a; see also Kim et al., 2003). This is a massively bedded, planktonic foraminiferal packstone to wackestone, up to ~40 m thick (Pedley et al., 1976). It was probably buried less than 1 km. The faults appear to have developed in four stages. (1) At some time before the initiation of faulting, an array of en échelon joints develops parallel to the direction of greatest compressive stress (σ_1). These joints become pre-existing structures to the fault. (2) The direction of regional compressive stress rotates, causing slip across the joints. (3) Wing cracks (also known as *tail cracks*) develop at extensional steps between the en échelon joints, leading to the development of pull-aparts as displacement increases (Fig. 1a). (4) Further displacement causes rotation of blocks that are bounded by the pull-aparts, and the end result is a fault zone with brecciation. This sequence of events (1–4) is illustrated in Fig. 1b–e.

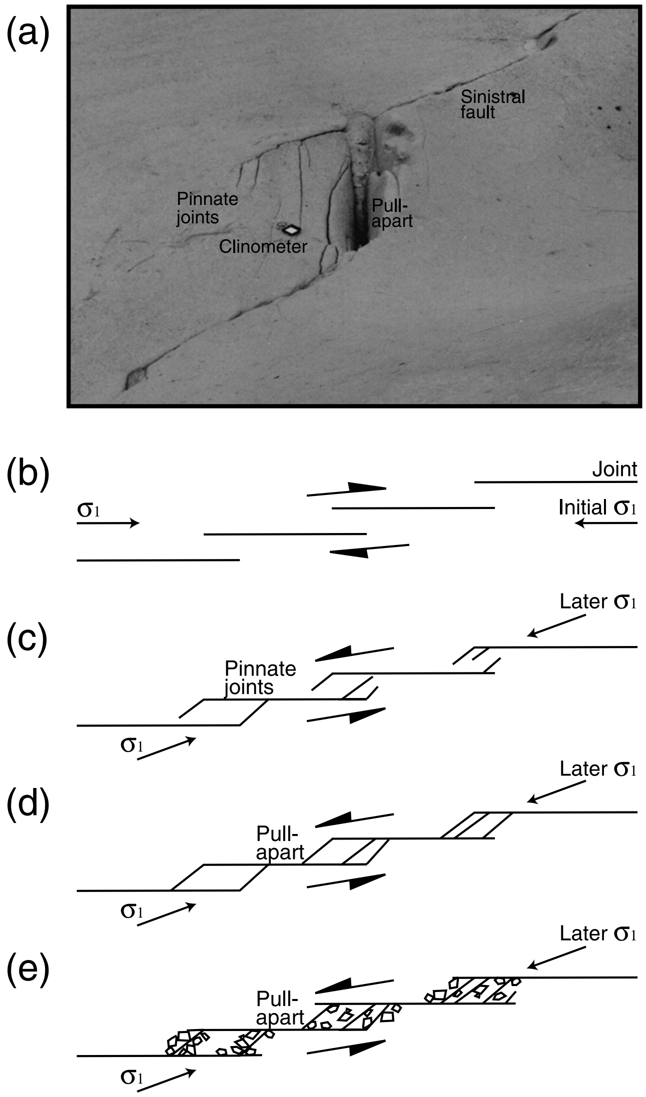


Fig. 1. (a) Photograph (oblique view of a horizontal surface) of a sinistral strike–slip fault zone that initiated as stepping joints, at Marsalforn, Gozo, Maltese Islands (see Kim et al., 2003). (b)–(e) Evolutionary model for the development of the strike–slip faults at Marsalforn. Here and in all other figures, σ_1 represents the maximum compressive stress. (b). An initial stress state produced an array of stepping joints (left-stepping in the example shown). (c) The stress system changed, driving sinistral shear on the joints, and causing the development of wing cracks in the extensional quadrants of the faults. (d) Pull-aparts developed between faults linked by wing cracks as slip increased. (e) A through-going fault developed, with brecciation at the pull-aparts.

Pre-existing extension fractures also control the development of strike–slip faults in granites of the Sierra Nevada, California (Segall and Pollard, 1983; Martel et al., 1988; Martel, 1990; Bürgmann and Pollard, 1994), where the faults followed meter-scale joints that were formed in a different stress field. A similar evolution of the resulting fault zone is observed: wing cracks propagated from the ends on the faulted joints, at a high angle to the joint plane, and lengthening of the faults occurred by the linkage of the faulted joints by wing cracks. Granier (1985) describes fault development in other granites, gneisses and schists along pre-existing joints that are linked by wing cracks. Cruikshank et al. (1991) and Myers (1999) describe similar structures in sandstone, while Petit and Mattauer (1995, fig. 8) suggest that the faults in the limestones of the Matelles exposure, Languedoc, France, formed along pre-existing joints. Wilkins et al. (2001) illustrate structures associated with bed-bounded joints in sandstone that become reactivated as normal faults. In cross-section, the relationship between the faulted joints and secondary fractures is the same as strike–slip faults in map-view (mode II). The geometry and mechanics of wing cracks around fault terminations and that link fault segments are further described by Fletcher and Pollard (1981), Martel (1990), Cruikshank et al. (1991), Cruikshank and Aydin (1994), Cooke (1997), Martel and Boger (1998), and Willemse and Pollard (1998).

2.2. Faults that develop with precursory structures

2.2.1. Precursor joints

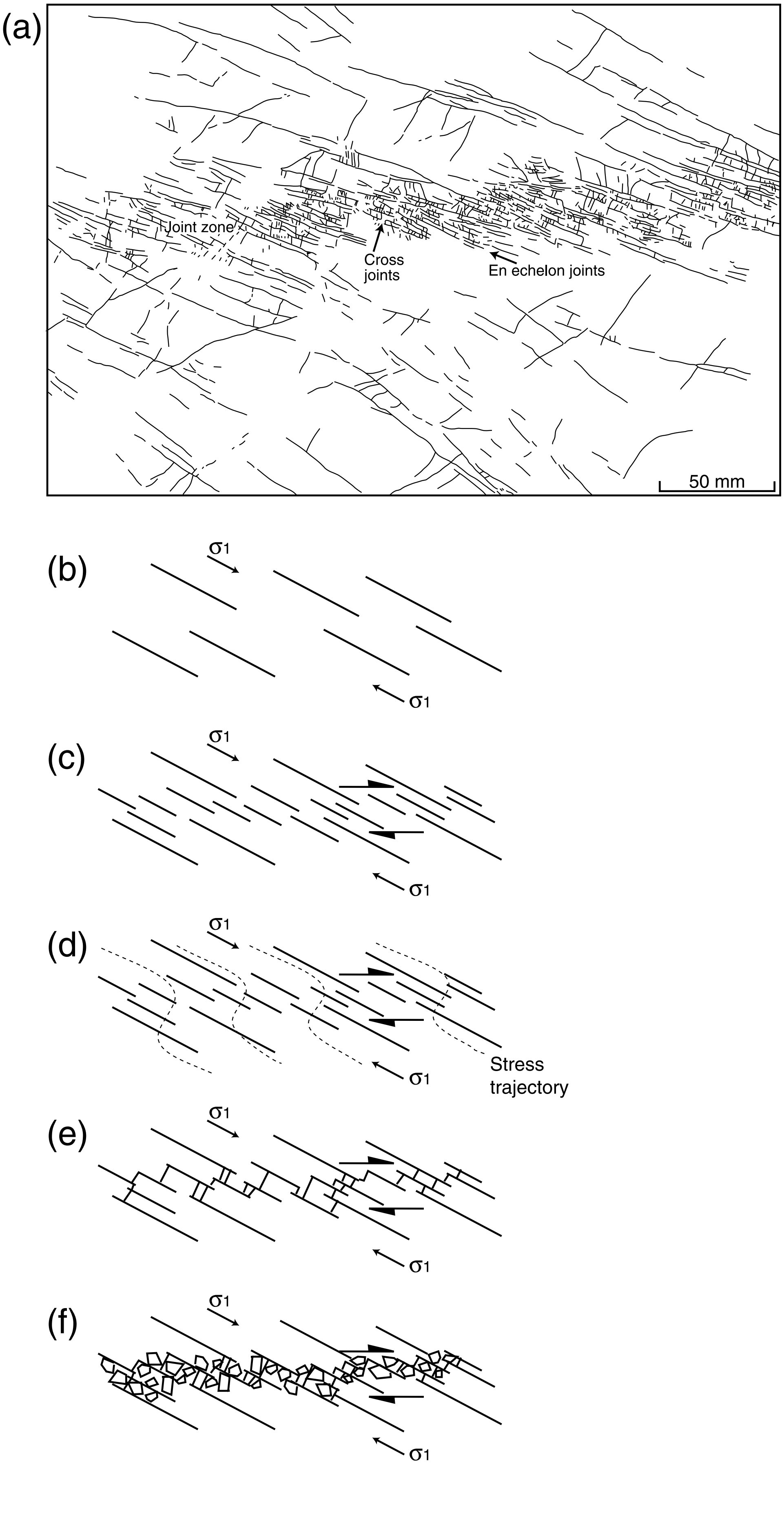
Mollema and Antonellini (1999) describe strike–slip fault zones in dolomites of the Sella Group, in the central part of the Dolomites, northern Italy (latitude 46°31′N, longitude 11°51′E). These Mesozoic dolomitized reef carbonates were deformed during the Alpine Orogeny, and were probably not buried below about 1 km (Mollema and Antonellini, 1999). The faults are surrounded by a dominant joint set, with blocks of wall-rock commonly included in the fault zone. An incipient fault zone is shown in Fig. 2a, and a model for these faults is shown in Fig. 2b–f. This model involves the development of a system of joints parallel to the σ_1 direction (Fig. 2b), which became locally intense in an incipient shear zone (Fig. 2c). A set of cross-joints developed to link the first set of joints. Cross-joints commonly develop perpendicular to pre-existing closely spaced joints (Rawnsley et al., 1998) due to local re-orientation of σ_1 (Bai et al., 2002). Peacock (2001) interprets the cross-joints as having developed as the σ_1 trajectories were perturbed into the developing shear zone (Fig. 2d and e). A through-going fault developed as shear continued (Fig. 2f). Using this process, faults develop in a rock without pre-existing fractures or mode II fracture propagation, within a single evolving stress system (cf. Martel et al., 1988; Peacock and Sanderson, 1995a). A similar case is described next, where strike–slip faults initiate as en échelon veins.

2.2.2. Precursor veins

Strike–slip faults that conjugate about N–S occur in sub-horizontal Jurassic limestones and mudrocks at Kilve, Somerset, England (latitude 51°11′N, longitude 3°15′W). Peacock and Sanderson (1995a,b) and Willemse et al. (1997) describe the geometry and development of these faults. The faults (Fig. 3) initiated as vein arrays in the limestone beds. Pressure solution seams occur in the bridges between the vein segments, approximately perpendicular to the strike of the veins. Shear occurred on the pressure solution seams as the veins widened and the bridges were rotated (Willemse et al., 1997). The pressure solution seams or fractures that linked portions of the veins allowed the development of pull-aparts (Fig. 3), with a through-going fault developing as the pull-aparts overlapped. Thus, the fault evolves from precursory structures and no regional stress rotation is required (Willemse et al., 1997). Peacock (1996) presents a similar model for semi-brittle shear zones in the Plymouth Limestone. Rispoli (1981), Gamond (1983, 1987), Petit and Mattauer (1995), and Petit et al. (1999) describe similar structures in limestones.

The excellent exposures of Upper Triassic and Lower Jurassic limestones and mudrocks on the Somerset coast (latitude 51°11′N, longitude 3°15′W) have also provided valuable information about the geometry and development of normal faults (Peacock and Sanderson, 1991, 1992, 1994, 1996). Peacock and Sanderson (1992) observe that normal faults exposed on limestone bedding planes die out laterally into veins that strike parallel to the faults (Fig. 4a). This suggests that the faults initiated as veins (or as open fractures that were later mineralized). The faults propagated laterally through the limestones by linkage of relatively simple extensional or slightly transtensional vein arrays (Figs. 4a and 5), and commonly propagated vertically through precursory monoclinal folds (Fig. 4b). The faults are usually steeper in the more brittle limestones than in the mudrocks, and consist of mud- or calcite-filled pull-aparts in the limestones (Peacock and Zhang, 1994, fig. 3a; Canole et al., 1997, fig. 3; Gross et al., 1997a). Veins at the lateral fault terminations have an average width of about 20 mm at the point where throw on the faults decreases to zero, suggesting that faults can propagate from the extension fractures into the mudrocks when the extension fractures are about 20 mm wide. Davison (1995) describes crack-seal in the pull-aparts along normal faults in Somerset, and similar examples are described by Lee et al. (1997) and Lee and Wiltschko (1999).

A model for normal fault development in multi-layered sequences is shown in Fig. 5. Veins develop in the more-brittle layers, with the less-brittle layers undergoing ductile deformation, as a rock sequence is strained with the maximum compressive stress perpendicular to bedding (Fig. 5a). Fractures propagate from the thicker veins into the adjacent less-brittle layers (Fig. 5b). An irregular through-going fault eventually develops (Fig. 5c). A similar sequence occurs in strike–slip faults that are at a high angle to vertical beds (Peacock, 1991). This model is similar to that presented by Eisenstadt and DePaor (1987), who


(a)
Joint zone
Cross joints
En echelon joints
50 mm
(b)
σ1
σ1
(c)
σ1
σ1
(d)
σ1
σ1
Stress trajectory
(e)
σ1
σ1
(f)
σ1
σ1

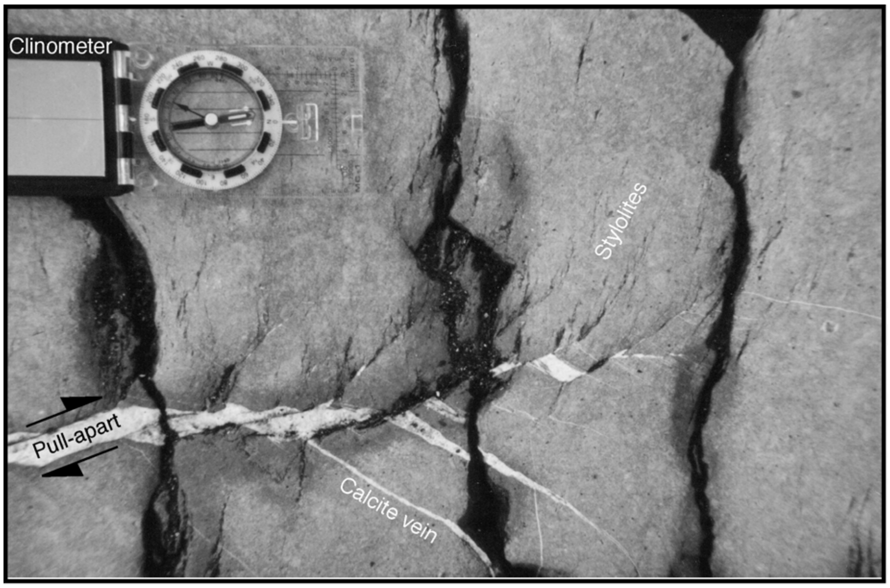


Fig. 3. Photograph of a strike–slip fault zone on a Liassic limestone bedding plane at East Quantoxhead, Somerset (horizontal surface). The fault consists of en échelon vein segments linked by shear fractures to form pull-aparts (e.g. Peacock and Sanderson, 1995a; Willemse et al., 1997). The fault dies out into a zone of contraction marked by pressure solution seams (in the upper portion of the photograph) and a zone of extension marked by thin veins (in the lower portion of the photograph).

show thrust faults that initiate as ramps in more brittle beds, later propagating into less brittle beds.

2.2.3. Precursor pressure solution seams

Pressure solution seams may also serve as discontinuities from which faults evolve. The Tutt Head Thrust Zone (Fig. 6a) is a zone of Variscan-age contractional deformation in a coarse grainstone unit within the massively bedded Carboniferous Limestone. The thrust zone is well exposed in a coastal cliff section near the Mumbles, South Wales (latitude 51°35′N, longitude 3°59′W). It consists of one main thrust, with a series of smaller synthetic and antithetic thrusts in both the footwall and the hanging-wall. Strain associated with the thrust zone is described by Hyett (1990), who shows that the main thrust plane dips at ~43° towards 110°, and has a maximum observed displacement of ~3.5 m. Slickenside lineations on the thrust indicates hanging-wall transport towards ~018°. Hyett (1990, figs. 6 and 7) suggests that the exposed portion of the main thrust represents a propagating termination of a lateral ramp.

The minor thrusts at Tutt Head appear to have initiated as arrays of pressure solution seams (Fig. 6b), oriented at ~30° to the zone boundaries. The seams are linked by veins to produce a through-going fault. This model is similar to that shown in Fig. 2 (precursory joints), because it requires stress perturbation to link precursory structures. Knipe and White (1979), Ramsay and Huber (1987, fig. 26.44), Price and Cosgrove (1990, fig. 18.24) and Peacock and Sanderson (1995a, fig. 9d) show similar transpressional arrays with pressure solution seams. Ohlmacher and Aydin (1997) describe veins and pressure solution seams related to thrusts, and suggest that pressure solution seams form during inter-slip periods and fault-parallel veins form during slip events.

2.3. Faults from precursory shear zones

The foregoing examples required the presence or preliminary development of mode I or −I fractures for fault initiation. Other faults initiate through the development of shear zones. Deformation bands are thin zones of porosity reduction and shear across which millimeters of offset may be measured (Fig. 7a). Deformation bands in granular materials have been recognized and described by many workers (e.g. Aydin, 1978; Jamison and Stearns, 1982; Karig and Lundberg, 1990; Wilson et al., 2003). Deformation bands do not contain a surface of displacement discontinuity (Antonellini et al., 1994). Thus, strictly speaking, they are not themselves faults but are best described as semi-brittle shear zones (e.g. Davis, 1999). At least two types of deformation bands are recognized

Fig. 2. (a) Map of a zone of en échelon joints and cross-joints formed in dextral shear in the Sella dolomites of northern Italy (Mollema and Antonellini, 1999, fig. 4). These joint sets are precursors to a fault zone. (b)–(f) Evolutionary model for a fault that propagates as a network of joints (Peacock, 2001; cf. Mollema and Antonellini, 1999, fig. 15). (b) A system of joints develops parallel to the far-field σ_1 direction. (c) The joints become locally intense within an incipient fault zone. (d) Stresses are perturbed within the zone, with the stress trajectories becoming perpendicular to the stage 1 joints to allow the cross joints to form at 90° to the stage 1 joints. (e) Cross-joints develop in the perturbed stress field. (f) A through-going fault develops within the zone, involving the rotation and disruption of the joint-bound blocks (Mollema and Antonellini, 1999, fig. 13).

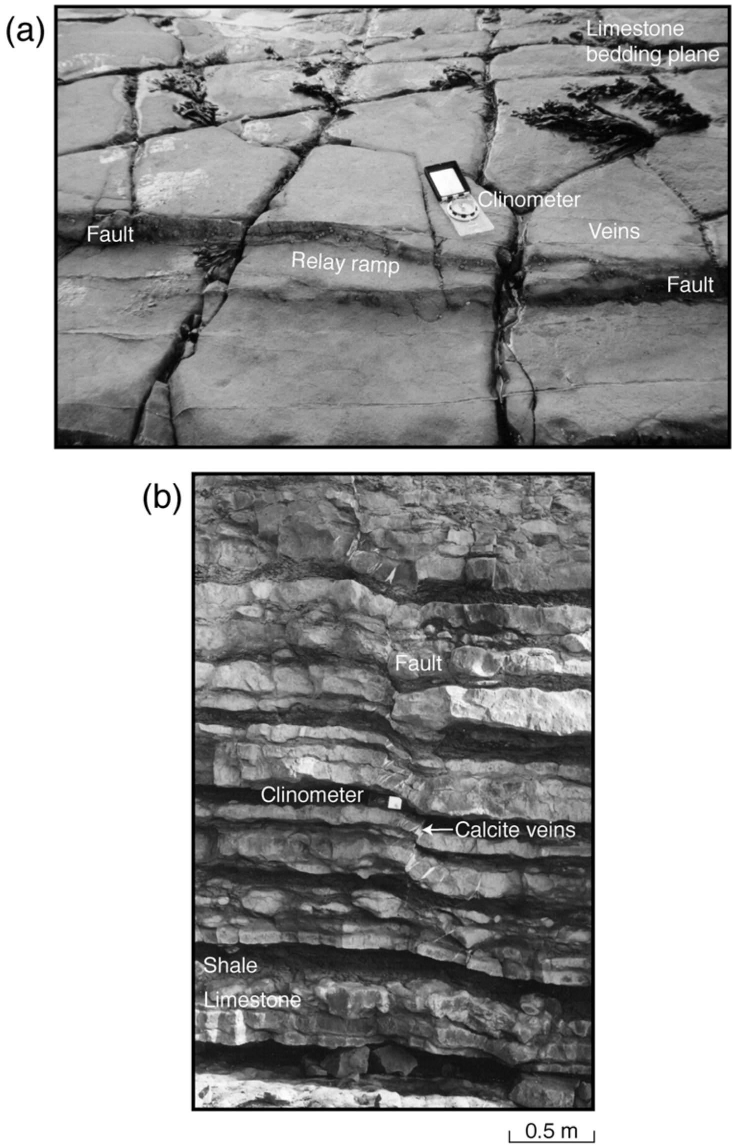


Fig. 4. Terminations of normal faults in Liassic limestones and mudrocks in the Bristol Channel Basin. (a) Photograph (oblique view) of a breached relay ramp and associated fracturing between two normal faults exposed on a limestone bedding plane at Kilve, Somerset (Peacock and Sanderson, 1994). The faults die out laterally into veins. (b) Photograph (vertical section) of a normal fault in a cliff near Southerndown, on the south Wales coast. The fault dies out downwards or upwards into a monoclinal fold, with calcite veins in the more brittle limestones (also see Childs et al., 1996).

(Antonellini et al., 1994): deformation bands with cataclasis in the zone of porosity reduction (DB1 of Davis, 1999) and deformation bands without cataclasis (DB2 of Davis, 1999). DB1 have porosity reduction compared with the undeformed host rock of up to an order of magnitude and permeability reduction of three orders of magnitude (Antonellini and Aydin, 1994). Exposures of this style of deformation band commonly form fins resistant to erosion (e.g. Aydin and Johnson, 1978; Fig. 7a).

We summarize published observations of deformation bands and the development of faults in sandstone. In a well-sorted, porous sandstone, strain is localized where the cement is weaker or where the shape of sand grains permits rolling. The rearrangement of grain contacts (from, for example, hexagonal to cubic packing) reduces contact area among grains and increases contact stresses, leading to fracturing (Antonellini et al., 1994; Fig. 7b). With continued shear and cataclasis, the grains have increased contact area, eventually increasing friction so that stresses are not sufficient to continue deformation in the band. The sequence is repeated nearby, until a zone of deformation bands have accumulated a moderate amount of slip (Fig. 7c). The zone is sufficiently large to concentrate stress as a stiff inclusion in a soft medium (Aydin and Johnson, 1983), producing a slip surface (fault) on the edge of the zone.

Other styles of continuous deformation may precede faulting. Fault propagation folding involves the bending of beds ahead of a fault termination, and may include other precursory structures described in this paper. Fault propagation folds have been described extensively in the

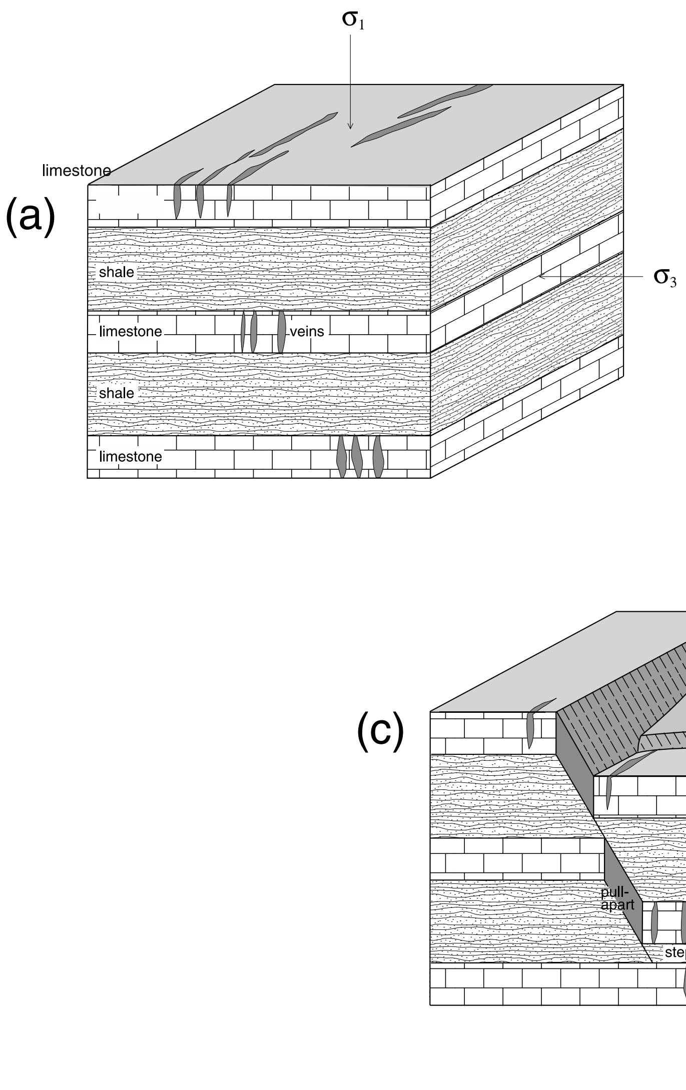


Fig. 5. Block diagrams illustrating the development of normal faults and relay ramps in a limestone–mudrock sequence (cf. Peacock and Sanderson, 1992, fig. 9). (a) Veins initiate in the brittle limestones. (b) Extension continues, with shear across the veins and connection of veins across the mudrocks. Relay ramps develop at steps between faults in map view. Displacement minima typically occur at steps (e.g. Ellis and Dunlap, 1988; Peacock and Sanderson, 1991). (c) Linkage occurs between segments, with breaching of relay ramps (Peacock and Sanderson, 1994), to produce an irregular fault zone.

literature, particularly for thrust faults, and so are not discussed here. Johnson (1995) developed a postulate of premonitory shear zones with significant evidence from precursory kink bands, deformation bands and fault propagation folds.

2.4. Fault initiation with combination styles

The Koae Fault System, Kilauea Volcano, Hawaii (Fig. 8a; latitude 19°22′N, longitude 155°17′), shows excellent examples of normal fault segments that have propagated by a series of folds, cracks and ruptures (Duffield, 1975; Parfitt and Peacock, 2001; Peacock, 2001; Peacock and Parfitt, 2002). The Koae Fault System is a zone of faulting about 2 km wide perpendicular to strike, extending from the East Rift Zone towards the South West Rift Zone. It has a dominant downthrow to the north, with individual fault segments having a throw of up to about 20 m. The Koae Fault System has accumulated about 25 m of heave since the last re-surfacing event (Duffield, 1975), 400–750 years ago (Holcomb, 1987; Wolfe and Morris, 1996). Several recent episodes of ground cracking and seismic activity in the Koae Fault System have been related to intrusion of magma from the rift zones (Kinoshita, 1967; Klein et al., 1987). Earthquake hypocenters have depths of 1–4 km (Klein et al., 1987), suggesting that the faults originate, and have greater slip, at depth (Parfitt and Peacock, 2001). The Koae Faults are likely to have greater slip at depth if they are growth faults, periodically covered by lava. Explanations for the origin of the Koae Fault System include that it is a 'tear-away' zone related to the extension of the East Rift Zone (Duffield, 1975; Swanson et al., 1976), or that it accommodates footwall uplift north of the Hilina Fault System (Parfitt and Peacock, 2001). Several relay ramps are well developed

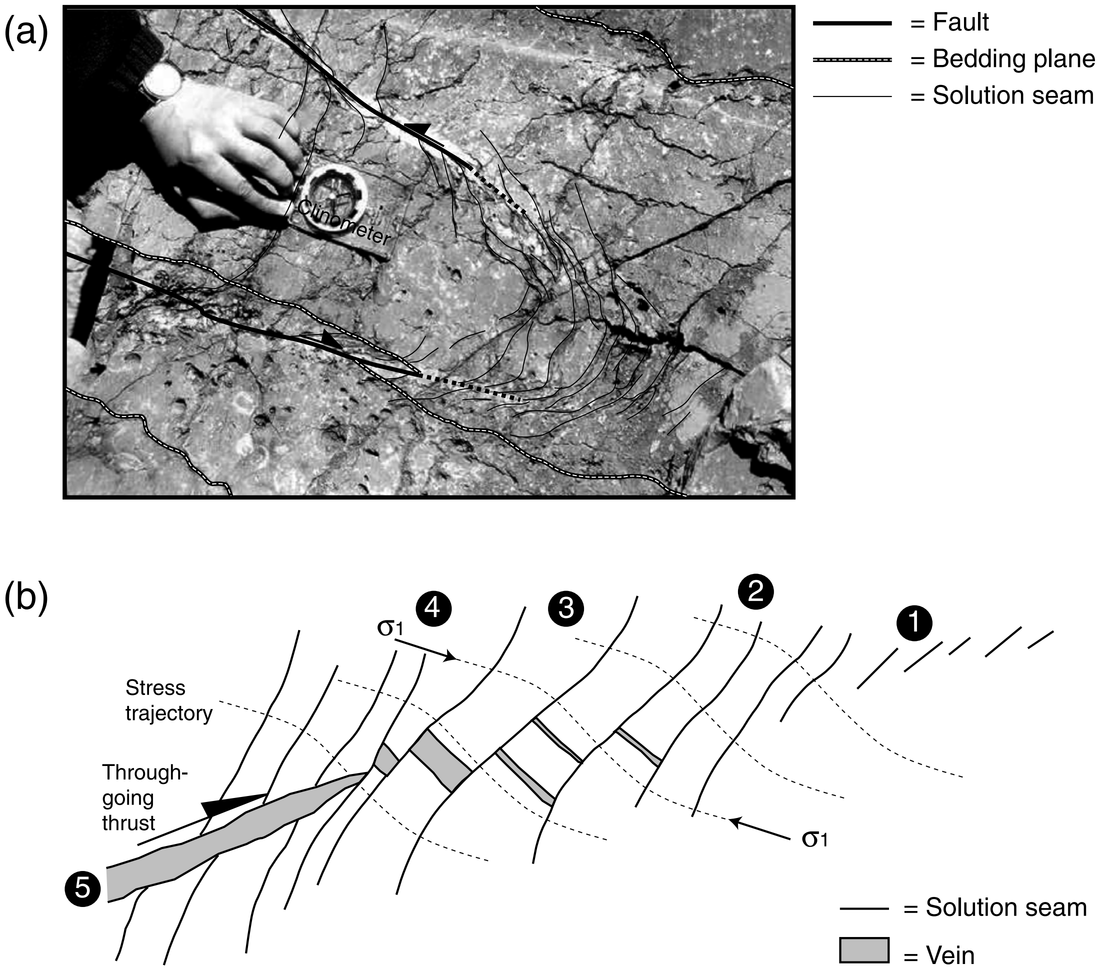


Fig. 6. (a) Photograph (vertical section) of the terminations of two conjugate dip–slip faults at Tutt Head, South Wales (see Hyett, 1990). The faults occur in thickly bedded Carboniferous Limestone, which dips at about 15° to the south. Cleavage curves between the faults. (b) Schematic diagram of a thrust termination zone at Tutt Head that can represent spatial or temporal evolution. (1) A transpressional shear zone consists of pressure solution seams furthest beyond the termination. These represent the first stage in fault development. (2) The pressure solution seams curve out of the shear zone, indicating that stress perturbation occurs around the zone. The curved solution seams may have been partly folded into their present position, but the related veins have not been rotated, indicating curved stress trajectories. (3) Calcite veins develop in the zone, linking the pressure solution seams. (4) Pull-aparts develop as the pressure solution seams and veins become linked, with shear developing. (5) A through-going thrust is developed, with calcite several millimeters thick along the fault plane.

in the Koae Fault System (Macdonald, 1957; Peacock and Parfitt, 2002); a schematic block diagram of such a relay ramp is shown in Fig. 8b.

Four sets of structures can be observed along the trace of each fault (Fig. 8b). (1) A monocline extends for tens to hundreds of meters laterally ahead of the termination of each fault. It is possible that the folds are lateral fault-propagation folds (e.g. Gawthorpe et al., 1997), or that the faults are present beneath the monoclines (S. Martel, pers. comm., 2001). (2) Vertical cracks develop along the hinges of the monoclines, especially in the outer arc (footwall side). When the cracks occur on both hinges, they form sub-parallel traces that have a tramline-like appearance on the USGS photoquads. These cracks typically follow pre-existing cooling joints in the basalt (Peacock, 2001). (3) Some of the cracks develop throw to become normal faults, possibly as they link with a normal fault at depth. One crack is commonly an open fissure while the other has vertical throw. Some faults have heave on the hanging-wall crack, while others have heave on the footwall crack (Duffield, 1975). (4) A rollover fold with associated cracking develops in the hanging-wall. Rollover probably occurs because the normal fault is vertical at the Earth's surface, but curves to a gentler dip at depth. These four sets of structures are inferred to represent stages of fault development: precursory folding, fissuring of pre-existing joints, throw across the fissures, and 'rollover' folding as the new surface faults link to a fault at depth. This sequence shows two styles of precursory structures before faulting: monoclinal folding (shear zone) and fissuring along pre-existing joints (mode I fracturing).

Fissures and monoclinal folds are also observed in the early stages of faulting of basalt outside of Hawaii. Acocella et al. (2000) describe similar structures associated with

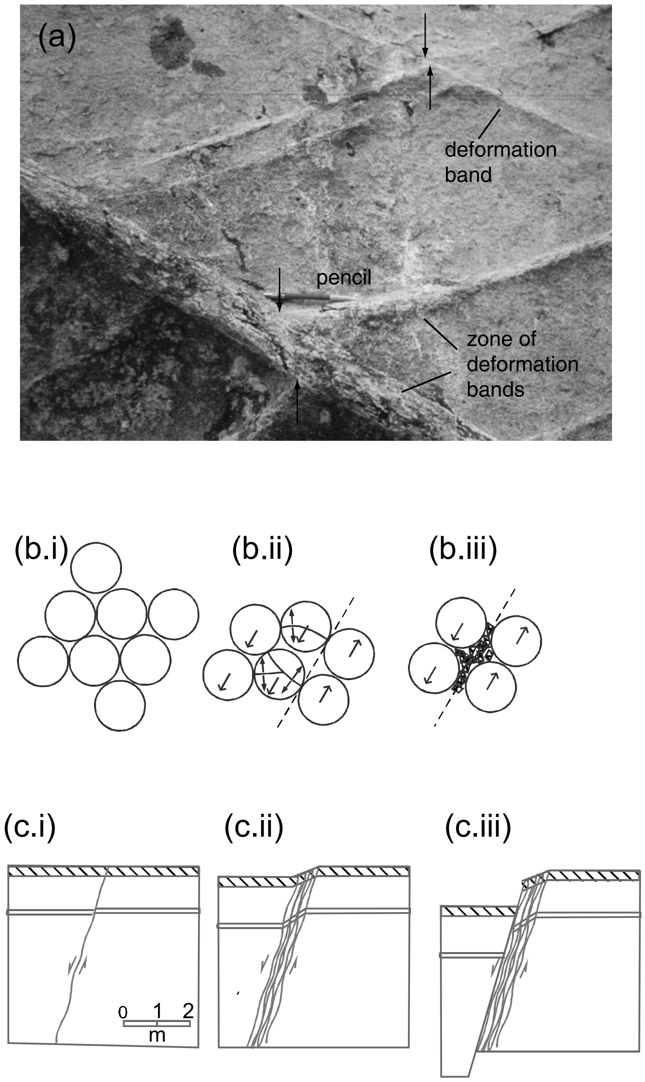


Fig. 7. (a) Photograph (vertical section) of conjugate deformation bands with apparent thrust offset (arrows) in massive sandstone at Wagon Caves Rock, Santa Lucia Range, California (Anderson, 1999). Apparent slip is greater for the wider zone of deformation bands. Deformation bands form the resistant ledge on which the pencil rests. The pencil is 14 cm long. (b) Microstructural evolution of a single deformation band. (b.i) Initial hexagonal packing of sand grains. (b.ii) Grains sliding along shear plane have fewer points of contact. Resulting stress concentrations lead to fracturing of the grains. (b.iii) Fractured grains form a zone of cataclasis and porosity reduction, i.e. a deformation band (redrawn from Antonellini et al., 1994). (c) Stages in the evolution of a deformation band fault. (c.i) A single deformation band deflects the marker horizon. Strain hardening prevents further slip on the band. (c.ii) A zone of deformation bands forms as each successive band reaches the strain hardening limit. The marker horizon is deflected in a continuous shear zone. (c.iii) A fault forms at the margin of the deformation bands when the zone is large enough to concentrate stresses as a stiff inclusion. The fault is a surface of displacement discontinuity (redrawn from Antonellini and Aydin, 1995).

active normal faults in basalt in Iceland. Muffler et al. (1994) and Crider (1999) document monoclines and fissures near the terminations of Quaternary normal faults in the volcanic plateaux of northern California. At Canyonlands National Park in southeast Utah, pre-existing regional joints control the surface expression of kilometer-scale normal faults in sandstone. There, fault-parallel fissures and hanging-wall rollovers are also observed. Moore and Schultz (1999) suggest a similar mechanism for the origin of those faults.

3. Factors that influence the style of fault initiation

We have reviewed seven different examples of detailed field studies. Of these, five examples require the development of mode I or −I fractures prior to the development of the fault, one shows the development of faults from zones of concentrated continuous deformation, and the last shows both folding and fracturing preceding faulting. From these we surmise the importance of precursory structures to faulting.

Rock architecture, including grain size and layering, and mineralogy appear to be primary influences on the nature of precursory structures. For example, it is evident that rocks with highly soluble minerals (calcite, dolomite) are more likely to host solution seams than rocks without these minerals. Pressure solution is also enhanced in fine-grained rocks relative to coarse-grained ones (e.g. Railsback, 1993; Andrews and Railsback, 1997). In those rocks, a greater surface area is exposed to fluids for dissolution.

Coarse sedimentary or mineral grains may influence the surface texture of pre-existing joints, providing asperities that may drive further cataclasis. Grain boundaries themselves are important precursory structures to faulting, in both the formation of faults in crystalline rock (Moore and Lockner, 1995) and in deformation bands (Antonellini et al., 1994).

Cementation and porosity may have a more important influence on fault initiation than grain size in sedimentary rocks. Faults in weakly cemented rocks may initiate as shear zones or deformation bands, while faults in stronger rocks more commonly begin as extension fractures. Shipton and Cowie (2001) observe that in damage zones around faults in low porosity rocks, the more grain contact there is, the more grain-scale damage-zone fracturing occurs (Anders and Wiltschko, 1994; Vermilye and Scholz, 1998). For high porosity rocks, grain-scale fracturing is observed only within deformation bands (Anders and Wiltschko, 1994). In these rocks, additional strain may be accommodated by grain-boundary sliding and rolling. A similar set of observations is made by Steen and Andresen (1999), who compare initiation of faults in poorly- and well-cemented sandstones. The well-cemented rock showed greater grain fracturing with a predominance of through-going cataclastic zones. Faults in the more porous, weakly cemented rocks appeared to initiate by pore collapse and grain reorganization.

Mechanical layering is important to the nature of initial fault structures. Bedding planes may act as discontinuities from which faults initiate (e.g. Dholakia et al., 1998).

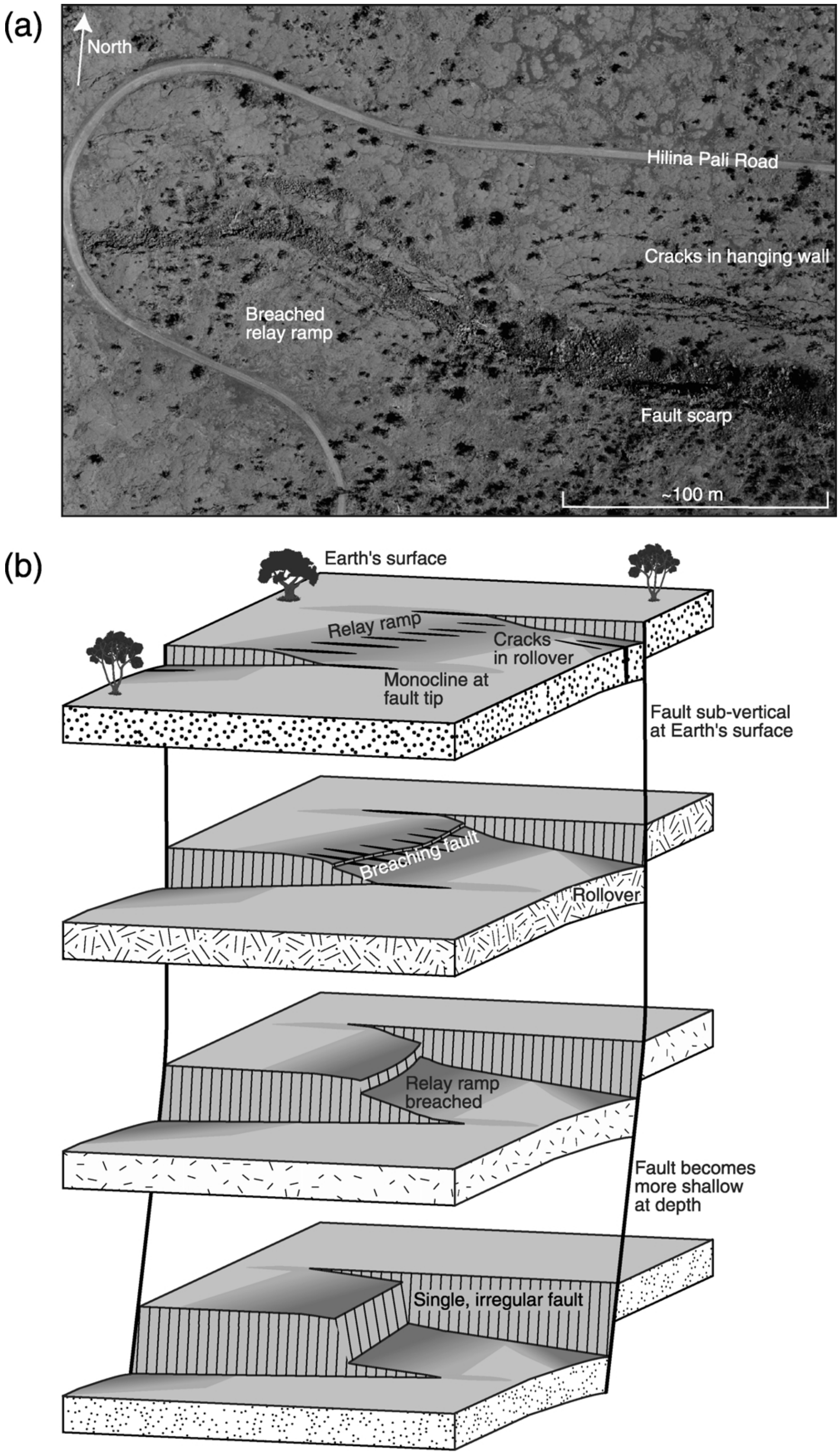


Fig. 8. (a) Aerial photograph of part of the Koae Fault System, Hawaii. The fault shown has a maximum throw of about 20 m down to the north at the right margin of the photograph. The Hilina–Pali Road loops up a relay ramp near the termination of the fault. (b) Block diagram of relay ramp in the Koae Fault System. The faults die out into monoclines and opened-up cooling joints (see Peacock and Parfitt, 2002, fig. 5).

Layering commonly controls the initial lengths and spacing of pre-existing joints (Narr and Suppe, 1991; Gross, 1993). Slip on the contact between layers may also inhibit fault propagation (Roering et al., 1997). Contrast in mechanical properties across layers influences how precursory structures interact and link. As described in Section 2.2, faults may initiate as steeply-dipping veins in brittle layers, and are connected by moderately-dipping surfaces in less brittle layers; thus, the faults appear to be deflected through the sedimentary sequence (Peacock and Sanderson, 1992; Peacock and Zhang, 1994). Laubach (1988) describes different styles of deformation in adjacent beds, with spaced cleavage in limestones and shales, and microfracturing in sandstones.

The geometric relationship between the orientation of principal stresses and layering is an important controlling factor on the style of fault initiation. Fig. 9 summarizes the relationships between layering and fault propagation, generalized from observations of small faults in layered sedimentary rocks. When the greatest compressive stress is perpendicular to layering, faults initiate as joints or veins in the more brittle layers (Fig. 9a). For example, normal faults in the sub-horizontal Triassic and Jurassic sediments of Somerset appear to have initiated as extension fractures in the more brittle limestones, and propagated as faults into the mudrocks when the extension fractures reached a critical width (Figs. 4 and 5). Strike–slip faults in sub-vertical Silurian turbidites in SW Scotland (Peacock, 1991) have very similar geometries (compare Figs. 5 and 10). When the intermediate principal stress is perpendicular to layering, faults may initiate by shear on échelon arrays of veins or solution seams (Fig. 9b). For example, strike–slip faults in Somerset initiated as transtensional to transpressional vein arrays in the limestones (Fig. 3), when the intermediate stress was normal to bedding. When the least compressive stress is perpendicular to layering (Fig. 9c), pressure solution surfaces and bedding planes may be important controlling structures for fault initiation. An example of this are the thrusts at Kimmeridge Bay, Dorset, England, described by Leddra et al. (1987) and Ramsay (1992).

The style of deformation in fault steps is also influenced by the relationship between the orientations of layering and principal stresses (Peacock and Zhang, 1994). Steps between normal faults in map view in the sub-horizontal limestone beds of Somerset are typified by bed rotation, veins, and by normal to oblique–slip faults (e.g. Peacock and Sanderson, 1994, fig. 7). Steps and bends along normal faults in cross-section in the sub-horizontal Cretaceous Chalk of Yorkshire are also strongly controlled by lithology, with fault refraction occurring as the faults pass through less brittle marl units (Peacock and Zhang, 1994). Contractional steps (map view) between strike–slip faults in the horizontal beds of Somerset are typified by more complex deformation, including bed rotation, veins, pressure solution seams, synthetic and antithetic faults, and thrusts (Peacock and Sanderson, 1995b, fig. 1).

4. Discussion

From the observations reviewed, we make some inferences about the nature and process of brittle faulting in the uppermost crust. While the conditions of deformation cannot be precisely determined from the geometric observations reviewed here, many of the rocks were deformed at or near the Earth's surface. Consequently, the rocks were probably deformed at low confining pressures and low differential stresses (Engelder, 1993). These conditions are particularly favorable for the formation of joints. Formation of joints and veins may also require increased internal fluid pressures (e.g. Secor, 1965; Sibson, 1998). Layered sedimentary sequences with contrasting permeability may result in a 'cap rock' arrangement, generating high fluid pressures in units (such as limestone) between layers of low permeability (such as mudstone). Pressure solution and deformation bands may form during burial and compaction. Thus the assemblages of structures described here may be unique to conditions of the uppermost crust.

Most of the faults studied here are extinct and propagation has been arrested. If propagation was arrested by a change in conditions, the final geometry may not be the same as when a fault was developing. We consider this to be unlikely, as our examples are drawn from many faults in many locations, and many similar observations are made in faults produced in experimental apparatus.

4.1. The role of local stress perturbation

Local stress perturbation is required to link many pre-existing or precursor structures to form an incipient fault zone. For example, stress is concentrated at the termination of faulted joints (e.g. Pollard and Segall, 1987), producing wing cracks and solution seams that connect to neighboring joints (Fig. 1). Cross joints grow between closely spaced precursor joints due to local rotation of the maximum compressive stress (Fig. 2; Bai et al., 2002). Shipton and Cowie (2001) suggest that stress perturbation ahead of a propagating normal fault can induce local deformation bands. Thus, local stress perturbation controls the initiation and early stages of faulting, enabling precursor structures to link and accumulate displacement.

Sibson (1996, 1998) provides a mechanical rationale for the irregular fault geometries observed in layered sedimentary rocks (e.g. Fig. 5), based on the two-dimensional Mohr–Coulomb failure criterion. The criterion permits simultaneous extension and shear failure in adjacent rocks of different strength under the same conditions. Field observations of fault terminations reviewed here show that the initial structures are typically extension fractures in brittle units.

The styles of fault initiation and growth described here operate at scales of centimeters to meters, the typical size range of joints, veins and other precursor structures. At fault

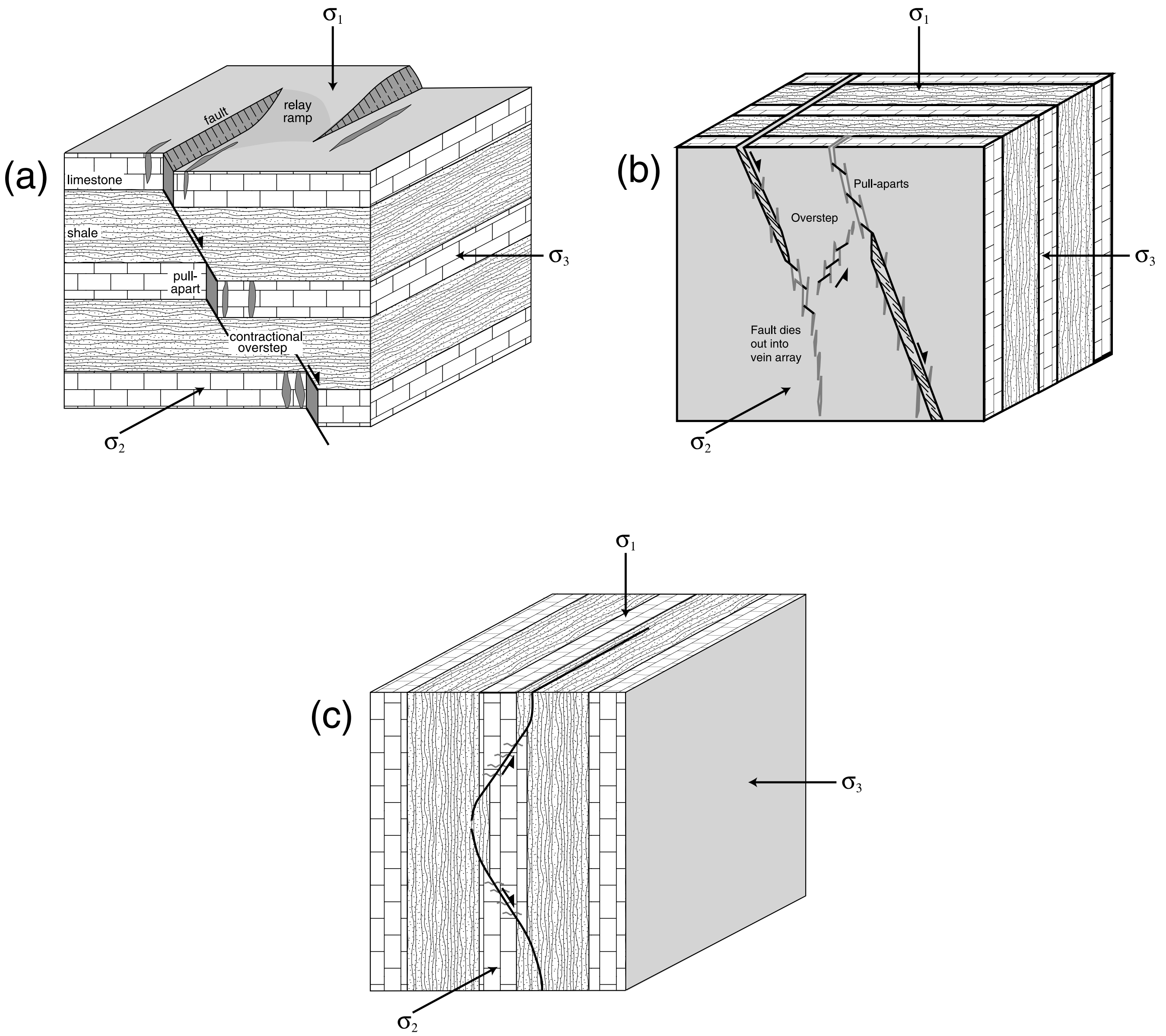


Fig. 9. Summary of the effects on fault initiation and propagation style of the relationship between the orientations of stress axes and layering (cf. Anderson, 1951; Peacock and Sanderson, 1992, fig. 12). For the purposes of illustration, the orientations of the principal axes are the same in each case, and the orientation of layering is changed. (a) σ_1 perpendicular to layering. The faults die out laterally and vertically into veins. A relay ramp occurs where the faults step in map view, with extensional or contractional steps in cross-section. (b) σ_2 perpendicular to layering. The faults have the geometries of strike–slip faults in horizontal beds. The faults die out into en échelon veins in the brittle beds. (c) σ_3 perpendicular to layering. The faults have the geometry of thrusts in horizontal beds.

lengths greater than tens of meters, the mechanisms of fault growth begin to become dominated by growth by linkage of distributed fault segments. Field observations have shown the linkage of fault segments to be important in the evolution of strike–slip faults (e.g. Segall and Pollard, 1983; Peacock, 1991; Willemse et al., 1997), normal faults (e.g. Peacock and Sanderson, 1991, 1994; Trudgill and Cartwright, 1994; Ferrill et al., 1999), deformation bands (Antonellini and Aydin, 1995; Shipton and Cowie, 2001) and thrust faults (e.g. Ellis and Dunlap, 1988). Linkage of segments takes advantage of stress perturbations and stress concentration in the zone between segments. The mechanics of fault–segment interaction have been described and modeled by (among many others) Chinnery (1961), Segall and Pollard (1980), and Kase and Kuge (1998) for the case of strike–slip faults and by Willemse (1997), Crider and Pollard (1998) and Crider (2001) for the case of normal faults. These studies show that local stress trajectories and magnitudes are perturbed from the regional values, producing secondary fractures and variations in slip distribution that correspond to field observations. Linkage of existing fault segments is a more efficient mechanism of growth than propagation from the fault terminations, because individual faults increase length in large jumps (e.g. Mansfield and Cartwright, 2001).

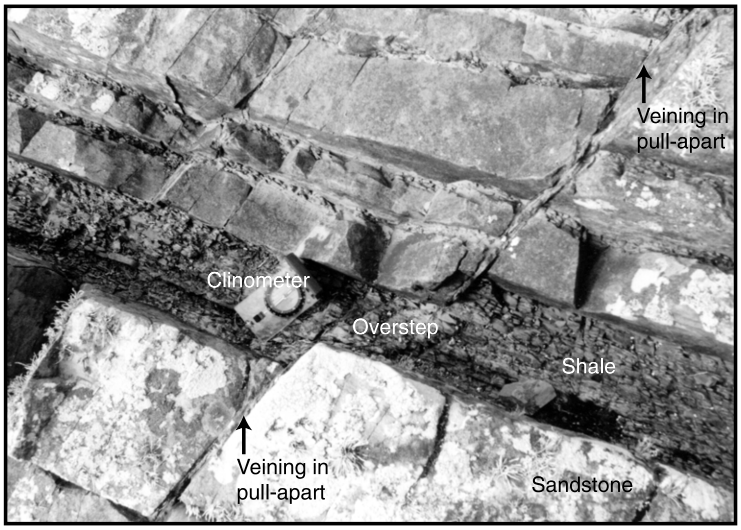


Fig. 10. Photograph of strike–slip faults cutting sub-vertical Silurian turbidites at Gipsy Point, Kirkcudbright, SW Scotland (horizontal surface; Peacock, 1991). Pull-aparts occur where a fault changed orientation through more brittle sandstone beds, with a contractional step accommodated by thinning of a shale bed. This fault appears to have initiated as veins in the more brittle sandstones (also see Figs. 4 and 5).

4.2. Scale variance of fault development

The mechanisms of fault initiation reviewed here give insights into the challenge some workers have faced in defining a power-law description of the relationship between the length of faults and cumulative slip across them (see the review by Bonnet et al., 2001). Indeed, it is evident from this review that the processes by which many faults initiate and lengthen, especially those that rely on pre-existing or precursory structures, are different from those processes that allow frictional slip across the resulting surfaces. Fossen and Hesthammer (1997) note a distinctly different scaling coefficient for deformation bands (precursory structures) than for the resulting faults nearby. Wilkins et al. (2001) observe that small faults that form from pre-existing joints have slip magnitude that is independent of length. Walsh et al. (2002) make similar observations for reactivated faults.

The role of lithology on the style of strain accumulation will necessarily influence the number of faults (or precursory structures). Steen and Andresen (1999) observe that for strain hardening mechanisms, such as deformation bands, there are more small structures than in rocks for which strain hardening is not observed. Thus, there is a different constant of proportionality between the number of faults and the slip across them for different lithologies (see also Wojtal, 1994). Bürgmann et al. (1994) show that changes in material stiffness along a fault can alter the slip distribution, and Gross et al. (1997b) illustrate mechanical contrasts in strata that influence slip–length scaling.

In addition, the mechanism by which faults grow changes with scale. For example, the propagation of faults by the linkage of en échelon veins or pressure solution seams is important at centimeter-scales, when fault slip is of the order of millimeters. At meter and larger scales, however, these processes appear to become relatively less important. When slip is of the order of meters and the faults are hundreds of meters long, fault propagation appears to become dominated by the linkage of segments. The change in mechanism necessarily has an influence on the slip–length scaling relationship (Mansfield and Cartwright, 1996). Wojtal (1994, 1996) shows a change in slip–length scaling relation with linkage of segments. Cartwright et al. (1995) and Mansfield and Cartwright (2001) illustrate how fault linkage can produce variability in the slip to length ratios of fault populations.

A slip–length ratio depends on the type of pre-existing or precursory structures, the lithology or lithologies, and the scale, number, and history of fault linkage. The initiation of faults from precursor structures limits slip–length scaling relations to dimensions larger than the initial structures. We suggest that the scale-dependant physical processes of faulting do not permit a universal slip–length scaling relation.

4.3. The importance of pre-existing and precursory structures

From the observations above, it is evident that most—if not all—natural faults initiate from pre-existing or precursor structures. In our examples, these premonitory structures are either brittle discontinuities (joints, veins, solution surfaces), or semi-brittle shear zones (deformation bands), or some combination (e.g. monocline and fissures). Of these, none are mode II fractures (‘shear fractures’). Deformation bands approximate mode II fractures, but they are not surfaces of displacement discontinuity. Similar observations are made in laboratory tests of rock (e.g. Horii and Nemat-Nasser, 1985), where microcracking precedes development of a through-going fault surface. While we cannot dispute reported observations of shear fractures, such as those observed in some rock failure tests (e.g. Paterson, 1978), it is apparent that shear fractures do not play an important role in the initiation of most faults in the upper crust (see also Petit and Barquins, 1988).

We suggest that all brittle faults in the upper crust faults initiate as mode I/−I fractures or as shear zones. For the case of semi-brittle shear zones, these necessarily involve fracturing and cataclasis. Thus, mode I fracturing controls the initiation of brittle faults.

5. Summary

The analysis of faults with millimeter- to meter-scale displacements from a range of rock types and tectonic settings has led to the identification of three distinct fault initiation styles: linkage of pre-existing structures, linkage of precursory structures, or initiation from a precursory shear zone.

Evolution of faults from the first two initiation styles follows a similar progression, regardless of the precursory or pre-existing structures involved. During stage 1, a fault utilizes pre-existing structures, or a set of precursor structures develops. Precursor structures tend to be mode I or −I fractures (veins, joints or pressure solution seams). This indicates that faults do not usually initiate as shear fractures. Stage 2 involves the development of linking structures (e.g. veins becoming linked by pressure solution seams), commonly in a locally perturbed stress field, allowing shear to occur across the zone. A through-going fault develops during stage 3, allowing shear to continue within the fault zone.

Precursor shear zones, such as deformation bands, appear to alter the strength of the host rock, thus inducing brittle failure.

Rock architecture, including layering, grain size, and cementation exert important controls on the style of fault initiation. In layered sedimentary sequences, the earliest structures appear to be mode I or −I fractures in the brittle units. The orientation of principal stresses with respect to layering controls the generation of the initial fractures.

Mode I fracture dominates fault initiation in the upper crust. Complex failure behavior occurs as the initial structures link to form a through-going fault. The various fault initiation and propagation styles described in this paper are scale variant, dominating only when displacements are at the scales of millimeters to meters. Propagation of larger faults tends to be dominated by the linkage of fault segments. This change in failure mechanism with scale should influence slip–length scaling relationships, and, very likely, our choices of failure criteria.

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