Recent advances in the understanding of fault zone internal structure: a review

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Abstract: It is increasingly apparent that faults are typically not discrete planes but zones of deformed rock with a complex internal structure and three-dimensional geometry. In the last decade this has led to renewed interest in the consequences of this complexity for modelling the impact of fault zones on fluid flow and mechanical behaviour of the Earth’s crust. A number of processes operate during the development of fault zones, both internally and in the surrounding host rock, which may encourage or inhibit continuing fault zone growth. The complexity of the evolution of a faulted system requires changes in the rheological properties of both the fault zone and the surrounding host rock volume, both of which impact on how the fault zone evolves with increasing displacement. Models of the permeability structure of fault zones emphasize the presence of two types of fault rock components: fractured conduits parallel to the fault and granular core zone barriers to flow. New data presented in this paper on porosity–permeability relationships of fault rocks during laboratory deformation tests support recently advancing concepts which have extended these models to show that poro-mechanical approaches (e.g., critical state soil mechanics, fracture dilatancy) may be applied to predict the fluid flow behaviour of complex fault zones during the active life of the fault. Predicting the three-dimensional heterogeneity of fault zone internal structure is important in the hydrocarbon industry for evaluating the retention capacity of faults in exploration contexts and the hydraulic behaviour in production contexts. Across-fault reservoir juxtaposition or non-juxtaposition, a key property in predicting retention or across-fault leakage, is strongly controlled by the three-dimensional complexity of the fault zone. Although algorithms such as shale gouge ratio greatly help predict capillary threshold pressures, quantification of the statistical variation in fault zone composition will allow estimations of uncertainty in fault retention capacity and hence prospect reserve estimations. Permeability structure in the fault zone is an important issue because bulk fluid flow rates through or along a fault zone are dependent on permeability variations, anisotropy and tortuosity of flow paths. A possible way forward is to compare numerical flow models using statistical variations of permeability in a complex fault zone in a given sandstone/shale context with field-scale estimates of fault zone permeability. Fault zone internal structure is equally important in understanding the seismogenic behaviour of faults. Both geometric and compositional complexities can control the nucleation, propagation and arrest of earthquakes. The presence and complex distribution of different fault zone materials of contrasting velocity-weakening and velocity-strengthening properties is an important factor in controlling earthquake nucleation and whether a fault slips seismogenically or creeps steadily, as illustrated by recent studies of the San Andreas Fault. A synthesis of laboratory experiments presented in this paper shows that fault zone materials which become stronger with increasing slip rate, typically then get weaker as slip rate continues to increase to seismogenic slip rates. Thus the probability that a nucleating rupture can propagate sufficiently to generate a large earthquake depends upon its success in propagating fast enough through these materials in order to give them the required velocity kick. This propagation success is hence controlled by the relative and absolute size distributions of velocity-weakening and velocity-strengthening rocks within the fault zone. Statistical characterisation of the distribution of such contrasting properties within complex fault zones may allow for better predictive models of rupture propagation in the future and provide an additional approach to earthquake size forecasting and early warnings.
Fault zones influence the mechanical properties and seismogenic behaviour of the crust, the migration and trapping of hydrocarbons and mineralizing fluids, regional hydrology and hydrogeology, and the morphology of the land surface (e.g., Handy et al. 2007). In particular, localization of shear within relatively narrow zones results in the formation of fault rocks, characterized by specific mechanical and hydrological properties dependent on a complex interplay between many factors. These fault rocks change during deformation, due to the operation of physico-chemical mechanisms related either directly or indirectly to deformation and often controlled by fluids. The resulting zone of fault rock is typically therefore highly heterogeneous, leading to large uncertainties in understanding and predicting the resulting mechanical and fluid flow behaviour of the fault. Nevertheless, the fact that faults are not discrete surfaces but zones, of finite width, of fault rock with different properties to surrounding host rocks has profound implications for the way in which we should assess the impact of faults on fluid migration and seismogenic behaviour in the Earth’s crust.

Fault zone architecture and related permeability structures form primary controls on fluid flow in upper-crustal, brittle fault zones. Compacting and dilatant regions of the fault zone will lead to the establishment of distinct structural and hydrogeologic units (e.g., Chester & Logan 1986). These units reflect the material properties and stress conditions within a fault zone, and dictate whether a fault zone will act as a conduit, barrier or combined conduit–barrier system (e.g., Caine et al. 1996). The permeability of fault zones is an aspect of particular interest in the fields of economic geology and seismogenesis, being a key factor in determining fluid pressure distributions and the volume of fluid that can pass through a fault zone. The temporal and spatial evolution of permeability in fault zones depends on the host rock type, stress and strain rate, deformation mechanisms, fault architecture, and many other parameters. Numerical modelling is a powerful tool to investigate the effect of varying permeability on the fluid flow patterns in terms of the stress regime, which can provide useful constraints for mineral exploration when coupled with reactive transport simulation (e.g., Zhang et al. 2008), and for understanding the role of faults within petroleum systems in both exploration and production contexts. Field and laboratory studies provide the input for such models in terms of both data and the appropriate fluid flow and mechanical rheological laws.

The seismogenic behaviour of a fault is also known to depend greatly on fault zone internal structure (Scholz 2002; Rice & Cocco 2007). At the large scale, the continued propagation or early arrest of ruptures depends largely upon the geometry of fault bends, dilational jogs and the connectivity or distance of step-over of individual strands in the fault system (King & Yielding 1984; Sibson 1985; Wesnousky 2006). Frictional properties of the fault materials, and their spatial variation, also control seismogenic behaviour and are key influences on whether the fault slips unstably or moves by steady slow creep. At smaller scales, more mature, evolved (higher displacement) fault zones are: (i) likely to have better developed smoother slip zones for localising rapid slip than rough slip planes of less-mature faults (e.g., Sagy et al. 2007); (ii) have larger overall fault zones of increasing complexity which may conversely encourage the diversion of the main rupture off the main slip zone onto a branch fault of different mechanical and permeability properties which may favour arrest (e.g., Shipton et al. 2006; Boutareaud et al. 2008a). These two competing factors may greatly impact on the frequency of the seismicity emitted, particularly in the early stages of earthquake slip, the repartition of the energy budget of the earthquake by different energy-sink mechanisms, and hence the operation and efficiency of various dynamic slip weakening mechanisms which govern the overall magnitude of the earthquake.

Experience shows that the complexity of a fault zone is strongly dependent on host rock lithology, displacement and pre-existing structure (including interactions with the mechanical layering of the host rocks). Depth (pressure and temperature) and stress regime and its evolution during the life of the fault can also play a role (e.g., Sibson 1977; Butler et al. 1995). Nevertheless, a single fault may also show strong changes in complexity along-strike or down-dip, even over relatively short distances (e.g., Childs et al. 1997; Schulz & Evans 1998). Despite such variation, careful documentation of the internal structure of a large number of faults, in terms of host lithology and displacement increase, is valuable in building up a general picture of the mechanisms of fault zone growth.

The purpose of this paper is to briefly critically review research into the heterogeneity of fault zone internal structure and fluid flow properties, how they may evolve during fault growth, and impact on improving our understanding and prediction of hydrocarbon migration and seismogenic behaviour of the crust. The emphasis is placed on recent and advancing concepts with suggestions for future work.

Fault zone evolution

Fault zone evolution has been examined for over two decades by comparing examples of fault
zones at different scales, the assumption being that small fault zones preserved in a population represent the early growth stages of the larger ones. Typically data on fault zone thickness, usually considered to be the thickness of highly deformed fault rock in which no original fabric is identifiable, have been compared for faults and shear zones – their deep crustal equivalents – of different displacement magnitudes in order to examine how a fault zone might develop with increasing offset (e.g., Robertson 1983; Wallace & Morris 1986; Scholz 1987; Hull 1988). Compilations of data over several orders of magnitude of displacement showed general linear trends of fault zone thickness increase with displacement (e.g., Fig. 1a), suggesting that continuous wear of the fault walls occurs as a function of increasing displacement (e.g., Scholz 1987), and that the fault therefore does not really change its rheology as it becomes larger. Numerous objections were raised to this kind of approach (e.g., Blenkinsop 1989; Evans 1990): definition of fault zone thickness is often difficult in highly heterogeneous zones with complex internal structure; comparing data over a range of scales necessarily requires log–log plots which mask the often large scatter; and datasets from different contexts (e.g., lithologies, tectonic settings, depths, etc.) were not always separated out, thus increasing the risk of irrelevant comparisons. Shipton et al. (2006) demonstrated the importance of separating out the fault zone ‘width’ into the thickness of different components such as fault zone core and damage zone (see Fig. 1a).

An increased awareness of these problems has led to more careful descriptions and data presentation/interpretation, such as describing faults in a single given field area and hence context, and separating out different components of the fault zone (e.g., Shipton et al. 2006). Nevertheless, the heterogeneity of fault zones, especially in cases where slip has localized onto several strands separating slivers of almost intact protolith means that interpretation of such data in terms of wear processes and fault mechanics should be done with caution. Furthermore, such an approach to examining fault zone evaluation depends on the assumption that the small faults observed in a given population are representative of the early stages of development of the large faults in the population.

Fig. 1. Compilations of fault zone thickness–displacement data over a wide range of scales, illustrating a general linear data trend, typical scatter in this trend, and examples of particular situations where data lie a long way off this trend. Note the log–log scales. (a) Synthesis of data for fault gouge zones in siliciclastic rocks (solid symbols) compared with measurements of the width of fault zones including zones of fracturing, and/or lens formation at linked segment relays (open symbols). Data from fault gouge zones are from Robertson (1983), Wallace & Morris (1986) and some unpublished field and microstructural data from the Lodeve Basin in S. France measured by the first author. A general envelope of these data is shown as a dashed oval. Data from the fractured zones are small-scale (thin section and experimental rupturing) fault tip-related rupturing from Otsuki (1978), Amitrano & Schmittbuhl (2002) and field measurements in granite from the Pelvoux Massif, western Alps by the first author (squares), lenses formed by relays at segment linkage zones (circles) (van der Zee et al. 2008) and larger-scale damage zones in sandstones (triangles) (Fossen & Hesthammer 2000). Data from sliding on discrete pre-cut surfaces are also shown for comparison (crosses), from Tallis & Weeks (1986), Yoshioka (1986), Blanpied et al. (1987) and Power et al. (1988). (b) Data from faults in high-porosity Cretaceous sands from S. France (crosses and open symbols) (Wibberley et al. 2007) and phyllosilicate-rich faults in crystalline basement of cataclastic origin (solid circles) (Wibberley 2005) as two examples of data exhibiting relatively inhibited fault zone growth in comparison to the general envelope of data in (a). Small-scale cataclastic deformation bands (crosses) show evidence of work-hardening in the field and do not lie significantly off a linear relationship. However, localized cataclastic faults (diamonds) and clay-rich larger faults (triangles) do not increase thickness as a linear function of increasing displacement.
Role of peripheral fracturing and host rock weakening

Emphasis in the last decade has been placed on the fault zone also often containing a zone of peripheral fracturing (often termed 'damage zone'; Caine et al. 1996) generated by a variety of mechanisms. Whether this is ignored or measured (if present) in a measurement of the thickness of the fault zone may have important impact on the interpretation of the data in terms of mechanical evolution of the fault zone during growth (Shipton et al. 2006).

Dilatant fractures formed during initial rupture of low-porosity rocks may result in a fracture zone around the initial rupture fracture right from the outset (e.g., McGrath & Davison 1995; Moore & Lockner 1995; Wibberley et al. 2000a). Faults in fracture-dilatant rocks such as crystalline rocks and low-porosity sandstones are thought to be governed largely by the interaction between the generation of peripheral fractures and the main fault zone. Dilatant fractures formed, such as mode I cracks and synthetic and antithetic Riedel microfaults (e.g., Tchalenko 1970; Brosch & Kurz 2008), result in weakening of the host rock adjacent to the developing fault, so that fault zone widening may be controlled as much by evolving host rock properties as by fault rock rheology (Fig. 2).

In natural fault zones, much of the geometric complexity can be related to wall rock fracturing and is therefore likely to be inherited from such processes occurring early on in the life of the fault. A wealth of theoretical and experimental data suggest that shear fractures form due to the interaction and coalescence of numerous tensile mode I microcracks (e.g., Paterson 1978; Lockner et al. 1992). Healy et al. (2006) have shown that the elastic stress fields developed around tensile microcracks in three dimensions lead inevitably to mutual interactions that promote the development of brittle shear fractures – and shear fracture intersections – that lie significantly oblique to all three principal stresses. This accounts well for the development of anastomosing and coalescing polymodal faulting patterns observed in many natural fault zones, even at low strains, in which fractures lie in a wide range of orientations at angles of up to 25° or more relative to the \( \sigma_1 \) and \( \sigma_2 \) axes. The importance of such peripheral wall rock fracturing may extend to controlling the entire rheological behaviour of the fault at the large scale by inducing stress rotations due to contrasting elastic moduli of such

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Fig. 2. Illustration of examples of different processes contributing to fault zone growth at different scales, and their effect on the evolution of fault zone growth as represented on a thickness–displacement plot. The thickness–displacement plot (a) summarizes data from Fig. 1a, with arrows suggesting possible evolution pathways as a result of different deformation processes. In the block diagram (b) the newly incorporated material in each process is indicated by dark shading, the pre-existing fault zone gouge denoted by diagonal shading. Each process first weakens the surrounding wall rock by fracturing (step i, open arrows on the graph), followed by incorporation of the fractured wall rock material into the fault zone during continued displacement (step ii, shaded arrows on the graph). The processes illustrated are tip-process zone microfracturing (1), segment linkage by relay breaching and lens formation (2) and splay-faulting and re-connection forming sidewall rip-outs (3). Note that each process increases the fault zone thickness above the general gouge trend during step (i), but the system strives to regain the general (wear?) trend as displacement continues during step (ii). The graph also attempts to illustrate the possible evolution of active slip bands or deformation zones (dotted arrows) during strain localization events, although the representation of this concept is dependent on the spatial and time-scales concerned.
zones with respect to the surrounding country rock (e.g., Faulkner et al. 2006).

The continued development of the fault zone as displacement increases may incorporate the fractured material into the developing fault zone as a function of the original width of the fractured zone. Eventually the material becomes sufficiently comminuted to form a zone of ultracataclasite, which, if forming a continuous zone, will localize further deformation and inhibit further widening of the fracture ‘damage’ zone (e.g., Micarelli et al. 2006). Hence such dilatant fractures on the small scale will diminish in impact as the fault zone increases in thickness. Indeed, continued deformation may result in the localization of strain and consequent narrowing of the active slip zone, whilst the overall finite fault zone thickness remains constant (Fig. 2a). Nevertheless, the generation of larger-scale splay faults (such as Riedel faults), fracturing around fault intersection zones, and cutting of asperities at a wider range of scales, will also result in bulk weakening of the host rock and favour incorporation of blocks of wall rock material into the fault zone as it gets larger. Examples of these processes are the incorporation of ‘sidewall rip-outs’ (Swanson 2005) and short-cuts of bedding-parallel slip asperities in the fault zone (Watterson et al. 1998), often classed as tip-line and asperity bifurcations respectively (Childs et al. 1996). In limestones, dilatant fracturing may be also important in the early stages of faulting, but the high rates of dissolution–precipitation processes are likely to govern later behaviour (e.g., Benedicto et al. 2008).

Influence of segment linkage

Studies of fault linkage, particularly in stratified rocks such as interbedded sandstones and shales or limestones and marls, suggest that if faults link at diastrophic jogs they may form relays containing lenses of relatively undeformed protolith (e.g., Peacock & Sanderson 1992; Childs et al. 1995; van der Zee et al. 2008; Micarelli & Benedicto 2008). Thus the thickness of the fault increases suddenly as two initially narrow faults link and form a wider relay zone (step i in Fig. 2a and b). Increased displacement will result in these relay zones evolving to a lens surrounded by fault gouges or other high strain fault rocks, without the overall width of the zone having significantly increased (step ii in Fig. 2a and b). Different fault zones, and indeed perhaps even different parts of the same fault zone, will have ceased activity at different stages of this relay evolution. Hence this model evolution predicts variations in displacement – thickness statistics. The thickness of mechanical layered units such as competent sandstones or limestones is thought to be an important scale control on this process (e.g., Wilkins & Gross 2002; Soliva & Benedicto 2005), and hence is likely to also be an important scale-related control on fault zone thickness resulting from this process. Thus a number of processes may operate, possibly repeating themselves at different scales, which together contribute to generating the structure of the fault zone. Furthermore, due to this potentially large number of processes contributing to fracturing of wall rock at different scales, the stepped evolution of the thickness–displacement trend is likely to be much more complicated than is illustrated in Figure 2. This is particularly so because a fault may cease its activity as a particular process, at any one scale, is in an early, mid or late stage of its evolution and/or repetition, and this may also vary along the length of the fault. Thus the superposition of a number of ‘stepped’ thickness–displacement patterns would naturally give a general linear trend with also the two- to three- orders of magnitude scatter that these data typically show. Nevertheless, the scale-dependence of some of these damage processes on features such as bed or mechanical layer thickness may give rise to similar scaling patterns for different faults and fault systems.

Effect of changes in fault rock rheology

In high-porosity sandstones the widespread presence of ‘deformation band’ faults has attracted much interest as these structures are thought to influence fluid migration in many good quality sandstone reservoirs (e.g., Aydin & Johnson 1983; Underhill & Woodcock 1987; Fowles & Burley 1994; Antonellini & Aydin 1994; Fossen et al. 2007). They typically appear as one or several white strands on the order of 1 mm wide clustered in zones a few millimetres to centimetres wide, accommodating displacements on the order of a few millimetres to tens of centimetres. These features have often been interpreted as being due to ‘work-hardening’, i.e. that the fault rock becomes stronger with displacement, so that the system has to wear the adjacent host rock to accommodate further displacement (e.g., Aydin & Johnson 1983; Underhill & Woodcock 1987). Abutting relationships of later deformation bands against earlier formed ones also suggest that the material in the deformation bands is stronger than that of the high-porosity host sandstone (Wibberley et al. 2000b). Studies of overprinting also suggest that deformation band networks also act to strengthen the system in the bulk sense. This may cause widening of the deformation zone initially (e.g., Johansen & Fossen 2008), but localization of deformation onto fewer, larger faults is eventually favoured during later deformation. These patterns of behaviour are probably a feature specific to high-porosity materials because they typically compact under most stress conditions, to
reduce their porosity as they shear, becoming stronger in the process. Most examples of deformation bands in the literature are cataclastic, with intragranular (hertzian) fracturing at grain–grain contacts resulting in the generation of angular fragments. The changing nature of the grains/fragments in the evolving deformation band may contribute to increasing fault strength as well as to the operation of shear-enhanced compaction of the high-porosity material. Yet studies of deformation band populations and larger (greater than metre-scale displacement) faults in high-porosity sandstones suggest that this work-hardening does not continue indefinitely. Larger faults show narrower thickness/displacement ratios (e.g., Fig. 1b) and evidence of deformation localization in slip planes as displacement continues (e.g., Antonellini & Aydin 1995; Shipton & Cowie 2001; Davatzes et al. 2005; Wibberley et al. 2007). Fault rock material in these cases is usually a ‘mature’ ultracataclasite, with a large proportion of very fine-grained fragmented and comminuted matrix supporting a small number of nearly intact original sand grains. Thus ‘large’ faults are very different in character and microstructural properties to small faults in high-porosity sandstones.

Integrated field, microstructural and geochemical evidence suggests that weakening of the fault rock, such as by syn-kinematic alteration, may encourage localization of the deformation onto a narrowing part of the active fault zone (reaction-enhanced ductility, White & Knipe 1978; e.g., Fig. 1b). This is particularly favoured in crystalline basement fault zones, where syn-kinematic alteration of feldspars to phyllosilicates in granitic upper/middle crust or olivine to serpentine in the upper mantle may encourage such localization of deformation (e.g., Janecke & Evans 1988; Wintsch et al. 1995; Imber et al. 1997; Stewart et al. 2000; Handy & Stüntitz 2002; Wibberley 2005; Jeffries et al. 2006). Nevertheless, care must be made in interpreting narrow fault zones in terms of absolute strength of the fault alone, because the localization of deformation is potentially related to the competence contrast between host rock and fault zone rather than in absolute changes in fault zone strength. However, the development of fine-grained phyllosilicate-rich fault rocks has been shown by both analogue experimental studies (e.g., Bos & Spiers 2002; Nieuwmeijer & Spiers 2005) and field studies of natural weak fault zones such as low-angle normal faults (e.g., Collettini & Holdsworth 2004) to promote the operation of ‘frictional-viscous’ mechanisms where grain-scale slip is accommodated by pressure solution allowing faulting at significantly reduced friction coefficients ($\mu$ as low as 0.2), i.e. absolute weakening (see Imber et al. 2008).

**Summary**

A number of processes operate during the development of fault zones, both in the developing fault zone itself and in the surrounding host rock, which may encourage or inhibit further fault growth. The complexity of the evolution of a faulted system requires changes in the rheological properties of both the fault zone and the surrounding host rock volume, both of which impact on how the fault zone evolves with increasing displacement (Fig. 2). Processes encouraging fault zones to grow include tip zone rupturing, formation of relay zones and splay faulting, all of which weaken the wall rock in a relative sense and allow incorporation of fragments or lenses of material into the fault zone. Processes inhibiting fault zone growth include work-hardening of the host rock volume before localization of a through-going fault, and strain-weakening in the fault zone itself, often influenced by syn-kinematic metamorphic reactions in the fault zone. Studies of ‘very large’ faults, i.e. those likely to cut the entire upper crust, show that the thicknesses of high strain zones are usually much narrower in relation to displacement than for small faults (e.g., Wibberley, 2005), implying that further wear by generation of splay and their incorporation into the zone diminishes once this scale of layering has been reached. Nevertheless, ‘large’ faults at this scale often occur in a set such as duplexes or strike-slip arrays. If the entire array is included in the definition of width of the ‘zone of faults’ (different to a ‘fault zone’ of near-exclusively high-strain fault rock), then the interpretation drawn may be somewhat different. Indeed, the application of ‘wear’ concepts to faults in the Earth’s crust seems to be a generalization of a wide range of mechanisms for weakening the surrounding host rock and incorporating it into the fault zone, most of which are scale dependent to some degree (Fig. 2), but when averaged out over a wide scale range, give an overall impression of scale-independence of fault zone growth processes.

Despite many studies into the internal structure of fault zones at different scales, evolution of the fault zone is usually inferred by comparison with smaller faults. Yet examination of the final structure can rarely distinguish whether deformation spread through time, or that the entire width was generated early on with later localization of deformation, or a combination of both (Means 1995). For large fault zones with complex internal structure and surrounding fracturing, a way ahead may be to look for methods of relative or absolute dating of the generation of structures around the high-strain part of the fault zone in order to resolve this question. This may be for example...
by careful observation of the evolution of peripheral deformation with respect to reliable time markers around faults. In the case of recent alluvial deposits around seismogenic faults from California, Ferrill et al. (2008) suggest that the fault damage zone width is established early on in fault zone evolution, with the active portion of the fault zone narrowing with further displacement. Interestingly, a similar pattern in the evolution of deformation distribution is proposed from integrated structural, microthermometric and stable isotope data from hectometric-displacement fault zones in limestone in the Gulf of Corinth, which suggest localization of deformation onto narrow slip zones after relatively early distributed brecciation (Benedicto et al. 2008).

Fault zone permeability structure

Introduction to concepts

Since the recognition became widespread that faults are zones of deformed material rather than discrete surfaces, the need became apparent to understand the physical properties of the fault zone, especially the permeability and its heterogeneity. Relatively early studies focused on the low-permeability nature of clay-rich fault gouges from the centre of large strike–slip faults such as the San Andreas Fault (Chu et al. 1981; Morrow et al. 1984) and an exhumed ancient branch, the Punchbowl fault (Chester & Logan 1986). The latter case also documented high-permeability zones around the fault associated with the presence of peripheral fractures around the central zone. In siliciclastic sedimentary settings, focus was placed on the moderate permeability reductions of deformation bands and cataclastic fault zones in high-porosity sandstones and the more significant reductions with increasing clay-to-sandstone proportion in the fault zone (e.g., Antonellini & Aydin 1994; Gibson 1994). Through the need to generalize such variability and fault zone complexity, a simple framework emerged for fault zone permeability structure by considering the presence or absence of two main elements modifying host rock properties: a low-permeability fault zone core, providing a barrier to across-fault flow, and a high-permeability fractured damage zone around the central core, providing a conduit or drain for along-fault flow (Caine et al. 1996). Whether or not each of these elements is present, and how important they are, depends upon various factors such as host lithology and displacement, which help to understand whether a given fault can act as a lateral barrier to fluid flow, a fault-parallel conduit, or both.

More complex permeability structure models

Several studies have shown that the simple core zone/damage zone model does not always adequately represent fault zone complexity, particularly the heterogeneous nature of the core zone, despite the overall usefulness of the concept. Core zone complexity becomes particularly important, for example, when considering the core zone behaviour during the onset of earthquake dynamic slip, particularly in the low-permeability slip zones governing dynamic slip weakening by thermal pressurization (Wibberley & Shimamoto 2005) and their branching (Boutareaud et al. 2008) or the distribution of velocity-strengthening and velocity-weakening fault rocks in the core (Faulkner et al. 2008). Two contrasting models of detailed core zone internal structure and permeability heterogeneity were based on very different lithosphere-scale strike-slip faults. The model of Faulkner et al. (2003), based on the Carboneras fault in southeast Spain, suggests that very wide (on the order of 1 km) fault gouge zones of low-permeability may contain large lenses of high-permeability fractured (‘damage zone’) host rock. Thus, while fluid escape from the fault zone is hindered by the low-permeability nature of the clay gouge perpendicular to foliation (Faulkner & Rutter 2001), the lateral (e.g., up-fault) permeability could be much higher given sufficient connectivity of the fractured host rock lenses. Wibberley & Shimamoto (2003) use a study from the much narrower Median Tectonic Line in Japan to show how the juxtaposition of host rocks of contrasting mechanical properties results both in an asymmetric permeability structure and in the intense localization of deformation into a narrow central slip zone (several millimetres to centimetres wide), probably generated by episodes of rapid slip, at the boundary between the fault rocks derived from the host rocks either side of the fault. The intense grain size reduction and very finely foliated clay-rich nature of this slip zone gouge results in it having the lowest permeability of all the fault rocks in this complex large scale fault zone.

Permeability structure and behaviour of active faults

The above studies on fault zone internal structure were carried out from field studies complemented by laboratory permeability measurements not subjected to deformation apart from isotropic confinement (inducing compaction). Seront et al. (1998) measured the permeability, during deformation
experiments, of fracture-dilatant cataclasites and breccias sampled from the core of a seismogenic active fault, and showed how their complex arrangement in the fault zone core may result in the irregular distribution of dilatancy-hardening and contribute to local fluid storage and rupture arrest during an earthquake, similar to fault jogs. The heterogeneous distribution in the fault core of fracture dilatant fault rocks (typically cemented cataclasites) and fault gouges which deform by shear-enhanced compaction is significant in controlling permeability changes in the fault zone during active faulting (Uehara & Shimamoto 2004), and can determine whether or not up-fault leakage occurs. During rapid slip, the permeability behaviour – which can also feedback to the dynamic slip behaviour – is therefore governed by the propagation path of the through-going rupture in this heterogeneous fault zone.

Field observations of mineralization localized in fault breccias and veins around high-angle reverse faults in mid-crustal crystalline basement motivated Sibson (1990, 1992) to propose a fault-valve model of transient fault zone permeability. This model necessitates a large vertical pressure gradient supported by a low-permeability mid-crustal seal in the fault, periodically broken by earthquake slip to allow pressure re-equilibration by fast up-fault fluid flow. The rapidly decreasing temperature of the mineralising fluid induces precipitation, thereby sealing the transiently high permeability of the fault until the next slip cycle. This model assumes dilatancy and permeability increase by fracturing during rapid slip, likely to be the case in crystalline basement. As argued by Sibson (1995), such a fault valve behaviour is more likely to be valid in the case of reverse reactivation of high-angle faults (e.g., Cox 1995) than normal faulting, although case studies in normal fault zones show similar features to those predicted to be caused by fault-valve behaviour such as evidence for the seismic release of high-pressure fluid in the fault zone followed by interseismic sealing by mineralization (e.g., Bruhn et al. 1994).

Another approach to considering/modelling permeability of active faults is to draw on the assertion that fault gouges are fluid-saturated granular media of a given porosity. Despite the increasing wealth of permeability data from such fault rocks, truly coupled porosity–permeability data of fault gouges are rare because of the technical challenges of measuring both properties at the same time, particularly under a range of pressure and anisotropic stress conditions. Nevertheless, such porosity–permeability relationships are potentially extremely useful, because the framework of critical state soil mechanics provides a quantitative link between porosity changes and the evolution of differential stress and effective pressure for granular materials, and have already been used to explain the mechanical and porosity evolution of deformation bands and cataclastic fault zones in high porosity sandstones (Sheldon et al. 2006; Wibberley et al. 2007). Combined with porosity–permeability relationships, critical state soil mechanics should provide a new framework in predicting permeability changes during the evolution of fault activity and stress state in active fault zones. The validity of this approach needs to be evaluated with appropriate porosity–permeability data for fault rocks under a range of effective mean stress and differential stress conditions.

As an illustration, a previously unpublished series of experiments on a suite of phyllosilicate-rich (muscovite/ilite and chlorite) gouges of different quartz clast content and grain size (collected from the Median Tectonic Line, Japan; Wibberley & Shimamoto 2003) were performed to evaluate porosity–permeability relationships of these fault rocks and their variation with mean effective stress and differential stress, using nitrogen gas as a pore fluid at a pore pressure of 50 MPa.

**Fig. 3.** Permeability v. porosity of a suite of clay-rich granular fault gouges from the Median Tectonic Line, measured under isotropic stress conditions with nitrogen gas as a pore fluid (pore pressure 50 MPa) with confining pressure ranging from 80 to 200 MPa. CWG = coarse white (quartzofeldspathic) gouge, CFG = coarse foliated gouge, FG = foliated gouge, FFG = fine foliated gouge, SZ = narrow central slip zone gouge (two samples, denoted SZA and SZE). All permeability measurements are parallel to foliation, obtained using the pore pressure oscillation method. Porosity was measured from an initial pore volume measurement, followed by pore volume change measurements after each change in confining pressure. General trends in the data, as described in the text, are illustrated by the numbered dashed arrows.
Figure 3 shows the porosity–permeability relationships for the suite of gouge samples without axial loading, i.e., confining pressure changes only. The suite of samples shows porosities ranging from 10% down to 4% for the coarsest to the finest gouges, and permeabilities from $10^{-16}$ down to around $10^{-21} \text{m}^2$ parallel to foliation. These results illustrate three pertinent findings:

1. Porosity and permeability diminish with increasing confining pressure during compaction with an approximately log-linear relationship (Fig. 3, trend 1).

2. The deconfining parts of the experiments, and any later pressure cycling not exceeding the maximum previous confining pressure, systematically show a different porosity–permeability relationship to the initial confining phase (Fig. 3, trends 2a and b respectively).

3. Porosity and permeability decrease with decreasing grain size working in towards the centre of the fault zone (Fig. 3, trend 3), yet the different samples have similar gradients of the porosity–permeability relationship.

These data show that, whilst porosity varies by a factor of 2–2.5, permeability varies by 4–5 orders of magnitude. Permanent compaction and later elastic dec ompaction have different effects on the relationship of the pore volume (porosity) and pore connectivity/tortuosity of flow paths (permeability).

Figure 4a shows porosity and permeability data for a foliated quartzo-feldspathic gouge sample first subjected to two confining pressure cycles.
(i.e., no axial deformation, isotropic stress), followed by axial shortening by a constant piston displacement rate at constant confining pressure during which axial stress was measured. In the first case the mean effective stress \( P = (\sigma_1' + 2\sigma_3')/3 \) is simply the difference between confining pressure \((P_c)\) and pore pressure \((P_p)\), where \(\sigma_1' = \sigma_3' = (P_c - \alpha P_p)\) and assuming that the coefficient \(\alpha\) equals 1. In the second case there is a significant deviatoric component to the stress regime. The data show that:

1. Both porosity and permeability have log-linear relationships with mean effective stress during axial deformation, at least while deformation remains relatively homogeneously distributed throughout the samples.
2. For a given mean effective stress, both porosity and permeability are significantly lower if there is a differential component to the stress regime.

The porosity kick (Fig. 4a i and iii) may be due to spaces being created at the slip zone – jacket interface during localized slip. This phenomenon was apparently never severe enough to short-cut flow and affect the sample bulk permeability measurement. Figure 4b shows that the permeability of a clay-rich foliated gouge sample has a log-linear dependence on mean effective stress during axial deformation (at constant confining pressure).

Care must be taken when applying these findings to foliated gouge behaviour in natural fault zone systems because the result depends largely on foliation orientation in the experiment. First, permeability is anisotropic, so porosity–permeability relationships will be different for permeability in different directions with respect to the foliation. Second, the mechanical response of the pore spaces, folia and flow pathways to axial loading may be different if the foliation is perpendicular to piston displacement direction or parallel to it. Healthy scepticism is also required for quantitative application of these data in modelling – both the role of pore fluid in the experiments (nitrogen gas rather than water) and the effect of barrelling of the samples during deformation need to be properly considered. Nevertheless, these findings show that fault gouge behaviour is consistent with behaviour predicted by critical state soil mechanics and that, with future carefully planned experiments, fault gouge permeability may also be predictable within this framework.

**Summary of advancing concepts: active fault permeability structure**

The original concept of a two-component fault zone system affecting host rock properties, the fault core and surrounding damage zone model has more recently been modified for large, potentially seismogenic or creeping lithosphere-scale fault zones with complex fault core structures. To understand and predict permeability behaviour during active deformation, a knowledge of the fault rocks and their response to (anisotropic) stress changes and deformation is critical. The basic two-component fault rock model is improved upon by using the above results on fault gouge, coupled with previous experiments on deformation of cemented cataclasites. Although the more recent models for permeability structure suggest that lenses of fractured fault rock with the low-permeability gouge core, or an asymmetric fractured cataclasite–gouge distribution with intense slip localization at the boundary, the same premise nevertheless runs that two fault rocks are present of contrasting poro-mechanical behaviours (e.g., Agosta 2008), namely:

1. a low-permeability granular material which can be modelled by critical state soil mechanics (possibly coupled with pressure solution);
2. a generally higher-permeability fracture-dilatant material which can be modelled by elastic-fracture permeability laws coupled if necessary with the effect of cyclic cementation.

Thus faults which contain predominantly clay gouge cores are likely to behave in a compaction-creep manner during steady low-deformation rate fault activity. During rapid slip they may nevertheless act as temporary conduits simply due to the irregular large-scale slip surface or slip zone geometry inducing dilatancy, although this is still an open question. Indeed, the three-dimensional structure of fault zones must be borne in mind when considering all the models described in this section, particularly as fracture–fault intersection zones and dilational jogs often provide localized vertical conduits for fluid flow or infiltration if sufficient pressure gradients are present (e.g., Sibson 1996; Baietto et al. 2008; Micarelli & Benedicto 2008).

Faults containing sufficient lenses of fracture-dilatant cataclasites may drain fluid more easily up the fault during deformation depending on whether or not these lenses are sufficiently connected. Poor connectivity between lenses may result in dilatancy hardening in the fault zone (e.g., Seront et al. 1998), depending on the rate of dilatancy with respect to fracture connectivity. Faults containing predominantly fracture-dilatant cataclasites in their cores are likely to behave in a mechanically more intermittent manner with fracture dilatancy inducing high fluid flow rates, cyclically sealed by cementation. These concepts may provide the basis for evaluating the dynamic slip v. continuous creep behaviour of faults independently
of (or in addition to) the velocity-strengthening/velocity-weakening behaviour which is currently only studied as a function of mineral composition by laboratory rotary shear experiments.

**Impact on hydrocarbon sealing and migration**

Fault zone internal structure has major implications in hydrocarbon exploration and production. Faults frequently form side-seals to petroleum traps, as well as providing conduits for fluid flow from deeper to shallow levels in basins. During hydrocarbon production, faults often act as barriers or baffles to fluid flow along the reservoir layers. Understanding these behaviours requires knowledge of fault-zone structure in the subsurface at an appropriate scale – a challenging task for the petroleum geoscientist.

A fault can provide a long-term seal (on geological time-scales) to hydrocarbon movement if the following conditions are met:

1. the reservoir rocks are juxtaposed against sealing lithologies across the fault, by virtue of the fault displacement;
2. or, reservoir–reservoir juxtapositions at the fault-zone are characterized by sealing fault-rock with high capillary threshold pressure;
3. and, the stress conditions on the fault do not promote flow up the fault plane.

**Geometrical implications**

To determine the first of these conditions requires a detailed mapping of the fault network and the sequence which it cuts. Typically such mapping is based primarily on the interpretation of 3D seismic reflection data. From the layer and fault interpretation, ‘Allan diagrams’ (named after Urban Allan who popularized these diagrams; Allan 1989) can be constructed. These are displays of the fault plane(s) showing the juxtaposition relationships between reservoir layers on the two sides of the fault. Allan diagrams provide the first-order control on the potential connection topology (plumbing) across faults. In simple circumstances, it is assumed that reservoirs juxtaposed against sealing lithologies will experience a side-seal at the fault, whereas reservoir–reservoir juxtapositions have the potential to provide a fluid pathway across the fault. Interpretational and structural integrity is critical to the creation of accurate Allan diagrams, and industry-standard mapping practice is often inadequate (e.g., auto-tracked horizons, no fault interpretation).

Detailed fault-zone structure also adds uncertainty to Allan diagrams because these details (<10 m size) are below the resolution of the seismic reflection technique (Hesthammer & Fossen 2000). They may include multiple fault planes within the fault zone, and local deformation (ductile or brittle) of the wall-rocks. A common occurrence (Childs et al. 1996, 1997) is for two or more separate slip planes to be present, sharing the total displacement that is seen on seismic data, with an intact sliver of rock between them (Fig. 5a). In such cases, individual reservoir layers which appear offset in the seismic image have the risk of being self-connected if their thickness is more than half of the total throw. Three-dimensional studies of fault zone internal structure (e.g., Childs et al. 1996) suggest that such a system of separate slip surfaces can be surprisingly variable along strike and up and down the fault (Fig. 5b). Hence such complexity is extremely difficult to predict, although it is likely that such heterogeneity is increased by the presence of strong mechanical layering in the host rock and tilting during deformation (e.g., van der Zee et al. 2008). A common wall-rock deformation is a small-scale (<100 m) ‘drag’ fold in one or both walls of the fault – in extreme cases, such folds may accommodate up to 90% of the total offset (e.g., Fossen & Hesthammer 1998).

**Impact of fault rock composition**

The sliding of wall rocks past each other creates new rock types within the fault zone, potentially having completely different hydraulic properties from the juxtaposed formations. The nature of fault rock depends on three key factors: the composition of faulted sequence, the stress conditions at the time of faulting, and the post-faulting burial history (especially temperature); (Antonellini & Aydin 1994; Fisher & Knipe 2001; Fulljames et al. 1997; Gibson 1998; Sperrevik et al. 2002). In clay-rich sequences (typically >40% clay beds), clay smears are a common component of the fault zone. They are characterized by a tapering wedge geometry, with clay being sheared into the fault zone from upthrown and downthrown halves of an offset clay layer (Weber et al. 1978; Aydin & Eyal 2002; Takahashi 2003; Eichhubl et al. 2005; van der Zee & Urai 2005). Thicker clay layers tend to be particularly effective at contributing clay into the fault zone, and examples observed at outcrop include smears up to 1 m thick continuous over 70 × 400 m of fault plane (Lehner & Pilaar 1997). In impure sands (15–40% clay content), faulting creates a ‘clay gouge’ (or shaly gouge), also known as a phyllosilicate-framework fault-rock (PFFR) (Gibson 1998; Fisher & Knipe 2001). PFFRs exhibit a deformation-induced mixing of clay and sand grains, often with a fabric
aligned parallel to the fault plane. Faults cutting clean sands (<15–20% clay) show the greatest variety of fault rocks, dependent on the stress-temperature history (Fossen et al. 2007). Faulting of clean sand near to the surface results only in a local grain rearrangement, to produce a rock type called a disaggregation zone which has similar hydraulic properties to the host sandstone (Fisher & Knipe 2001). However, faulting the same sand at higher stresses (e.g., >1 km burial depth) results in grain fracturing (cataclasis), with the smaller grain fragments clogging the pore-space (Fowles & Burley 1994; Crawford 1998). With subsequent post-faulting burial, both disaggregation zones and cataclasites (but especially the latter) can be further modified by quartz dissolution and local reprecipitation – this thermally activated process becomes significant deeper than c. 3 km (for an average geothermal gradient), (Fisher et al. 2003). Severe quartz overgrowths can convert a cataclastic

Fig. 5. The problem of resolution from seismically imaged faults and impact on across-fault reservoir connectivity and possible fluid flow pathways. (a) Fault zones more complicated than a single slip plane will have reservoir–reservoir connectivities and the potential for fluid migration (bold dashed arrows), which are difficult to predict, as a function of multiple slip planes and/or normal drag. (b) An example of along-strike changes in fault zone structure (from Childs et al. 1996).
deformation band to something resembling a ‘sheet of glass’ in the reservoir.

For fault seal prediction, the huge variety of fault-rocks, described above, may seem daunting. Which fault-rocks are relevant to a particular hydrocarbon prospect? A widely used methodology in the petroleum industry uses the Shale Gouge Ratio algorithm (SGR) to map out the main areas of different fault-rocks automatically on the Allan diagram. SGR is computed from the net clay content of the section sliding past each point on the fault, as this represents the material that can potentially be entrained into the fault zone (Yielding et al. 1997; Yielding 2002). Under conditions of perfect mixing, SGR would represent the upscaled composition of the fault rock, but in reality it does not represent the detailed internal structure (e.g., Fig. 6), where many different components (e.g., smears, cataclasites) might be present together (e.g., Foxford et al. 1998; van der Zee et al. 2008). Nevertheless, recent stochastic modelling of that detailed structure implies that at higher ratios of throw-to-bed thickness the SGR is an effective and pragmatic approximation (Childs et al. 2007).

In exploration contexts, fault hydrocarbon retention capacity is then usually evaluated by considering the capillary threshold pressure of the fault rock likely for any SGR value, supposing that the fault zone is saturated in water. In order to perform such an evaluation, a relationship between SGR and threshold pressure needs to be calibrated. SGR values have been calibrated using two different approaches:

1. empirical comparison with in-situ hydrocarbon and pre-production pore-pressure data at proven fault traps (Yielding 2002; Bretan et al. 2003; Bretan & Yielding 2005); and
2. analogy with small-scale samples recovered from simple fault zones (Sperrevik et al. 2002).

In the empirical approach, it is observed that an SGR of around 15–20% often represents a threshold for a leak-seal transition (e.g., Yielding et al. 1997; Gibson & Bentham 2003). This is consistent with fault-rock samples because it corresponds to the distinction between clay-poor fault rocks, such as disaggregation zones, and more clay-rich rocks such as PFFRs and smears. On a more quantitative basis, the SGR values can be calibrated by the in situ hydrocarbon buoyancy pressures in observed fault-bound traps, to indicate the strength of the fault seal in terms of the trapped column height (Bretan & Yielding 2005). Similar calibrations can be performed using fault-rock samples, but lab-derived Hg–air threshold pressures must then be converted to hydrocarbon-water values using uncertain fluid properties (O’Connor 2000; Nordgård Bolås et al. 2005). In a field development context, the relevant physical property is the fault-rock permeability, upscaled to the size of the cell–cell connections in a geocellular simulation model. The fault-rock property is typically input in terms of modifiers to the cell–cell transmissibilities across the fault – the transmissibility is simply the mean permeability of the juxtaposed cells, weighted by their cross-sectional contact area and the inverse of the distance between their centres (Manzocchi et al. 1999).

The impact of heterogeneity in structure and composition

The three-dimensional heterogeneity of fault zone internal structure becomes important in predicting hydrocarbon flow in a number of ways. Fault zone capillary threshold pressures vary down the fault plane as a function of juxtaposed stratigraphy (Yielding et al. 1997). A sandstone reservoir unit juxtaposed against a sand–shale cover series will have a higher SGR working up the fault plane to the top of the reservoir because a higher percentage of the juxtaposition history will have involved the more shaley cover. Thus capillary threshold pressure will increase up the fault plane accordingly, and such a threshold pressure profile will have to be compared with the buoyancy pressure profile of a possible hydrocarbon column to predict the leak point and hydrocarbon column height retained against the fault (Brown 2003; Bretan & Yielding 2005). Along-strike variations in throw will strongly alter reservoir–reservoir
juxtaposition connections as imaged on Allan diagrams, where reservoir self-juxtaposition becomes more important approaching fault tips and low-displacement segment connections, but also because the branching nature of fault strands can change drastically along strike as described above, strongly influencing reservoir juxtapositions (Hesthammer & Fossen 2000). In the third dimension, across the fault zone, heterogeneity in composition will result in a wide range of fault rock capillary threshold pressures and permeabilities even at the same ‘point’ on a two-dimensional Allan diagram. Thus treating a three-dimensional fault zone as a two-dimensional surface necessarily adds uncertainty by not taking into account such heterogeneity. For threshold pressures, the threshold pressure of the fault rock directly adjacent to the reservoir wall rock will initially control the hydrocarbon infiltration into the fault zone, followed by the geometry of percolating pathways within the internal structure of the fault zone. Nevertheless, various fault seal prediction studies have attested to the robustness of the SGR approach (at least in the hydrostatic regime), despite this uncertainty, suggesting that this third dimension heterogeneity somehow gets averaged out over the large scale (Yielding 2002; Childs et al. 2007).

For permeability, various algorithms have been proposed to predict fault rock permeability with respect to percentage clay content (e.g., Manzocchi et al. 1999; Sperrevik et al. 2002). These vary widely between each other and are mostly calibrated from samples taken from small (centimetre-offset) faults, although these may not have the same fabrics as more ‘mature’ faults. A heterogeneous composition across the fault zone will result in across-fault flow rates which are much harder to predict than for a single composition in a homogeneous fault zone. Here the large-scale ‘bulk’ permeability is unlikely to be the thickness-weighted average of the individual fault rock permeabilities in the fault zone, particularly when permeability anisotropy and tortuosity come into effect. One suggested way forward is to use stochastic flow simulations (e.g., Lunn et al. 2008) to model flow in heterogeneous fault zones composed of clay contents with a defined statistical variation both across and along the fault zone, coupled with a permeability-clay content algorithm taking into account permeability anisotropy, and compare the results with single-component systems. A second parallel route forward is to ‘calibrate’ fault zone permeability behaviour with large (e.g., field) scale examples where fault permeability has been estimated, such as from field tests like tracers or four-dimensional seismic, or production history matching, for example (see Jolley et al. 2007 for a comprehensive overview of current best practice).

**Impact of stress state**

Whatever the ability of the fault zone to act as a seal to across-fault flow, fluid flow within the fault zone may be promoted by the *in situ* stress state. A fundamental criterion for such flow is that the fault be critically stressed and close to frictional failure (Barton et al. 1995). Flow probably occurs in arrays of dilatant microfractures which are associated with the slipping (or near-slipping) fault surface (e.g., Losh et al. 1999). This behaviour was initially quantified by studies of fractured basin rocks, but has now been observed in seismic reflection studies of an actively slipping growth fault in the Gulf of Mexico, where a large pulse of overpressured fluid appears to be moving up the fault zone at >100 m per year (Haney et al. 2005). For predictions of whether such flow may occur, it is necessary to know the *in situ* stress state and pore-pressure, and the approximate strength of the fault zone (assumed to be weaker than the surrounding rock, but see Dewhurst & Jones 2003). Different parts of the fault network feel a different imposed stress according to their orientation in the stress field. The nearness to failure of a particular fault plane is assessed by parameters such as ‘slip tendency’ (ratio of shear stress to normal stress, Morris et al. 1996) and ‘fracture stability’ (a measure of the pore-pressure increase that would induce slip, Wiprut & Zoback 2000; Mildren et al. 2005). Such parameters may help in ranking different fault trends for potential seal breach, ranking different faults for stability during drilling, and understanding migration pathways from deep to shallow parts of the section (Mildren et al. 2002).

**Implications for seismogenic processes**

Earthquakes are the result of ruptures that nucleate, grow and terminate along, in most cases, pre-existing faults (Gilbert 1884; Koto 1893; Scholz 2002). Earthquake dynamics is concerned with both the behaviour of the volume around the fault, its accumulation of elastic strain energy and catastrophic release thereof (e.g., Reid 1910). The behaviour of the fault itself is a contact problem in which understanding fault friction (e.g., Scholz 1998) and shear fracture development (Ohnaka 2003) are key features. The recognition that most faults are in reality zones of complex and unpredictable internal structure and composition in which the rupture nucleates and through which it propagates (e.g., Sibson 1983), has deep-reaching implications for
our understanding of earthquake-related processes and how we model them. In this section we will briefly review some recent results of the role of rheology of fault materials and of geometry of faults in controlling earthquake nucleation, propagation and arrest.

**Complexity of seismic fault zone structure**

As the contributions to this Special Publication illustrate, faults are often very heterogeneous zones with respect to their structure and physical properties. In the case of strike-slip faults, for example, field investigation of fault zones exhumed from different depths in the crust reveal different architectures that vary with host lithology and grade of maturity of the fault (displacement, exhumation history and reactivation for example). In the case of shallow crustal deformation, the Punchbowl fault, exhumed from 2–5 km depth in the San Andreas Fault system, accommodated about 30 km of slip on a metre-scale wide cataclasite fault core surrounded by a fracture damage zone several hundreds of metres wide (e.g., Chester et al. 1993; Chester & Chester 1998). In this case it has been shown that most of the deformation is localized on individual 100 µm to centimetre-scale thick ultracataclasite slipping zones, so that only a small proportion of the highly deformed fault core was thought to have accommodated the most recent rapid slip events. Larger-displacement fault zones from similar depth, such as the main active strands of the San Andreas fault intersected at 3 km depth by SAFOD drilling (San Andreas Fault Observatory at Depth) project, have several creeping sub-parallel fault cores, cutting a wider, up to 1 km thick, damage zone (SAFOD phase III preliminary results, Hickman 2007). A spectacular example of the complexity of a mature plate-boundary fault zone exhumed from 1.5–4 km depth is the Carboneras Fault in Spain (Faulkner et al. 2003). The fault zone, which accommodates up to 40 km of displacement, consists of a 1 km thick fault zone composed of continuous and anastomosing strands of phyllosilicate-rich fault gouge bounding lenses of dolomite hundreds of meters long. The wide nature of the fault gouge and lack of evidence for localization of rapid slip deformation within the structure suggests that the fault operated as a stably creeping plate boundary fault (Faulkner et al. 2003, 2008).

With increasing depth, fault zone internal structure and the range of fault rocks vary with the degree of maturity of the fault (e.g., Holdsworth et al. 2001). For instance, long-lived active faults exhumed from about 10–15 km depth that accommodated hundreds of kilometres of slip, like the Median Tectonic Line in Japan, have complex fault cores, with mylonites and foliated cataclasites (representative of the deeper activity of the fault, see Jefferies et al. 2006) cut by phyllosilicate-rich foliated gouges and clay-rich cores produced by continued fault activity during exhumation to shallower levels (Wibberley & Shimamoto 2003). Less mature fault zones exhumed from similar depths (9–11 km), but with cumulated displacements of several kilometres at most, may consist of hundreds of centimetre-thick cataclasite–mylonite zones cutting a poorly developed damage zone (e.g., Gole Larghe Fault Zone, Italy, Di Toro & Pennacchioni 2005; Fort Foster Brittle Zone, Maine, Swanson 1988; Ikertorq Brittle Fault Zone, Greenland, Grocott 1981). In this latter case, the fault zone hosts abundant pseudotachylytes (usually less than 1 cm in thickness) within the fault rocks (a marker of ancient seismic activity, e.g., Cowan 1999), which are not found in the other faults. Thus for these examples of faults active at greater depths in the Earth’s upper crust, the range of fault rocks and complexity of their distribution increases with fault displacement and degree of exhumation during activity. Though these field studies give a flavour of the complexity of natural fault zones as well as how difficult it is to determine and infer fault structure at depth, recent experimental and numerical modelling studies, based on field observations and seismological data, have helped greatly to understand the role of heterogeneity and complexity in fault zone internal structure on the seismogenic behaviour of fault zones.

**Rupture nucleation**

The nucleation of crustal earthquakes is generally thought to be the consequence of frictional instabilities along fault surfaces (Brace & Byerlee 1966) similar to stick–slip phenomena observed during sliding of metals (Bowden & Tabor 1966). Although friction is the result of geometrical, chemical and atomic interactions (Coulomb 1785; Bowden & Tabor 1950; Rabinowicz 1965; Persson 2000; Gerde & Marder 2001; Urbakh et al. 2004), here we only consider rock friction as the result of the interaction of geometrical asperities or protuberance of the sliding surfaces; the importance of fault healing (e.g., Muhuri et al. 2004) being beyond the scope of this review. Roughness is typical of all fault surfaces over a wide scale range (Power et al. 1988; Power & Tullis 1992; Renard et al. 2006; Sagy et al. 2007). Fault irregularities may produce detectable pre-earthquake stress heterogeneities. The distribution of static stress drops during earthquakes suggests that fault surfaces are irregular and only
accumulate stresses over a small fraction of their area at a limited number of asperities, emphasizing the importance of the asperities in the rupture process (Bouchon 1997; Sammis et al. 1999; Fletcher & McGarr 2006).

Seismic ruptures can only nucleate in velocity-weakening rocks (see Scholz 2002 for a discussion). However, not all rocks are velocity-weakening for slip rates lower than 0.001 m s$^{-1}$ (velocities considered to be appropriate to the nucleation zone of an earthquake). Fault gouges containing abundant clay minerals such as montmorillonite, chlorite and illite are consistently velocity strengthening (Marone et al. 2008), whereas smectite may show a complex dependence with slip rate, normal stress and temperature (e.g., Saffer et al. 2001). For $T < 200 \degree C$, brucite, chrysotile- or antigorite-serpentinite gouges are velocity-strengthening or velocity-independent under drained conditions (e.g., Moore et al. 1996, 2001), thereby inhibiting dynamic instability and promoting stable creep (Moore et al. 2004). The direct extrapolation of these experimental observations to natural conditions implies that a heterogeneous distribution of different velocity-weakening and velocity-strengthening fault materials is one of the factors controlling earthquake nucleation (Marone & Scholz 1988). For example, in the case of the complex Carboneras Fault zone structure, this extrapolation predicts that earthquakes would nucleate in the dolomite patches (velocity-weakening) and arrest in the continuous clay-rich layers (velocity-strengthening) (Faulkner et al. 2003, 2008). The preliminary examination of the samples recovered from cuttings of the SAFOD borehole, located close to the northwestern end of the 2004 M5 Parkfield earthquake rupture zone, shows the presence of velocity-strengthening materials such as talc (derived from host rock serpentine by alteration in the fault zone). Although further work is needed to better quantify how much talc is present, this could provide an explanation for the lack of large earthquakes on this portion of the San Andreas fault (Moore & Rymer 2007). Yet the fact that there are earthquakes at all, albeit small ones, suggests the presence of compositional heterogeneity with both velocity-strengthening and velocity-weakening materials (Wibberley 2007) similar to the behaviour suggested for the Carboneras fault as mentioned above. Indeed, recent borehole data (SAFOD, phase III preliminary results, Hickman 2007) show that aseismic creeping fault strands have a $1\text{–}2\text{ m}$ thick clay-, serpentinite- and talc-bearing central zones (Moore & Rymer 2007), whereas microseismicity is localized in metric to decametric scale patches (the so-called ‘Hawaii Islands’) located about 150 m below the cored faults. Though we might speculate that the ‘Hawaii Island’ microearthquakes are located inside velocity-weakening patches (e.g., sandstones lenses embedded in the creeping sections), further studies (and drilling) are necessary to solve the issue.

**Rupture propagation**

**Dynamic weakening.** The diversity of material-dependent frictional responses observed at sub-seismic slip rates (velocity-strengthening v. velocity-weakening), disappears approaching seismic slip rates ($V = 1 \text{ m s}^{-1}$), as illustrated by the synthesis of laboratory rock friction data in Figure 7. In fact, all the rocks tested so far, including cohesive rocks (tonalite, diorite, gabbro, peridotite, serpentine, limestone, marbles, siltstone, serpentine: Hirose & Shimamoto 2005a, b; Spray 2005; Di Toro et al. 2006a, b; Hirose & Bystricky 2007; Han et al. 2007), and non-cohesive rocks (clay-rich fault gouges, Mizoguchi et al. 2007; Boutareaud et al. 2008b), have a low friction coefficient (on average, 0.2) and a strong velocity dependence at seismic slip rates, independently of the specific weakening mechanism involved (Figs 7 & 8). The experimental data are reliable, since (1) these low values for friction were obtained in different experimental configurations and apparatus (e.g., rotary torsion, rotary shears, torsion bars, Di Toro et al. 2004; Hirose & Shimamoto 2005a; Prakash & Yuan 2004) and (2) other materials, such as aluminium, in the same apparatus and under the same deformation conditions, are velocity-strengthening (Han, pers. comm.). In the case of frictional melting, field (Di Toro et al. 2006a) experimental (Hirose & Shimamoto 2005a; Spray 2005; Di Toro et al. 2006a) and theoretical analyses (Fialko & Khazan 2005; Nielsen et al. 2008) all indicate lubrication of fault surfaces in the presence of melts. Such a review of data from different velocity ranges shows that the low value for friction at $V = 1 \text{ m s}^{-1}$ is in contrast with the finding of the conventional experiments performed at $V < 0.01 \text{ m s}^{-1}$ and displacements of few centimetres at most, where $\mu$ is 0.6–0.8 (Fig. 7; Stesky et al. 1974; Byerlee 1978). The dramatic weakening found in these high-velocity experiments, might explain several seismological and geophysical observations, including: (i) why dynamic stress drops (the difference between initial stress and frictional stress while the fault is slipping) are larger than static stress drops (the difference between the average shear stress on the fault zone before and after the earthquake); (ii) the rupture propagation mode (see below); (iii) the increase in the ratio of radiated energy v. seismic moment with earthquake size (Mayeda & Walter 1996); and (iv) the production of heat during seismic slip (e.g., Lachenbruch 1980).
Another observation is that clay-rich gouges are also velocity-weakening at seismic slip rates (Fig. 7). This has important implications for the role played by the internal structure of the fault: earthquakes nucleating in velocity-weakening parts of the fault zone may propagate into and through zones such as clay gouges (or at greater depths, phyllonite) of velocity-strengthening (at low slip rates) behaviour if the rupture energy is sufficient to overcome the low-slip-rate barrier (Fig. 9). This will either bring this material into the high-velocity weakening regime and/or allow propagation through into another velocity-weakening region (Fig. 9c). Thus although a rupture tip propagating through a velocity-strengthening material will have more energy absorbed than it would do in a velocity-weakening material, the inference that the strengthening material will change the polarity of its velocity dependence at high velocity to that of weakening suggests that the dampening impact on propagation may not be as severe as previously thought (e.g., Scholz 2002). Such behaviour could be considered as ‘conditionally stable’ in a slightly different sense than that defined by Scholz (1998), but with the same outcome: a sufficient velocity kick will make it unstable, allowing rupture propagation. Whether or not the rupture tip can give the required velocity kick depends upon the slip rate behind the tip zone when it entered the velocity-strengthening material. This will be a function of the seismic moment up to this point, and hence rupture surface area in the velocity-weakening material, suggesting that there is a strong scale influence of heterogeneity on this process. Other key controls will be the rate of dampening during slip acceleration to the velocity-weakening state, and the relative and absolute sizes of ‘nucleating’ velocity-weakening zones and ‘conditionally’ stable zones, which are initially velocity-strengthening (Fig. 9c). The impact of other weakening processes such as those listed below can also affect the success of the rupture tip in continuing to propagate by absorbing less energy on the parts of the rupture surface already slipping at high velocity, thus retarding dampening. Hence, in many heterogeneous fault zones, ruptures may continue to propagate beyond velocity-weakening (at low slip rates) zones depending upon a variety of factors in which the spatial distribution of different fault zone materials plays an important part (Fig. 9c). This might in some cases reduce the degree of complexity of earthquake source physics.
The weakening mechanisms activated in the laboratory at seismic slip rates share the strong dependence with temperature. Examples of such weakening mechanisms are: water pressurization and vaporization (e.g., Brantut et al. 2008; Boutareaud et al. 2008b); dehydration (Hirose & Bystricky 2007) and decarbonation (Han et al. 2007) reactions; gelification (Goldsby & Tullis 2002; Di Toro et al. 2004; Roig-Silva et al. 2004) of silica-bearing rocks; local (flash) heating and eventually melting of the asperity contacts (Goldsby & Tullis 2003; Beeler et al. 2008) or bulk melting (Hirose & Shimamoto 2005a; Spray 2005; Di Toro et al. 2006a, b) of the fault surfaces and materials. Given that frictional heating depends upon the shear strain rate and hence (coseismic) slip zone width, the typical narrowness of slip zones (a few hundreds of microns to a few centimetres at most, see Sibson 2003 for a review) makes coseismic temperature increase an important issue. Furthermore, frictional heat diffusion is typically limited to a few millimetres in the slipping zone due to the low thermal diffusivity of rocks, so that the localization of heat in the slipping zone means that earthquake mechanics and dynamic weakening mechanisms are mainly controlled by local temperatures (Rice 2006; Rice & Cocco 2007; Pittarello et al. 2008). The narrowest slip zones are often (but not always) those generated by the most extreme localization of deformation at competence contrast boundaries between different lithologies or different fault rock materials. Hence it is likely that the localization of rapid deformation at competence contrast boundaries within fault zones or at the fault zone/host rock boundaries will favour dynamic weakening.

In all these cases, the continuity of the fault zone material(s) giving rise to the particular dynamic weakening process is an important issue – without this continuity in the slipping zone, the weakening process cannot operate efficiently and the rupture propagation may terminate. For example, water-saturated gouge slip zones have sufficiently low permeability to trap frictionally-heated pore water and hence cause dynamic weakening by thermal pressurization, provided they are laterally
and vertically continuous (Wibberley & Shimamoto 2005). Heterogeneity in permeability properties of the slip zone, such as could happen if the rupture front propagates onto a branch fault which splays into a higher-permeability damage zone, would diminish the efficiency of the dynamic weakening, leading to possible rupture termination and seismic asperities (Wibberley & Shimamoto 2005, Boutareaud et al. 2008).

The critical slip distance \( D_c \). One important parameter in earthquakes is the critical slip distance \( D_c \) over which strength drastically decreases because it controls the size of the rupture nucleation dimension, the magnitude of pre- and post-seismic slip and the length scale over which dynamic stress is concentrated at the rupture front (Marone 1998 for a review). In experiments performed at sub-seismic slip rates, \( D_c \) has a length comparable with the asperity diameter (c. 5–50 \( \mu \)m) of the bare surfaces. So \( D_c \) is related to the initial roughness of the sliding surface and is interpreted as the slip distance to renew the asperity contacts (Dieterich 1979). In the case of gouge in an experimental fault zone, \( D_c \) is controlled by the thickness of the zone of localized shear strain (Marone & Kilgore 1993). However, in nature, if a slip weakening distance exists, it is of the order of metres (Ide & Takeo 1997; Tinti et al. 2004; Ma et al. 2006). Simple upscaling of experimental data from drained gouge zones suggests that a deforming zone in the order of a hundred metres wide is needed to explain seismically-derived \( D_c \).
values (Marone & Kilgore 1993). Whilst gouge zones may indeed be this wide, field studies of fault zone internal structure suggest that typically much narrower slip zones are active during any one rapid slip event (Chester & Chester 1998; Sibson 2003; Wibberley & Shimamoto 2005; Di Toro et al. 2006a). The discrepancy between experimental and seismically-inferred \( D_c \) values might firstly be because in natural faults gouges are typically of low permeability causing undrained conditions during rapid shearing, so that thermal pressurization controls \( D_c \) (Wibberley & Shimamoto 2005), or by considering that asperities are larger in nature. This latter possibility requires the determination of fault roughness (Power et al. 1988; Power & Tullis 1992; Renard et al. 2006; Sagy et al. 2007) over a wide range of scales, particularly as geometric irregularities at larger scales than the thickness of the seismic slip zone are likely to be the most important. However, in high-velocity rock friction experiments, performed on smooth surfaces, the slip weakening distance is of the order of metres, and decreases with increasing normal stress, suggesting that it is the thermal history and the rheological evolution of the fault materials that controls \( D_c \) (Hirose & Shimamoto 2005b; Nielsen et al. 2008), as well as possibly its dependence on slip velocity in these experiments. Further studies are necessary to investigate the relationships between fault geometry and weakening mechanisms and their role in controlling \( D_c \) as well as integrating the results of constant-velocity experiments into a real earthquake model for widely varying velocity regimes.

**Type and speed of rupture propagation.** The type of rupture propagation during earthquakes is strongly influenced by fault geometry and fault zone structure. The two end member models of rupture propagation are the expanding crack model and the self healing pulse model (e.g., Heaton 1990; Zhang & Rice 1998; Beeler & Tullis 1996; Nielsen et al. 2000). In the expanding crack model, the earthquake nucleation region slips for the duration of the earthquake, and slip ceases on the entire fault when the rupture arrests. The self-healing pulse considers that only a small patch of the fault slips at any one time. Quantitative models that produce self-healing pulses consider large dynamic stress drops (e.g., Heaton 1990), not included in the rate and state friction law (Dieterich 1979; Ruina 1983) that describes rock friction experiments conducted at slip rates <1 mm s\(^{-1}\), but consistent with the large dynamic stress drops found in high-velocity rock friction experiments (Fig. 8), and with the presence of strong pre-rupture stress heterogeneity related, for instance, to fault geometry (Nielsen et al. 2000).

Preliminary field studies suggest that the rupture speed during earthquakes is controlled by fault geometry. For instance, though most earthquake ruptures propagate at velocities approaching the shear wave velocity (the so-called Rayleigh speed which is on average in most rocks 3 km s\(^{-1}\)), some ruptures propagate at supershear velocities (4.5–5 km s\(^{-1}\)) (Bouchon & Vallée 2003). Ongoing research suggests that supershear speed is achieved along straight fault segments, indicating that fault smoothness is one of the main parameters controlling rupture speed (Das 2007; Bouchon pers. comm.).

**Asperities.** High-velocity experiments indicate that the friction coefficient decreases at seismic slip rates to about 10–20% of its initial value (Fig. 7) and is slightly dependent on the normal stress (e.g., melt lubrication, Nielsen et al. 2008). Extrapolating these observations to natural conditions suggests: (1) the possibility of large coseismic stress drops in nature (up to 100–150 MPa at 10 km depth); and (2) that stress drop should increase with fault depth. Although **dynamic** stress drops as large as 100 MPa have been estimated for discrete fault segments of the ML 6.7 Loma Prieta earthquake (Bouchon 1997), and Fletcher & McGarr (2006) calculated an increase in the **static** stress drop with depth for some earthquakes, experimental results are at odds with the seismological observation that static stress drops are typically between 1 and 30 MPa, irrespective of earthquake size (e.g., Hanks 1977). One appealing explanation for the discrepancy – although not the only possible explanation – is the role of fault geometry. In the experiments, given the small size of the samples (usually less than 25 mm in diameter), it is not possible to reproduce the roughness of natural fault surfaces. The presence of geometrical asperities might impede the smooth sliding simulated in the laboratory, and experiments performed along pre-existing smooth surfaces underestimate the mechanical work expended (1) to override the asperities (dilatancy), and (2) to break the asperity junctions and short-cut the contractional jogs. The bulk effect of the introduction of surface roughness could be to increase the frictional resistance: roughness could buffer the dramatic decreases in strength suggested by the weakening mechanism discussed earlier. The problem could be tackled by numerical models which include the fault roughness and the constitutive laws of the new dynamic weakening mechanisms. The input roughness of the fault for the numerical model could be imported directly from the quantitative field reconstruction of the fault geometry (e.g., Sagy et al. 2007).
Rupture arrest

Looking at the larger scale, one of the most critical aspects of fault geometry is their action as barriers to earthquake propagation. For instance, the presence of kilometric scale dilational and contractional jogs and step-overs is one of the most reliable physical barriers to arrest the propagation of large earthquakes (King & Yielding 1984; Sibson 1985, 1986; Wesnousky 1988, 2006). However, ‘barriers’ are not only geometrical. For instance, several strike-slip earthquake ruptures stop along planar surfaces, suggesting that other mechanisms, such as the presence of velocity-strengthening materials (clays, serpentinites, etc.) might stop the propagation of the rupture. In the case of the repeated M5 earthquakes occurring in the Parkfield segment of the San Andreas Fault, one rupture tip could be associated with the southern fault bend where the 1857 M8.1 Fort Tejon earthquake, propagating from the south, stopped (geometrical barrier), whereas the northwestern rupture end of the Parkfield segment could be explained by fault rheology (Bakun et al. 2005).

However, rupture arrest may also occur because the elastic strain energy released from the wall rocks is not able to sustain the propagation of the rupture in the slipping zone. This, for instance, could be the case in the lower crust, where (i) fault rocks are velocity-strengthening (e.g., at temperatures >300 °C in the case of granite), and (ii) crystal-plastic deformation in the shear zones releases the elastic strain energy stored in the host rocks which, by decreasing the rupture driving forces, progressively stops the propagation of the rupture at depth. As the presence of phyllosilicates such as talc, phengite and chlorite in large fault zones is extremely frequent, and these minerals show temperature-activated creep at much lower temperatures than most other silicates, the role of crystal plastic deformation in rupture arrest even at relatively shallow levels in the Earth’s upper crust should be considered (Wibberley 2007; Imber et al. 2008). Alternatively, the energy loss in the damage zone (i.e., outside the slip zone) by fracturing the wall rocks and by absorbing the elastic strain energy released during rupture propagation may determine the arrest of the rupture (Andrews 2005).

A corollary stemming from findings into the role of fault geometric barriers in stopping earthquake propagation is that earthquakes therefore do not always know a priori if they are going to be small or large, and could be similar at least during their nucleation phase (e.g., Wesnousky 2006). Other than fault geometry, such earthquakes may differ only in the amount of elastic strain energy stored in the wall rocks and of its release during rupture propagation. Yet this suggestion conflicts with other studies that found much higher proportions of high-frequency radiated energy during the early stages of small earthquakes than large ones, suggesting that small and large earthquakes could be different from their very early initiation, perhaps due to differences in the initial stress (Olson & Allen 2005). The answer to this paradox may lie in considering the evolution of fault zone structure with growth of the fault, and in particular the roughness of the seismic slip zone (e.g., Dolan 2006). Small magnitude earthquakes, which may occur on small faults as well as large faults, often generate higher-frequency seismic waves in the early stages of the earthquake than do larger earthquakes, which necessarily occur on larger, more mature and probably ‘smoother’, faults. Hence the correlation of earthquake magnitude with proportion of early high frequency waves could simply be reflecting slip zone roughness within a complex fault zone structure. In particular, Sagy et al. (2007) found that fault surface roughness evolves with cumulated slip: parallel to the slip direction, small-slip faults are about one order of magnitude rougher than large-slip faults. The difference in geometry implies that the nucleation, growth and arrest of earthquakes should be different for small compared to large earthquakes and explains why the amount of high-frequency radiated energy should be smaller for large earthquakes. The way in which fault zone internal structure and heterogeneity in physical properties may control the final size of the rupture (and of the magnitude of an earthquake) is of great relevance in earthquake hazard mitigation such as early warning systems (e.g., Gasparini et al. 2007), and this necessitates further field, theoretical and experimental studies.

Concluding statements

Field studies of fault zones show a wide variety of internal structure depending on a number of geological variables such as displacement, lithology, depth and pre-existing fabric. A number of different mechanisms for incorporating wall rock material into the fault zone have been identified, each of them scale-dependent, and each relying on weakening of the surrounding host rock by a specific scale-dependent mechanism. The sum effect of these scale-dependent mechanisms in data complications over a wide range of scales may give the inaccurate impression that fault growth mechanisms are scale-independent. Nevertheless, whilst a certain amount may be gleaned from field studies of fault zones in terms of their growth mechanisms, relative chronological markers in the growth stages of a fault are not often
present. Hence inferences on growth patterns of fault zones from studying the final state alone must be treated with caution.

Concepts on permeability structure have advanced significantly in the past decade, but remain centred around the notion that two components can usually be identified in fault zones: a fracture-dilatant conducting component and a granular compacting barrier component. The relative distribution and properties, including anisotropy, of each component will govern the overall hydraulic behaviour of the fault. An advancing concept in this paper is that hydraulic and hydrodynamic behaviour of the fault during its activity may be predictable using concepts of critical state soil mechanics for the granular material and elastic crack mechanics for the fracture-dilatant component. Such advances have clear applications in considering the role of active faults in limiting hydrocarbon trap retention, evaluating the risk of slip instability during hydrocarbon production, and understanding the role of fluids in earthquake slip mechanisms.

Field studies illustrate why faults cannot simply be treated as discrete surfaces when modelling mechanical and fluid flow behaviour. Yet the complexity and heterogeneity in fault zone internal structure makes predicting fluid flow and seismogenic behaviour an extremely difficult task. For applications to the hydrocarbon industry, across-fault reservoir juxtaposition or non-juxtaposition is a critical geometric property to define in fault analysis. Yet juxtaposition is largely controlled by the geometry of accommodation of overall throw in the complex fault zone and adjacent volume: normal ‘drag’ folds may reduce reservoir offsets, as may the distribution of throw over several sub-parallel fault strands. Thus the recognition of this at the scale of seismic resolution or incorporation of such possible complexity at the sub-seismic scale in uncertainty estimates is crucial to prediction of across-fault reservoir communication. The last decade has seen significant advances in the way fault zone properties are quantitatively evaluated for estimates of hydrocarbon retention and production-induced pressure differences. These relatively recent approaches use rules-of-thumb to predict general fault zone structure based on previous experience and field examples, often by predicting average properties at any one point on a fault and calculating the variation in these properties over the entire fault ‘surface’ of interest. Such pragmatic approaches average out the complexity across the fault zone. Whilst such approaches have met with success up to a point, a next natural step is to evaluate the uncertainties induced in these averaging approaches by modelling statistical variations in properties across the fault zone and what impact they would have on the predicted outcome.

Linking fault zone internal structure characteristics to various aspects of seismogenic behaviour such as those described in this paper requires a leap of scientific imagination only possible by integrating structural observations with seismological data and laboratory frictional and petrophysical measurements on relevant fault rocks. The complexity of fault zones strongly influences a number of processes occurring during the lifetime of an earthquake rupture, particularly by the distribution of materials of contrasting mechanical and frictional behaviours. Indeed, recent findings suggest that whether the fault deforms by unstable earthquake slip at all, or suffers slow steady creep instead, is controlled at least partly by the contrast in velocity behaviours of different materials in the fault zone. Despite the recent increase in available laboratory data, the limits of scale, particularly in fault surface geometry, should never be neglected.

The authors would like to thank Bob Holdsworth for a very efficient and effective review of the paper. C. W. thanks the Geological Society publishing staff and book committee for encouraging this Special Publication to come to fruition. Data for Figures 3 and 4 were measured in a high-pressure gas deformation apparatus at Kyoto University in the laboratory of T. Shimamoto, who is thanked for his guidance. G. D. T. thanks M. Bouchon and S. Nielsen for discussions, T. Hirose, R. Han and K. Mizoguchi for sending their original data for preparing Figure 8 and R. Han for helping in assembling Figure 7. The costs of the work by G. D. T. were covered by grants from the Università degli Studi di Padova (Progetto di Ateneo 2007) and Progetti di Eccellenza Fondazione CARIPARO (CAassa di RIsparmio di PADova e ROvigo).

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