Journal of Structural Geology, Vol. 7, Nos. 3/4, pp. 437 to 457, 1985
Printed in Great Britain

0191-8141/85 $03.00 + 0.00

Brittle microtectonics: principles and practice

P. L. HANCOCK

Department of Geology, University of Bristol, Queen's Building, University Walk, Bristol BS8 1TR, U.K.

(*Received* 8 *February* 1984; *accepted in revised form* 17 *August* 1984)

Abstract—Brittle microtectonics as defined here is the application of mesofracture analysis to the solution of tectonic problems. For the determination of regionally significant stress or strain trajectories the ideal suite of structures comprises kinematic indicators such as mesofaults, shear zones, arrays of en échelon cracks, kink bands, stylolites and fibrous veins. However, in many tectonic settings joints are the only widespread structures capable of being analysed. The principal criteria for classifying joints into extension, hybrid and shear classes are fracture-system architecture and symmetry, surface morphology, dihedral angles and thin-section characteristics. The architecture of orthogonal extension joints is commonly T-shaped, the younger joint abutting the older one. Neighbouring conjugate hybrid or shear joints generally define X, Y or V shapes but some X patterns are artefacts of unrelated cross-cutting fractures. Conjugate joints enclosing dihedral angles of less than 45° are common; they are interpreted as hybrid failure surfaces initiated in the shear–extension fracture transition.

The orientations of many joint sets in platforms are related to far-field stresses generated during plate motion and the subsidence, uplift and inversion of basins. Even within a single joint set there is commonly field evidence to show that it developed during a multiphase failure sequence. Some joints are younger than folds but where a system was established before the close of folding the surfaces commonly become the sites of slip, dilation or pressure solution, and are thus transformed into structures no longer classed as joints.

PURPOSE AND SCOPE

BRITTLE microtectonics is regarded for the purposes of this account as the application of the techniques of mesofracture analysis to the solution of tectonic problems. Although brittle and semi-brittle mesofractures are also of intrinsic interest from the perspective of rock mechanics this aspect is not emphasized here. Field examples are selected largely from terrains of weakly deformed sedimentary rocks within which mesofractures may be the only widespread structures. Furthermore, mesofractures are more reliable indicators of regionally significant stress/strain trajectories in weakly deformed rocks than in thrust-fold belts, grabens and transcurrent fault zones where there may have been substantial vertical or horizontal rotations and complex strain histories. The use of the term mesoscale follows that of Turner & Weiss (1963) who employed it to embrace structures that range in size from less than a centimetre to a few metres, and that are observable in a single continuous exposure.

The structures considered in this paper include mesofaults, shear zones, arrays of en échelon cracks, kink bands, fissures, veins and pressure-solution seams, all useful brittle or semi-brittle kinematic indicators. The importance of these phenomena in microtectonic investigations has been emphasized already by workers such as Arthaud (1969), Blès & Feuga (1981), Choukroune (1969, 1976), Eyal & Reches (1983), Jaroszewski (1972) and Letouzey & Trémolières (1980). The benefits to be gained from analysing joints, the commonest of brittle structures, are highlighted here. In forelands and hinterlands they enable past and contemporary stress trajectories to be established (e.g. Engelder & Geiser 1980, Engelder 1982a,b, Hancock & Kadhi 1978, Hancock *et al.* 1984). Although joints and many other mesofractures are small-scale structures their regular arrangement within large areas (>1000 km^2) of weakly deformed rocks allows us to be confident that they are linked to tectonic processes. Indeed, Eyal & Reches (1983) have claimed that kinematically diagnostic mesostructures can yield more reliable stress directions than macrostructures.

METHODOLOGY

Principles of inferring stress/strain trajectories

Numerous experiments (reviews in Brace 1964, Price 1966, Hobbs *et al.* 1976, Jaeger & Cook 1976, Paterson 1978) have shown that when a brittle isotropic rock is loaded to failure in a conventional 'triaxial' compression test the resulting fractures are symmetrically orientated with respect to the three effective principal stresses ($\sigma_1' > \sigma_2' > \sigma_3'$; compressive stress positive); effective stress (σ') being total stress (σ) minus fluid pressure (p). The class of fracture that develops is related to the value of σ_3' and the stress difference (σ_1'–σ_3') compared with the tensile strength (T) of the rock. Three classes are recognized (Fig. 1, Table 1), each with its own range of dihedral (2θ) angles about σ_1', assuming a generalized composite failure envelope and a 30° angle of internal friction (ϕ). From a generalized failure envelope (e.g. Fig. 1) and from curves published by Price (1977, fig. 11) it is possible to estimate values of the effective normal stress (σ_n'), (σ_1'–σ_3'), σ_3' and p in terms of T. The orientations of the principal stresses can be determined knowing that at the time of failure an extension fracture is initiated perpendicular to σ_3' and in the principal stress plane containing σ_1' and σ_2', and that conjugate hybrid or shear fractures enclose an acute bisector parallel to σ_1'. Given a ϕ value of 30°, a 2θ angle of 45° within the hybrid

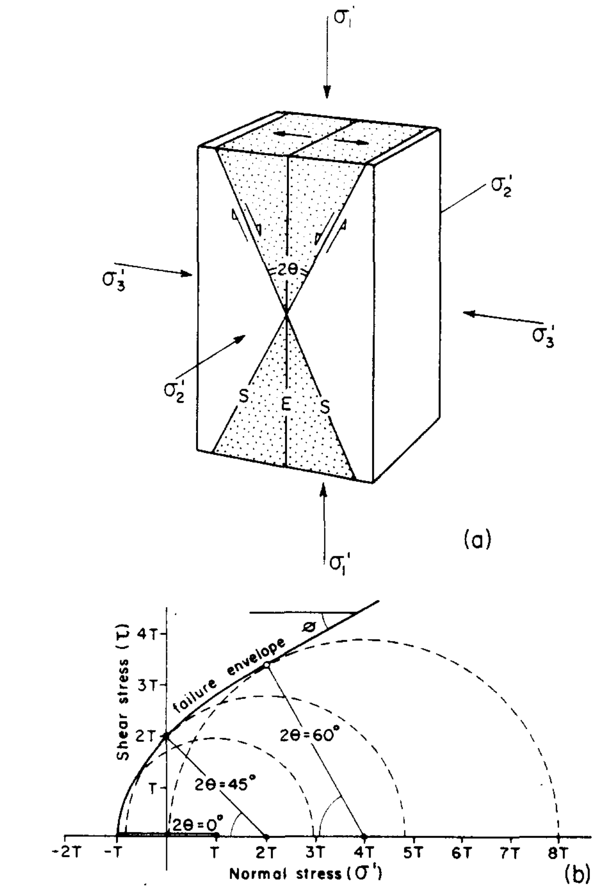

Fig. 1. (a) Block diagram showing relationships between effective principal stresses ($\sigma_1' > \sigma_2' > \sigma_3'$) and an extension fracture (E) and conjugate shear fractures (S) developed in a mechanically isotropic brittle rock. Stipple indicates the quadrants within which hybrid fractures form. (b) Composite failure envelope and Mohr circles constructed for 2θ = 0, 45 and 60°. T, tensile strength; ϕ, angle of internal friction.

fracture class divides it into two subclasses. When fractures enclose a 2θ angle of less than 45° both σ_3' and σ_n' are negative, whereas when 2θ is 45–60° the value of σ_n is positive, although σ_3' remains negative (Fig. 1). Dennis (1972, pp. 288, 291–295) called structures in the former subclass oblique extension fractures. Although a ϕ angle of 30° has been assumed for the purposes of the above exposition it is well known that it varies with lithology, generally being greater in more competent lithologies. As a consequence of strain, faults that develop in ductile rocks commonly enclose an obtuse angle about the compression direction (Ramsay 1980a, fig. 20). Rocks cut by pre-existing planes of mechanical anisotropy may fail by sliding on these planes, provided they are suitably orientated (Donath 1961, 1964), and, as McKenzie & Jackson (1983) have stated, there is no reason to believe that the relationship between the orientation of a stress field and a reactivated fault will be simple.

The simple relationship between conjugate brittle shears and principal stress axes has been much used by geologists since the publication of Anderson's (1942) book, *The Dynamics of Faulting*. Inherent in the Andersonian view of shear failure is that the influence of σ_2 is neutral and that the resulting deformation involves plane strain. Recently, Reches (1978, 1983), Aydin & Reches

Table 1. Classes of brittle fracture

Class	Failure mode, Etheridge (1983)	$(\sigma_1'-\sigma_3')$	Dihedral angle (2θ)
Extension fracture	Tensile failure	$<4T$	0°
Hybrid shear fracture	Extensional shear failure	$4T$–$8T$	1–60°
Shear fracture	Compressional shear failure	$>8T$	>60°

(1982) and Reches & Dieterich (1983) have discussed the geometry and kinematics of faults developed in a three-dimensional (i.e. triaxial) strain field, concluding that three or four sets in orthorhombic symmetry will form provided that the rock already contains suitably orientated pre-existing discontinuities. It is possible to combine Reches' ideas with those of Nelson (1981), who has pointed out that a single set of stylolitic solution seams result in uniaxial compaction. Conjugate faults or spatially associated extension fractures and stylolites can then be perceived as giving rise to biaxial (i.e. plane) strain (in stress conditions commonly known as triaxial in laboratory tests). Figure 2 illustrates relationships between principal stress/strain axes and structures developed in uniaxial, biaxial and triaxial strain fields. Note that the orientation of the stylolitic lineation (columns) on a solution seam is parallel to σ_1 or the Z strain axis (e.g. Buchner 1981). Techniques of determining stress/strain axes from lineations on faults bounding blocks displaced in a stress field unrelated to that which initiated the faults have been formulated by workers in the French schools of microtectonics, notably Arthaud (1969) and Angelier (1984). Their elegant methods are, however, of limited value unless a large population of lineated faults is present.

Data collection and analysis

There is a voluminous literature on the statistical aspects of sampling and analysing orientation data that for reasons of space cannot be reviewed here. The following suggestions are based on the author's experience and have proved to be practicable and cost/time effective in many situations. Microtectonic orientation data should be collected from small sampling sites (stations) that are structurally homogeneous domains. Collecting data from a locality within which there is, say, a change in layer dip or fold plunge, or which contains a major fault may mask the influence of these controls and lead to 'fuzziness' in the pattern. In general, tightly clustered point diagrams result from sampling small volumes of rock less than 5000 m^3. If a continuous exposure contains obvious variations in mesofracture pattern it is prudent to site several stations within it. Limitations of exposure generally prevent station locations being established on the basis of a grid. In addition to recording orientation data it is equally important to record details about fracture style, morphology, dimensions, separation, architecture and the influence of lithology; topics elaborated in later sections of this paper. In order to ensure that observations are made

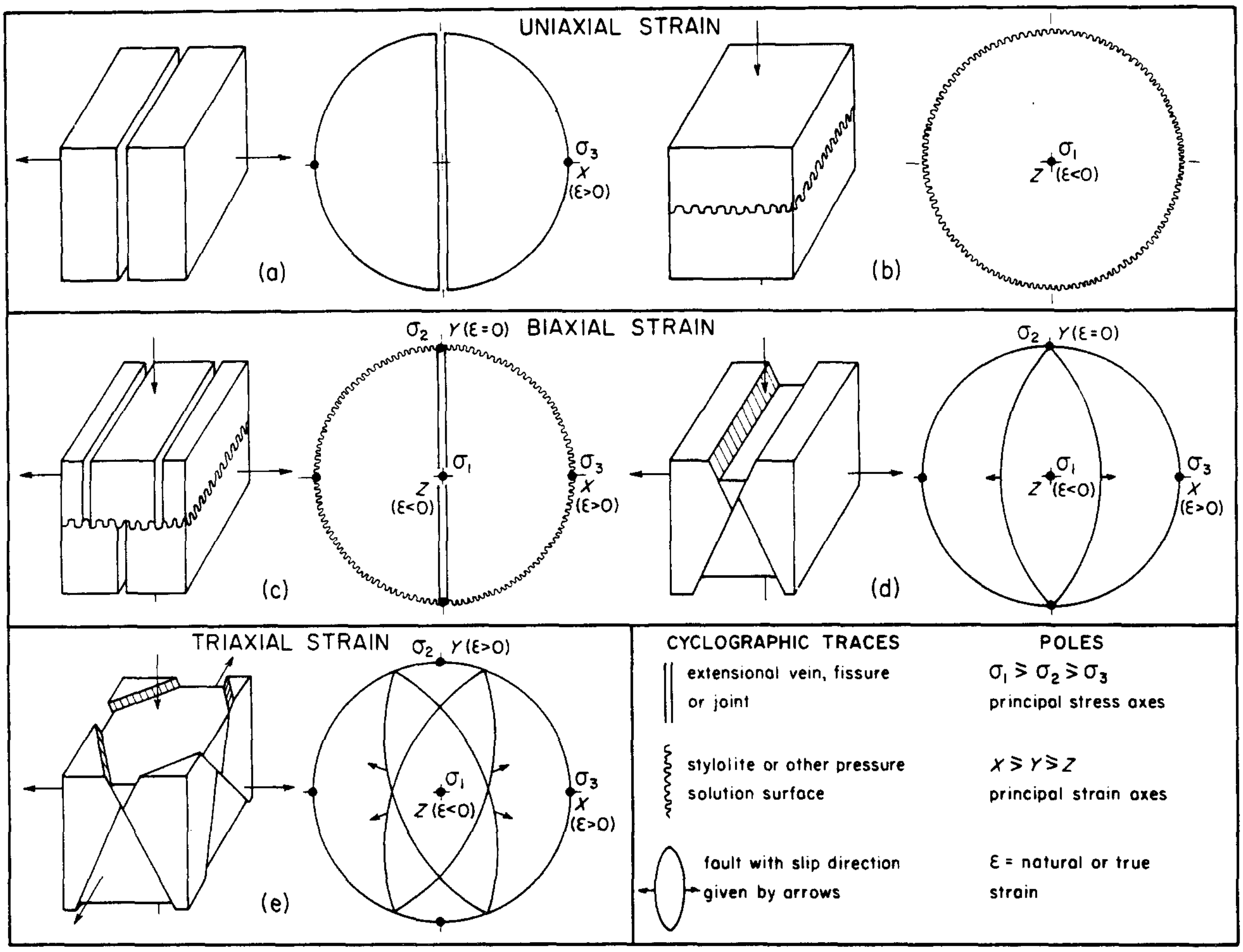

Fig. 2. Schematic block diagrams and stereograms illustrating (a & b) uniaxial, (c & d) biaxial and (e) triaxial strain resulting from the development of brittle mesostructures. (e) is after Reches (1983, fig. 1b).

systematically and are susceptible to statistical analysis it may be worthwhile devising a mesofracture logging form.

The initial aim during analysis is to understand what happened in a small volume of rock and then to integrate conclusions from it with those from adjacent volumes until subdomains can be recognized and a synthesis attempted. Each station should be analysed separately using stereographic techniques so that the symmetry of the pattern with respect to possible controls such as layer dip, fold plunge or azimuth can be established. Simplified data from each station can be plotted on synoptic maps or diagrams; especially effective in the final stages of analysis are trend-line maps showing the orientations of the inferred maximum and minimum horizontal stress axes, in addition to fracture traces. Horizontal stress trajectories correspond directly to σ_1 and σ_3 only in wrench (σ_2 vertical) regimes. For example, the maximum horizontal stress in terrains dominated by normal faults will be σ_2 in terms of the three-dimensional stress field.

KINEMATIC INDICATORS

In this section, field aspects of mesofractures and allied structures that provide unambiguous information about directions of shortening, dilation or shear are discussed. Categories of structure are not necessarily mutually exclusive and some, which are not the products of brittle behaviour, are included because of their close association with brittle mesofractures.

Mesofaults

As Ramsay (1980a) has noted there is a transition from brittle to ductile shear zones (Fig. 3). In this account the name fault is employed if a plane of frictional sliding is present (Figs. 3a & b) whereas if the strain gradient varies smoothly across the structure it is called a shear zone (Figs. 3c & d). A shear zone may subsequently fail as a fault. The arbitrary separation of mesofaults from macrofaults is taken at a displacement of about 5 m and a fault plane area of about 1000 m^2. The

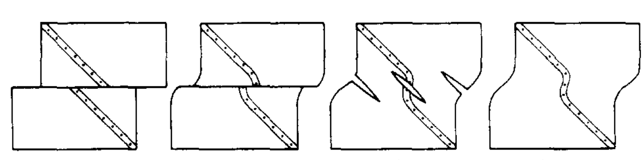

Fig. 3. Modes of faulting. (a) Brittle fault. (b) Semi-brittle fault. (c) Brittle-ductile shear zone. (d) Ductile shear zone. After Ramsay (1980a, fig. 1).

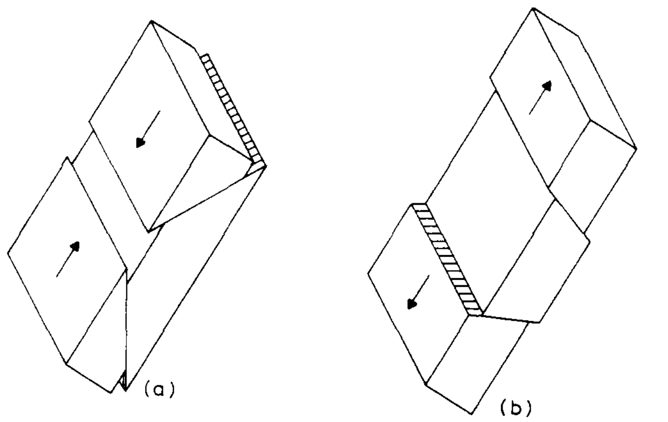

Fig. 4. (a) Conjugate contraction faults. (b) Conjugate extension faults.

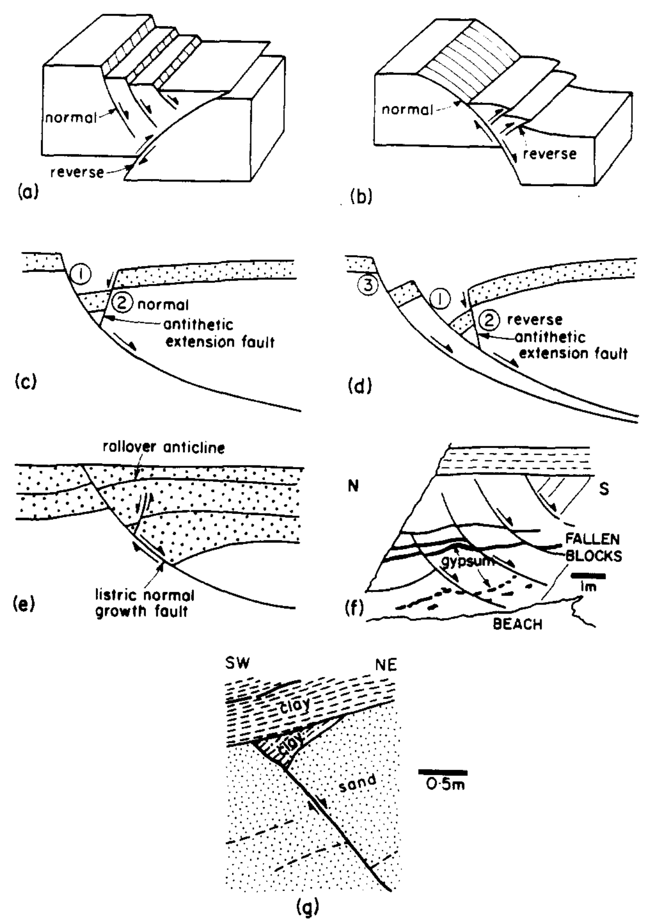

Fig. 5. Aspects of curviplanar faults. (a) Normal faults in the hangingwall of a convex-upwards thrust. (b) Reverse faults in the hangingwall of a convex-upwards normal fault. (c) and (d) Back rotation of an antithetic fault (2) into a reverse attitude in the hangingwall above a listric normal fault. Note that relative to layering, fault (2) remains an extension fault. Fault sequence is numbered (1)–(3). (e) Features associated with a listric normal growth fault. (f) Listric normal mesofaults in Triassic sediments, Watchet, Somerset, England. (g) Intraformational normal mesofault displacing Neogene sediments, Havza basin, Turkey. Because the bedding in the hangingwall has been back rotated the fault is probably listric, becoming gentler in inclination below the level of the exposure. (a) and (b) after Mattauer (1973, fig. 6.55), (c) and (d) based on Jackson *et al.* (1982, fig. 4), (e) after Reading (1978, fig. 6.55), (f) drawn from a photograph, (g) after Hancock & Barka (1981, fig. 4d).

tions of mesofaults are identical to those of macrofaults. The kinematic terms, normal, reverse and strike–slip are the most appropriate to use where faults are orientated symmetrically about the horizontal, but where dip–slip faults are symmetrically inclined about sedimentary layers it may be preferable to classify them as contraction or extension faults (Norris 1958) (Fig. 4). Employing Norris' classification (e.g. Bevan in press a) may simplify formerly cumbersome groupings of fault sets based on angles to the horizontal. Contraction faults at angles as small as 15° to layering commonly act as 'ramps' connecting bedding-parallel 'flats' within small-scale thrust imbricates.

Although many mesofaults in macrofault zones belong to the same class as the host macrofault this relationship is not universal, especially where there has been a complex strain history in the hangingwall of a thrust or low-angle normal fault, or where the macrofault is curviplanar and the mesofaults are accommodation structures (Mattauer 1973, Yielding *et al.* 1981, Gibbs 1983, 1984, Philip & Meghraoui 1983) (Figs. 5a & b). Furthermore, after about 25° of rotation in the hangingwall of a listric normal fault, some former antithetic normal faults become reverse (Jackson *et al.* 1982), although its extensional character relative to layering should remain clear (Figs. 5c & d). Although there are geological and geophysical observations or inferences in favour of some normal macrofaults being listric (e.g. Bally *et al.* 1980, Wernicke & Burchfiel 1982, Jackson & McKenzie 1983) there are fewer reports of listric mesofaults. Field criteria suggestive of a normal fault being listric include: (a) a curved profile; (b) back-rotation of the hangingwall (an effect which can also be a result of the domino-like rotation of blocks bounded by planar faults [Wernicke & Burchfiel 1982]) and (c) geometrically necessary accommodation structures, such as antithetic faults and rollover anticlines (Figs. 5 c–g). Planar high-angle normal faults characterize terrains that have experienced minor inhomogeneous extension (Wernicke & Burchfiel 1982). These faults may be accompanied by inward-facing monoclinal flexures and associated extension faults, especially where they frame grabens (Al Kadhi & Hancock 1980) (Fig. 6).

It is important to record whether a fault is syn- or post-depositional; that is whether it was active during sediment accumulation (i.e. a growth fault) or after it. Evidence in favour of a syndepositional origin includes: (a) preservation of a greater thickness of sediment on the downthrow side, provided there has been no erosion of the upthrow side (e.g. Gill 1979) (Fig. 5e); (b) the presence of a rollover anticline and/or antithetic faults if the principal fault is listric; (c) soft-sediment effects associated with the fault (e.g. Gill 1979, Pickering 1983) and (d) burrows cutting slump faults (Farrell 1984). A fault active during the deposition of a formation but not during sediment accumulation might show: (a) truncation of the fault plane by an intraformational erosion surface and (b) erosion of the back-rotated hangingwall if the fault is listric (Fig. 5g). Restriction of a fault to a layer is not by itself indicative of syndepositional or intraformational development because the distribution

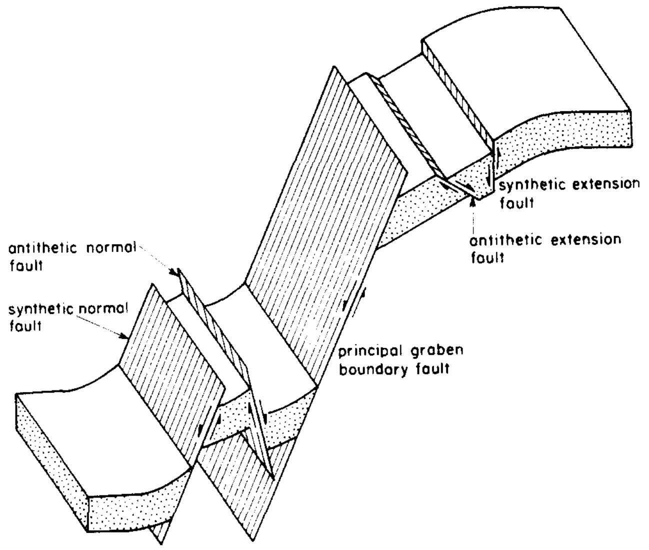

Fig. 6. Structures characteristic of planar high-angle normal fault zones such as those bounding grabens in terrains that have been inhomogeneously extended by a small percentage. After Al Kadhi & Hancock (1980, fig. 6).

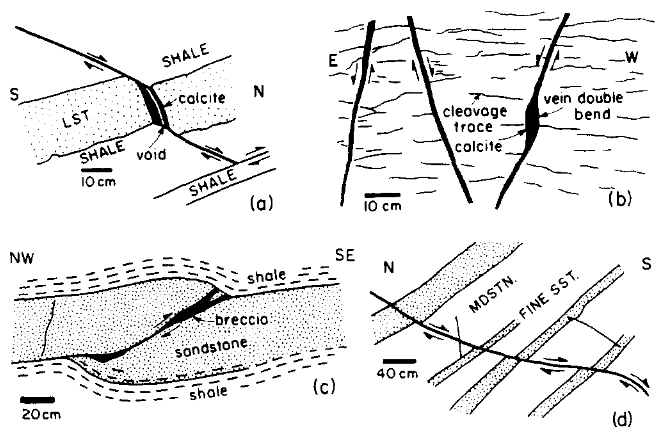

Fig. 7. Accommodation phenomena along curviplanar faults displaying a double bend. (a) Normal fault refracted through a Jurassic limestone interbedded with shales. Where there has been oblique extension in the limestone a calcite-filled 'pull-apart' vein has formed. Watchet, Somerset, England. (b) Calcite-filled 'pull-apart' vein exposed in plan on bedding plane in cleaved Palaeogene limestones about 60 km west of Sinop, Turkey. (c) Contractional mesofault with breccias developed in the extensional zones caused by sliding along a double flat-ramp topography within a Jurassic sandstone. Kimmeridge, Dorset, England. (d) Lazy double bend along an extensional mesofault displacing Triassic sediments. Note that there are no visible mesoscopic accommodation structures, probably as a consequence of the ductility of the fine-grained rocks. All the sketches were drawn from photographs.

of many mesofaults is related to the different mechanical responses of contrasting rock types.

Subparallel to the slip vector the traces of some mesofaults display a double-bend related to either lithologically controlled refraction or the presence of two offset extension fractures that have become connected during evolution of the fault zone (Segall & Pollard 1983a) (Figs. 7a & b). Where there has been extension across the bend a pull-apart vein will be developed, and if the fill is fibrous the axes of the fibres may indicate the extension direction. Compression across a double bend can give rise to an oblique stylolite lineation (Elliott 1976, Marshak *et al.* 1982). The accommodation effects associated with a contractional mesofault that is asymptotic with both the base and top of a bed are illustrated in Fig. 7(c). Adjacent to some gentle double bends there are no mesoscopic accommodation effects, especially if the faulted rocks are fine grained (Fig. 7d).

Some normal macrofaults are accompanied by mesofaults orientated approximately normal to bedding suggesting they were initiated as extension fractures. Angelier & Coletta (1983) have proposed that slip on such extension fractures commences when there has been about 25–30° of rotation of fault blocks and elongation reaches 50–150%.

Faulting in a porous rock, such as a weakly cemented sandstone, leads to dilatency followed by a volume decrease in the fault zone (see Aydin & Johnson 1983 for discussion). Cataclastic zones or veins in porous sandstones have received a variety of names including deformation band (Aydin & Johnson 1983), microfault (Jamison & Stearns 1982) and granulation seam (Pittman 1981). Individual granulation seams are generally a few millimetres in thickness, but commonly anastomose in zones up to a few centimetres wide across which displacement is measurable in centimetres. Aydin & Johnson (1983) suggest that striated slip surfaces bounding granulation seams mark a change from continuous to discontinuous deformation. Bevan (in press b) has noted that granulation seams in Cenozoic sands in southern England evolved within contractional fault zones but not extensional fault zones.

The faulting of indurated rocks results in an initial volume increase (fault breccia) but with continued displacement there is progressive cataclasis and the development of fault gouge of the type described by Engelder (1974). Care should be taken before deciding that all clay-like seams along fault zones are gouge; in some settings the 'clay' has been hydraulically injected from below, or even precipitated from hydrothermal fluids.

Some brittle mesofaults are accompanied by an array of pinnate (feather) joints that intersect the host fault normal to the slip vector and which subtend an acute angle with it that closes in the direction of displacement of the block containing the joints (Fig. 8a). Commonly, the joints are restricted to one side of the fault. Their distribution in the walls of the small contractional fault illustrated in Fig. 8(b) is interpreted as indicating that they preferentially developed in the extensional quadrants relative to a locking or sticking point on the fault (Fig. 8c). Repeated displacements from different locking points on a fault would lead to interfering arrays and the presence of pinnate joints on both sides of the fault along part of its length. As Fig. 8(b) also shows, there are two additional arrays of pinnate joints related to increments of slip on bedding planes. Engelder (1974) has demonstrated that microfractures cutting quartz grains in sandstones adjacent to a cataclastic fault gouge are orientated parallel to σ_1 and oblique to the fault zone. Pinnate joints might be mesoscopic analogues of such

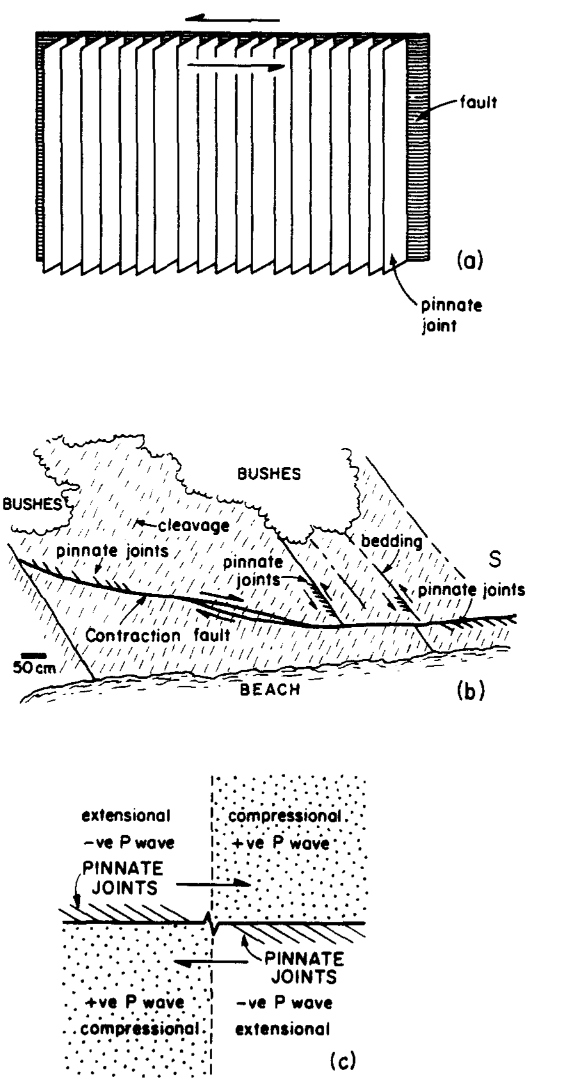

Fig. 8. (a) Pinnate joints related to a hypothetical sinistral strike–slip fault. (b) Cleaved Devonian mudstones cut by a contraction fault at the northern end of which pinnate joints are restricted to the hangingwall and at the southern end of which they are in the footwall. Milford Haven, southwest Wales. Drawn from a photomosaic. (c) Possible relationship between the distribution of pinnate joints and extensional quadrants about a locking-point on a fault.

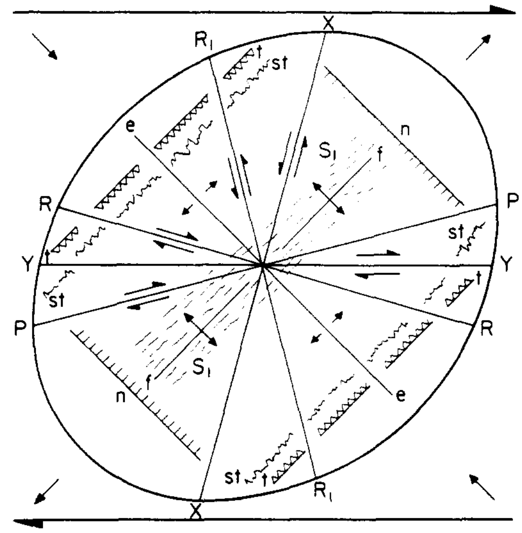

Fig. 9. Compilation diagram illustrating en échelon structures characteristic of strike–slip fault zones evolving during simple shear. R and R_1, Riedel and conjugate Riedel shears; P, X and Y, P-, X- and Y-shears; e, extension joint, fissure or vein; n, normal fault; t, thrust; st, stylolite; f, fold; S_1, cleavage or other foliation. Loosely based on Harding (1974) and Bartlett *et al.* (1981, fig. 3).

microcracks, which are also known to form in association with faults in experimentally deformed rocks (Friedman & Logan 1970). Wedge veins (Segall & Pollard 1983a) contained in the extensional quadrants on opposing sides of strike–slip faults in granite also possess the same geometry as many pinnate joints.

The upper crustal levels of many broad transcurrent fault zones are characterized by a variety of second-order en échelon structures (e.g. Tchalenko & Ambraseys 1970, Harding 1974, Hancock & Barka 1981, Gamond 1983). Bartlett *et al.* (1981) reviewed earlier authors' genetic classifications and integrated them with their own experimental results. A compilation diagram of possible en échelon structures in a right-lateral fault zone developed during simple shear is illustrated in Fig. 9. The type of structure that develops depends upon the ductility of the rocks in the zone, whether they are layered, the orientation of any layering, the relative magnitudes of the vertical and horizontal stresses, and the total displacement during any increment of movement. A fault parallel to the zone generally indicates that there has been substantial slip. Some second-order directions are themselves composite, for example the 'Riedel-within-Riedel' structures described by Tchalenko (1970). Where there has been horizontal shortening across a zone in addition to shear along it, the secondary compressional structures (folds and thrusts) initiate at smaller angles to the zone than when shear is simple, whereas extensional structures (normal faults and veins) form at greater angles (Sanderson & Marchini 1984, fig. 5). Shortening across the zone, in addition to transcurrent shear, reverses these relationships.

Some mesofaults are lineated and some lineations are accompanied by asymmetric steps roughly perpendicular to them (Fig. 10). The risers of the steps are generally incongruous if they accompany frictional-wear striations (Gay 1970) or oblique-stylolite columns, but are congruous where related to accretionary growth fibres. One lineation can be superimposed coincidentally or obliquely on another of the same or different type; and if there is a 180° change in slip sense accretionary fibres can grow in former solution hollows (i.e. combining the features shown in Figs. 10b & c). On a few faults the erosive tectonic tool responsible for frictional-wear striations remains in the pit at the distal end of the groove (Tjia 1972, Barka & Hancock 1984) (Fig. 10a). Engelder (1976) suggested that groove lengths of wear tracks (striations) are equal to or less than the slip distance during a displacement event. Polished and lineated, closely spaced fractures are characteristic of some deformed ('scaly') mudstones and have been reported from sediments in the Middle America trench slope by Lundberg & Moore (1981).

Where a lineation was generated during slip in a stress field of identical orientation to that responsible for

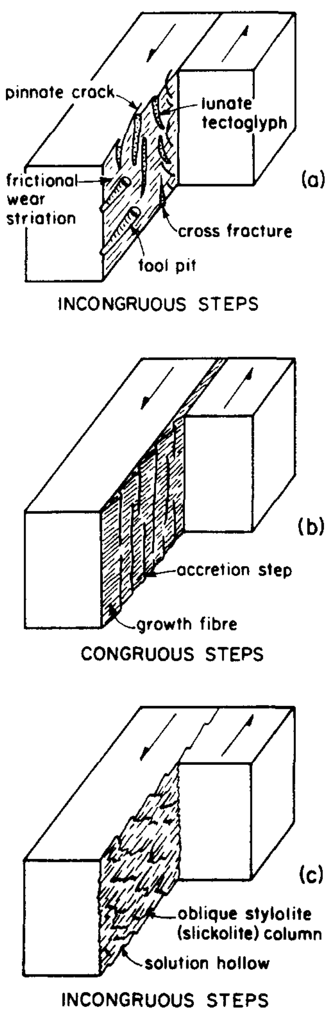

Fig. 10. Types of lineations and steps on fault surfaces. (a) Frictional wear. (b) Accretionary growth of crystal fibres. (c) Oblique pressure solution.

fracture initiation it is possible to infer the approximate attitudes of axes knowing that a plane containing the lineation and the normal to the fault defines the $\sigma_1\sigma_3$ plane, and that σ_2 was perpendicular to the lineation in the fault plane. σ_1 and σ_3 orientations can be estimated if the displacement sense is known and a realistic ϕ value is assumed. Stress axes should be inferred from a large population of lineations if slip occurred in a stress field unrelated to that which initiated the faults (see e.g. Arthaud 1969, Angelier 1984).

Shear zones

Mesoscopic shear zones in sedimentary rocks characteristically contain en échelon veins facing against the sense of shear or a cleavage and grain-shape fabric facing with it, or both (Fig. 11a) (e.g. Knipe & White 1979, Ramsay 1980a, Ramsay 1982, Ramsay & Huber 1983, Rickard & Rixon 1983). Generally, shear zones are better developed in sandstones and limestones, rather than mudstones, probably because in many mudstones shortening is achieved by the formation of a cleavage. Beach (1975) distinguished between two types of en échelon vein arrays (Fig. 11b), and proposed that veins in type 1 arrays are initiated as shear fractures while those in type 2 arrays are extension fractures. Other workers (e.g. Ramsay & Graham 1970, Ramsay & Huber 1983) have claimed that planar veins in type 1 arrays at 45° to zone margins (135° measured in the direction of shear) are extension fractures initiated normal to the maximum incremental extension within a zone of simple shear. Hancock (1972, 1973) has argued on the basis of the range of angles between veins and arrays, and the evidence for displacement along some veins, that they can occupy Riedel shears, hybrid fractures or extension fractures (Figs. 11c & d). Ramsay & Huber (1983, fig. 3.21) have explained the variation in angle between extensional veins and zone boundaries by relating them to positive or negative dilation across the zone, the angle being smaller where there has been positive dilation. If a rock mass contains suitably orientated pre-existing fractures they are likely to dilate in preference to the formation of new cracks (e.g. Ramsay 1967). 2θ values between conjugate arrays range from greater than 90° in ductile rocks to as small as 15° in brittle rocks (also see Ramsay 1982, fig. 18). Continued shear along a zone leads to it growing in width and the veins becoming sigmoidally distorted (Durney & Ramsay 1973). Because the distorted veins illustrated in Fig. 11(e) are limited to the central segment of the zone they may locate its nucleation point. The hinge lines of 'folded' veins are orientated normal to the slip vector provided the sense of shear has not changed since vein initiation. Some shear zones decay into faults or even joints (Simpson 1983), and likewise some mesofaults degrade into shear zones (Fig. 11f).

Kink bands

Although many kinks (Gay & Weiss 1974) are ductile structures formed in metamorphic rocks or unconsolidated sediments (e.g. Van Loon *et al.* 1984) some deflect brittle structures such as closely spaced joints, and are themselves parallel to associated brittle structures. Most 'brittle' kinks are negative (i.e. there is shortening in the plane of the pre-existing anisotropy) but positive kinks, comparable to shear bands (White *et al.* 1980), occur in some sedimentary rocks.

Stylolites and allied solution seams

Despite not being brittle deformation phenomena many widely spaced stylolitic and planar pressure-solution seams occur in close conjunction with brittle structures. They have been called anticracks by Fletcher & Pollard (1981) who have emphasized their association with veins and small faults. From either the attitude of the stylolitic lineation as defined by the axes of interlocking columns and sockets, or the apparent offset of two oblique markers the direction of shortening can be determined (Fig. 12a). Some workers (e.g. Buchner 1981) have employed the orientation of the lineation (the stylolite *sensu stricto* of some workers) to determine σ_1 axes at the time of stylolitization. The offset of a structure oblique to a seam across which there has been

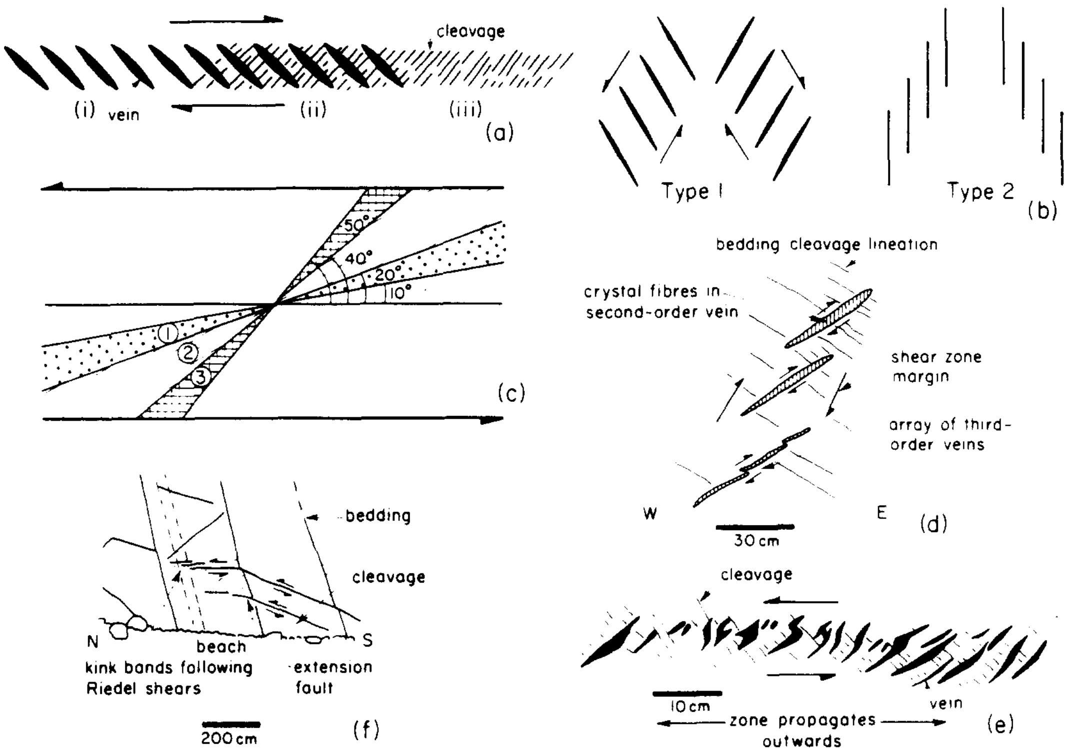

Fig. 11. Shear zone structures. (a) Structures of brittle–ductile zones (i) en échelon veins, (ii) en échelon veins and a secondary foliation, (iii) secondary foliation. (b) Type 1 and type 2 vein arrays according to the classification of Beach (1975, fig. 1). (c) Possible interpretations of undeformed en échelon veins subtending different angles with the margins of a simple shear zone, that is one across which there has been neither positive nor negative dilation: field (1) Riedel shears, field (2) hybrid fractures, field (3) extension fractures. (d) Part of an array of en échelon quartz veins exposed on the upper surface of a bed of Devonian sandstone, Freshwater West, southwest Wales. Note that the sense of displacement along the second-order veins can be inferred from growth fibres and the array of third-order veins. (e) Shear zone in cleaved Devonian mudstones, St. Ishmaels, southwest Wales. Note that the veins are more deformed in the central segment of the array. (f) Extension fault decaying into an array of en échelon Riedel shears in a cleaved Devonian mudstone, Freshwater West, southwest Wales. (d) to (f) were drawn from photographs.

perpendicular shortening is always 'normal' relative to the seam (Figs. 12a & c), and if the older structure was a formerly continuous vein an array of en échelon, but not overlapping, segments results. From the geometry of the vein segments (Hancock & Atiya 1979) or the maximum height of stylolitic columns (Fletcher & Pollard 1981) the width of dissolved material can be estimated. Coeval stylolites and extensional veins normal to each other have been called styloboudins by Mullenax & Gray (1984). Commonly, there is evidence for some stylolites being older than associated veins and *vice versa*, indicating that the development of even a single set of veins or stylolites is a multiphase process.

Jaroszewski (1972) and Mattauer (1973) have emphasized that there is a continuum of structures from stylolites that bear columns normal to a seam, through those on which the columns are oblique to the seam (slikolites) to those within which the lineation is subparallel to the seam (Figs. 10c and 12b).

Slow deformation by pressure solution in clast-supported conglomerates leads to pitting (Ramsay 1967). Pitted pebbles are ruptured if the matrix strain rate exceeds a critical level (McEwen 1981). The geometry of the failure surfaces offsetting clast margins is controlled by the orientations and magnitudes of the bulk principal stresses and by the shapes of the pits and pebbles.

Veins

Many syntaxial or antitaxial growth fibres in veins record the history of principal strain increments (Durney & Ramsay 1973, Ramsay 1980b, 1981, Ramsay & Huber 1983), although as Cox & Etheridge (1983) have noted this depends on nucleation and growth mechanisms. Undeformed fibres in extensional veins are perpendicular to the vein margin whereas those in hybrid fractures or shear veins are oblique to the vein margin (Fig. 12d). Where fibres subtend small angles with a vein this leads to a fibre sheet containing a fibre lineation (slickencrysts) within the treads of congruous accretion steps, the risers of which are the breaks across the fibres (Fig. 10b). The sense of dilation across a non-fibrous vein can be determined provided that it cuts two older planar structures oblique to each other and normal to the vein (Fig. 12d). According to Durney & Ramsay (1973) and Ramsay & Huber (1983) a change in the direction of dilation leads to curved fibres. Some antitaxial fibres contain wall-parallel inclusion bands indicating that they formed in response to a crack–seal mechanism (Ramsay 1980b). Brecciation of the wall rocks adjacent to a fracture zone is caused by hydraulic bursting when there is an abrupt fluid pressure drop within the main fracture zone (Phillips 1972).

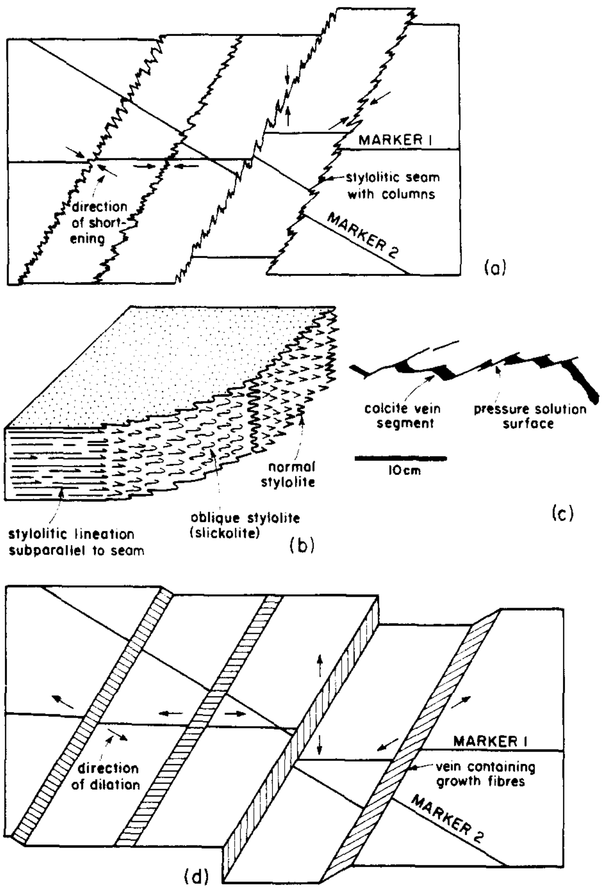

Fig. 12. (a) Different amounts and senses of offset of two markers across four parallel stylolites on which shortening directions are oblique to each other. (b) Transition from a normal stylolite through a slickolite to a surface bearing a subparallel stylolitic lineation. After Blès & Feuga (1981, fig. 2.4). (c) Array of vein segments related to normal pressure solution across surfaces oblique to a formerly continuous vein. Carboniferous Limestone, Hook Head, Ireland. (d) Different amounts and senses of offset of two markers across four parallel veins on which dilation directions are oblique to each other.

Fissures

A crack with a void space is here called a fissure. According to Jackson & McKenzie (1983) fissures are unlikely to penetrate to more than 500 m depth. Fissures generated during increments of seismic displacement are ephemeral structures, likely to be sealed or subsequently filled by vein material or sediment.

Sediment-filled mesofractures

Whereas many of the above types of mesostructure have been the focus of attention of structural geologists, sediment-filled fractures have attracted interest only recently and hence there is as yet no consensus about their classification and interpretation. In addition to structures related to purely sedimentary processes this neglected category includes a variety of fractures that from their internal geometry, symmetry and regular organization are interpreted as being of tectonic origin. Provisionally, two main classes can be recognized.

(1) *Intrusive sedimentary dykes*. This class comprises tabular zones up to tens of centimetres in width containing fluidized or brecciated sedimentary material that has been hydraulically injected, generally from below. Winslow (1983) has described important examples of 10 cm–1 m wide dykes that were intruded into former cross-joints that dilated in the leading edges of Andean thrust sheets which were being shortened during their emplacement. On a smaller scale the thin (<1 mm) *dewatering veins* described by Carson *et al.* (1982) from Japanese trench slope sediments also belong to this class because they are interpreted as having been hydraulically injected during tectonic dewatering. *Healed extension fractures* (Carson *et al.* 1982) are also 1–2 mm dark claystone seams but they contain aligned phyllosilicates and are interpreted as having been opened during dewatering, but closed during subsequent burial. Narrow clastic dykes are also injected into overlying sediments during seismic activity or as a consequence of dewatering from a basal fault (e.g. Farrell 1984).

(2) *Neptunian dykes*. Neptunian dykes result when sediment filling is from above, either contemporaneously with sediment accumulation or later. Intraformational neptunian dykes contain sediment derived from the enclosing formation and they may permit amounts of extension during basin filling to be estimated. When an already indurated rock is stretched, fissures will open and be filled either contemporaneously or later. The geometry of their walls is commonly controlled by that of pre-existing fractures. Some neptunian dykes of this type are related to deep-seated stretching, others (e.g. the filled 'gulls' of Pleistocene age in England) are related to more superficial movements.

JOINTS

Where kinematic indicators are rare or absent it is necessary to analyse systematic (Hodgson 1961a) joint systems to derive stress trajectories. The name joint (in use since at least the 18th century) is employed here as a serviceable field term to describe a barren, closed fracture on which there is no measurable slip or dilation at the scale of observation. If any mineral fill, including crystal growth fibres, is visible in the field the structure is better called a vein. Likewise, any detectable slip, even if only a few millimetres, places the fracture in the fault category, many fault planes also being shear veins. Fractures bearing frictional-wear striations or a stylolitic lineation are classed with faults and solution seams, respectively.

Although the organization of a joint system commonly mirrors some aspects of the bulk strain history of a region the development of the fractures achieves minimal strain. For example, Segall & Pollard (1983b) calculated from some unusually well-exposed joints in a Sierra Nevada granodiorite that jointing caused only about 0.01% elongation. Segall & Pollard (1983a) also showed that extension joints controlled the nucleation of younger strike–slip faults trending obliquely to the joints.

Extension joint? Shear joint? Hybrid joint?

Because joints are kinematically enigmatic structures their interpretation has generated controversy. For example, Scheidegger (1982, 1983) regards many joints, especially those belonging to orthogonal vertical sets, as shears even when, as in the Appalachian Plateau (Engelder & Geiser 1980) and Alberta (Babcock 1973, 1974), the same sets have been interpreted as comprising extension fractures. Investigators working in other settings such as the Canadian Rocky Mountain foothills (e.g. Muecke & Charlesworth 1966) and central Arabia (Hancock & Kadhi 1978, 1982) have attributed some sets to extensional failure, some to shear failure and others, enclosing small dihedral angles, to failure in the shear–extension fracture transition (i.e. hybrid fractures). The possibility of interpreting pairs of sets enclosing a small dihedral angle as belonging to the shear–extension fracture transition had been highlighted earlier by Muehlberger (1961), although the examples he cited from the Allegheny Plateau have now been reinterpreted by Engelder & Geiser (1980) and Engelder (1982a, 1985) as two extension fracture sets of different ages. Hence despite their characterless features it is important to attempt to assess, mainly in the field, whether a joint set comprises extension, hybrid or shear fractures. The following ten criteria can provide clues but it is emphasized that no single one is likely to be diagnostic.

(1) *Microscopic characteristics.* A thin-section across a crack classified as a joint in the field might reveal whether on a microscale there is evidence of shear or dilational offset (Hobbs *et al.* 1976, p. 293, Engelder 1982b). Irregular crack walls may be bilaterally symmetrical and thus indicative of dilation alone or they may match only if a component of shear, not detectable on a mesoscale, is allowed for (Fig. 13a). Furthermore, microscale growth fibres or offsets of primary components such as grain boundaries or fossil fragments may be visible. Evidence for later shear displacement along an older extension fracture is not uncommon on either the micro- or the mesoscale (e.g. Segall & Pollard 1983a).

(2) *Surface markings.* Two principal types of surface marks on joints are recognized: (a) hackle marks, including plume and herringbone structures (Fig. 13b) and (b) rib marks, also called arrest lines by Bahat & Engelder (1984). Plume structure has attracted most attention, and although there is agreement that it can be used to locate the point of initiation and direction and history of propagation of a fracture there is some disagreement about whether it indicates failure during extension or shear. Roberts (1961) and Syme-Gash (1971) favoured plume formation during shear while Bahat (1979), Engelder (1982b) and Bahat & Engelder (1984) claim that plumes indicate extensional failure. The balance of arguments favours plume formation on extension fractures but it is noteworthy that fringe joints (Hodgson 1961b) at the outer margins of plumes are arranged en

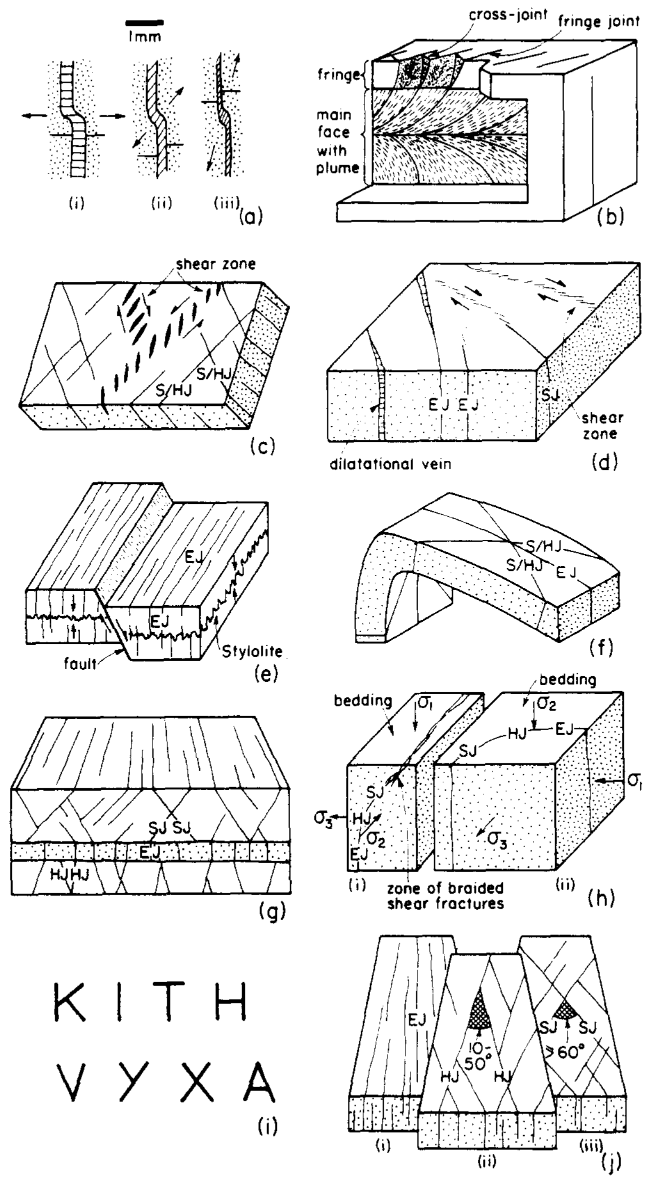

Fig. 13. Criteria for distinguishing between extension, hybrid and shear joints. (a) Microscopic characteristics: matching of crack margins, fibres in microveinlets, offset of a primary component: (i) dilation without shear, (ii) dilation and shear, (iii) shear with subordinate dilation. (b) Plume marks and fringe joints (after Hodgson 1961b, fig. 1). (c) Parallelism with a kinematic indicator, e.g. shear or hybrid joints and shear zones containing en échelon veins. (d) Continuity with a kinematic indicator; e.g. shear joint and a shear zone containing an oblique foliation; extension joint and a dilational vein; extension joint and en échelon cracks in an array. (e) Symmetry with respect to kinematic indicator, e.g. extension joints with reference to a normal fault and bed-parallel stylolites. (f) Symmetry of joint sets with respect to the fold containing them. (g) Lithologically controlled joint refraction across layer interfaces. (h) Systematic curvature of a joint plane. (i) σ_1 normal to layering (based on relationships observed near Merrimbula, New South Wales), (ii) σ_2 normal to layering. (i) Joint trace architectural styles characterized by reference to shapes of letters in the Latin alphabet. (j) 2θ ranges for (i) extension joints, (ii) hybrid joints, (iii) shear joints when $\phi = 30°$. EJ, extension joint; HJ, hybrid joint; SJ, shear joint.

échelon and obliquely to the main joint face. That is, their organization is comparable with that of some second-order cracks within fault or shear zones. Bahat & Engelder (1984) have shown that different generations of extension joints in different lithologies bear plumes of contrasting morphology. Cross-joints (*c*-fractures) within a fringe are generally irregular surfaces abutting the fringe joints; in cleaved rocks their orientations are commonly controlled by a micro-fabric.

(3) *Parallelism with a nearby kinematic indicator.* It is tempting to interpret a joint parallel to a kinematic indicator as belonging to the same failure class, but because the joint might be of a different age it is difficult to be certain about this. The conjugate joints illustrated in Fig. 13(c) are parallel to nearby conjugate shear zones (containing en échelon veins) and thus they could be interpreted as shear or hybrid-shear fractures. Joints parallel to extensional veins are more readily interpreted as extension fractures.

(4) *Continuity and parallelism with a kinematic indicator.* Reports of faults passing uninterruptedly into joints are rare, but Simpson (1983, fig. 7) has illustrated shear zones containing a secondary foliation degrading into joints (e.g. Fig. 13d). Dilational veins passing into extension joints are, however, abundant (e.g. Fig. 13d). A common form of continuity indicative of the likelihood of a joint being an extension fracture is its passage into a suite of en échelon cracks parallel to the joint but contained within an array oblique to it (Fig. 13d). The array is interpreted as following a hybrid or shear direction. Care should be taken to establish whether en échelon cracks are independent structural elements or whether they are part of a fringe assemblage bordering a main joint (Fig. 13b).

(5) *Symmetry with respect to brittle and allied kinematic indicators.* The symmetry of a joint set with respect to related kinematic indicators may be suggestive of its failure class. For example, in Fig. 13(e) the horizontal beds are cut by a steep normal fault and contain a layer-parallel stylolite in addition to a vertical set of joints striking parallel to the fault. The joints possess the orientation predictable for extension fractures accompanying a normal fault and horizontal stylolites.

(6) *Symmetry with respect to folds.* A pair of sets comprising approximately coeval fractures symmetrically related to the fold containing them is most readily interpreted as consisting of conjugate shear or hybrid joints, whereas joints normal to a fold hinge are probably extension fractures (Fig. 13f).

(7) *Joint refraction at an interface between contrasting lithologies.* Provided σ_2 was not perpendicular to layering during failure, joints can be refracted at bedding planes as a consequence of neighbouring beds of contrasting lithology being characterized by different ϕ angles (Fig. 13g).

(8) *Curviplanar joints.* Some joints within a single lithology display a systematic curvature that can be explained by interpreting the different sectors as representing different parts of a continuum of failure classes from extension to shear fracture. The joints illustrated in the two parts of Fig. 13(h) are consistent with their development in stress fields orientated symmetrically with reference to layering so that in part (i) the σ_1 axis was normal to bedding while in part (ii) it was layer-parallel. Possible reasons why failure mode might change from shear to extension fracture via the hybrid class are that the original value of σ_1–σ_3 varied throughout a layer, or that with time it decreased in the direction of crack propagation. As Fig. 13(h, part i) shows a single curviplanar joint surface can pass into a zone of braided fractures, a morphology that Engelder (1974) regards as diagnostic of shear failure.

(9) *Fracture-system architecture.* The architectural style of a joint system, as defined here, describes the spatial relationships of neighbouring surfaces and in plan or profile is readily visualized from the pattern of their traces, which can be characterized by reference to the shapes of capital letters in the Latin alphabet (Fig. 13i). Extension joints initiated in a nearly hydrostatic stress field display a mud-crack geometry (Engelder 1982b) and thus a K-shaped pattern of traces results, younger joints abutting older ones, commonly at right angles immediately adjacent to the butt. Unidirectional extension jointing gives rise to an I-shaped pattern, whereas two episodes of orthogonal systematic extension jointing yield a T-shaped pattern (position 1 in Fig. 14, Fig. 15a), again the younger trace abutting the older. If the later phase of orthogonal extension jointing involved the formation of non-systematic cross-fractures they will be short and hence an H-shaped pattern results (position 2 in Fig. 14). Conjugate joints generally make V-, Y- or X-shaped patterns (positions 3a–3b, 4 and 5, respectively in Fig. 14, Figs. 15 b & c and 16 a–c). The trace of a joint defining the 'arm' of a Y-shaped pattern may not abut the 'trunk', but die out within 1–2 cm of it (position 4 in Fig. 14), perhaps as a result of the node being a heterogeneity that acted as a nucleus from which fracture propagation started. Care should be exercised when interpreting X-patterns that the 'X' is not an artefact of crossing but unrelated joints (position 6 in Fig. 14). Non-systematic but relatively regular cross-joints superimposed on older conjugate joints give an A-shaped pattern (position 7 in Fig. 14, Fig. 15c). Note that both conjugate shear and conjugate hybrid joints define V-, X- and Y-patterns and that the distinction between shear and hybrid joints is based on 2θ angle. V-, Y- and X-shaped patterns also arise when members of an extensional set intersect or abut shear or hybrid joints in a genetically related system (position 8 in Fig. 14). Other criteria, such as symmetry or parallelism with nearby kinematic indicators, are then required to differentiate between the interpretations. It must also be emphasized that individual examples of T-, H-, V-, Y- and X-shaped patterns can be misleading but where they are repeated

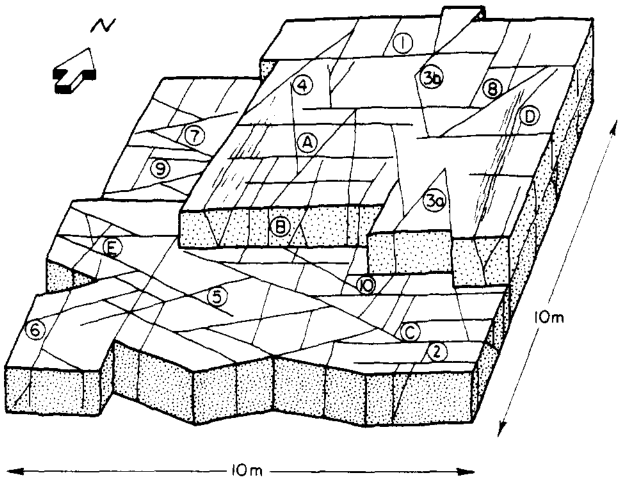

Fig. 14. Ten sets of joints cutting horizontal beds at an imaginary exposure. The relationships depicted are loosely based on geometries observed in central Arabia. For interpretation see text.

throughout a station they are more likely to be of significance. Pattern recognition is difficult where average joint separation is unusually great or the area of an exposure is small compared with average joint separation.

Adjacent beds in a multilayer sequence of contrasting lithologies may display contrasting architectural styles. For example, K-patterns in beds that did not act as stress guides may alternate with regular patterns in more rigid units. If the joints in some lithologies are different in age and origin to those in others there may be differences in architecture related to certain sets not being represented in all lithologies. Joints limited to certain lithologies will be of confined height and give rise to an effect which Engelder (1985) calls joint containment. The occlusion of individual joint planes and sets from the upper layer in a two-layer model is illustrated in Fig. 14 which shows sets striking E–W or enclosing an E–W acute bisector limited to the lower bed while sets striking N–S or enclosing a N–S acute bisector cut both layers. The WNW-striking joint that tracks from point 9 to point 10 in Fig. 14 is a vivid example of a joint occluded from the upper layer, while by contrast the joint planes enclosing a V-shape at position 3a leak down across the interface. A scenario explaining the assemblage of joints illustrated in Fig. 14 is given in the section on multiphase jointing.

(10) *Dihedral* (2θ) *angles between conjugate joints*. As explained earlier, the alleged range of 2θ values between conjugate joints is from 90° or greater to a few degrees. A range of natural 2θ values is to be anticipated knowing that ϕ varies with lithology and that there is a transition from single extension fractures to paired shear fractures (see Fig. 1). For a ϕ angle of 30° the following classification of joints based on 2θ angles can be applied. Extension joints occur as single sets enclosing a notional 2θ angle of 0° (Fig. 13j, part i); conjugate hybrid joints enclose small dihedral angles, say; for practical purposes it is possible to recognize them where 2θ values are in the range 10–50° (Fig. 13j, part ii); and conjugate shear joints enclose angles of 60° or greater (Fig. 13j, part iii). For different ϕ angles different 2θ ranges will apply. In the author's experience conjugate hybrid joints enclosing 2θ angles of 35–45° are common (e.g. Figs. 15c and 16c), and comparable moderate 2θ angles also characterize some conjugate mesofaults and shear zones. Despite the practical difficulties of detecting them some conjugate joint sets enclose 2θ angles of 20° or less (e.g. Figs. 15b and 16a). It should also be stressed that extension joints are equally, if not more, common than hybrid joints. For example, of the 1100 joint orientations measured in the horizontal rocks of the Qaradan segment of the central Arabian graben and trough system (Hancock & Al Kadhi 1982), 41% of the surfaces were classified as extension fractures while 16% were classified as belonging to two conjugate systems of hybrid joints. The remaining joints comprise either conjugate shears (28%) or fractures that were difficult to classify with certainty. Shear joints are especially common in systems consisting of fractures at moderate angles to bedding (e.g. Fig. 16b).

An implication of general significance following from the observation that many vertical or steeply inclined joints are extension or hybrid fractures is that during jointing in brittle sedimentary rocks the value of σ_3' is commonly negative, possibly as a consequence of high fluid pressures, and that $(\sigma_1'-\sigma_3')$ is small (Fig. 1, Table 1). Etheridge (1983) came to a similar conclusion for failure conditions during the development of syn-metamorphic veins.

Allocating joints to sets in the field

Except where joint separation is so great that it is difficult to see several surfaces at once, the allocation of a surface to a set should be accomplished in the field on the basis of visible angular relationships between neighbouring joint planes. Because more than ten sets can be present at a station and because 2θ between conjugate sets can be as small as 10° it is unrealistic to attempt to discriminate between sets solely on the basis of orientation statistics. This is especially so where the data are collected from a large volume of rock or the dispersion about the mean orientation of a set is equal to, say, half the 2θ value between two sets. Consider for example the ten sets illustrated in Fig. 14. By noting the angular relationships between surfaces in the neighbourhoods of subareas A–E (Fig. 14), where several sets are exposed either in plan or profile, an observer would be unlikely to confuse the vertical joints belonging to the conjugate system enclosing the N–S acute bisector with those enclosing the E–W bisector. On a point diagram the four sets would probably yield only two maxima as a consequence of the dispersion of individual orientations about the mean orientation of each set.

Discrimination between sets can also be achieved, in some instances, by considering their morphology or an aspect of their architecture such as abnormally high-frequency. For example, near subarea D in Fig. 14 closely spaced planes in a joint-swarm corridor locally

Fig. 15. Tesselated bedding-plane pavement in Jurassic limestones crossed by T- and H-shaped traces of vertical extension joints developed during a multiphase sequence of jointing. 80 km west of Riyadh, Saudi Arabia. Scale rule is 25 cm. (b) Small dihedral angle conjugate joint defining a V-shaped pattern of traces on a bedding plane in Jurassic limestones. One joint zone comprises an array of en échelon cracks at a very small angle to the array. 150 km south of Riyadh, Saudi Arabia. Pen is 13 cm long. (c) X-shaped pattern of vertical conjugate hybrid joints exposed as traces on a bedding plane in Cretaceous limestones 175 km NNW of Riyadh, Saudi Arabia. Non-systematic cross-joints connect the systematic joints to form A-shaped patterns. Hammer is about 30 cm long.

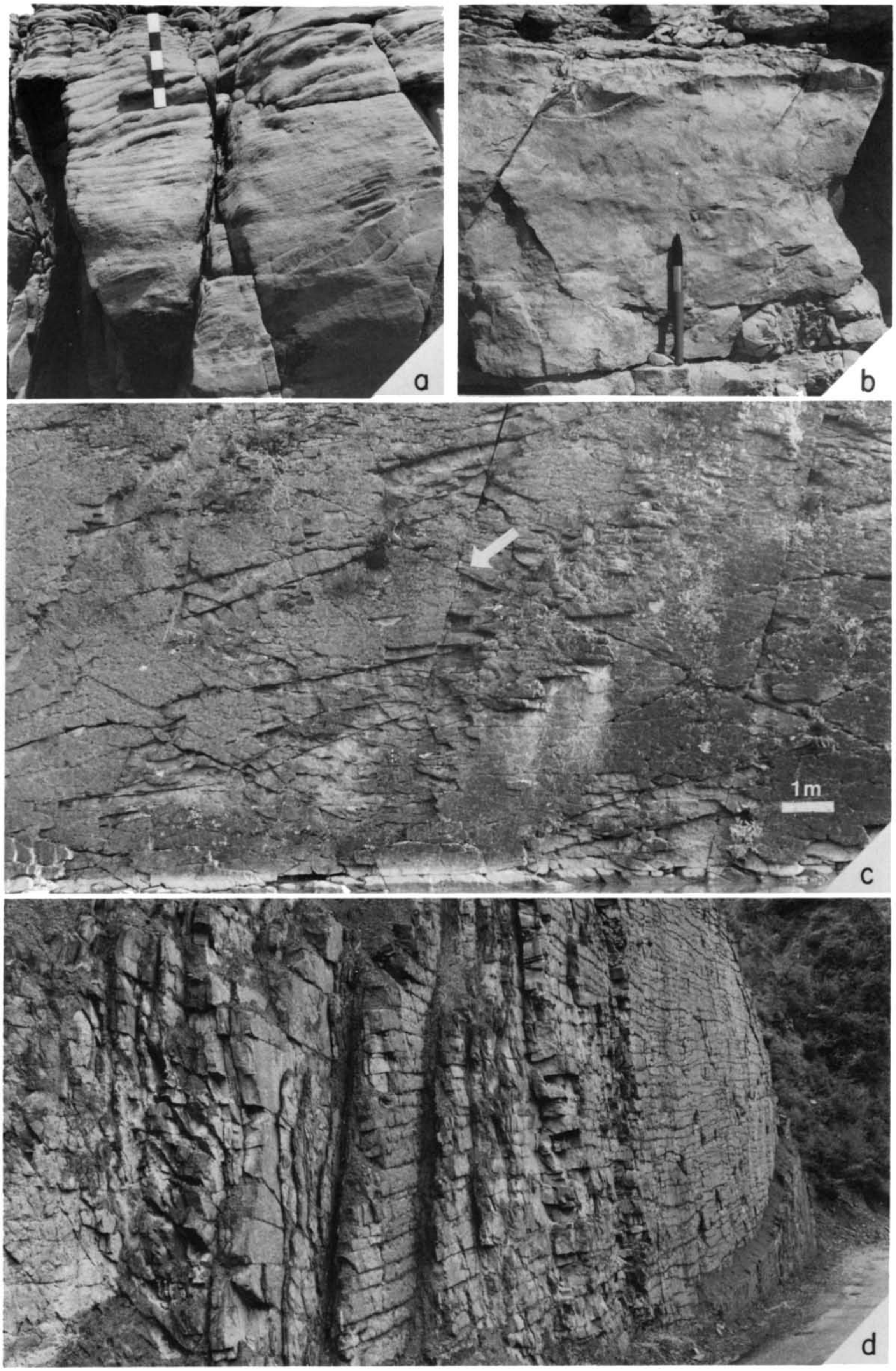

Fig. 16. (a) Steeply inclined hybrid joints enclosing a small dihedral angle and making an X-shaped pattern in Triassic sandstones about 80 km west of Riyadh, Saudi Arabia. Rule is 25 cm long. (b) Y-shaped patterns of moderately inclined shear joints enclosing a dihedral angle of 60° in Jurassic limestones about 125 km south of Riyadh, Saudi Arabia. Pen is 13 cm long. (c) X-, Y- and V-shaped traces of conjugate *hk*0 hybrid joints enclosing an acute dihedral angle about layer strike (subparallel to the *b* fabric axis) exposed in plan on a nearly vertical bedding plane in Palaeogene molassic sandstones about 25 km southwest of Jaca, Spanish Pyrenees. One *hk*0 surface is a small mesofault displacing an older *ac* joint (arrow). (d) Set of pervasive extension fractures in *bc* cutting vertical beds of Palaeogene molassic sandstones about 25 km southwest of Jaca, Spanish Pyrenees. Some of the fractures were transformed into small slip surfaces during the closing stages of folding. Height of exposure is about 3 m.

characterize the steeply inclined N–S striking sets. The general significance of joint-swarm corridors is uncertain: some are closely associated with faults (Pohn 1981) while others are unrelated to known surface structures. Combined fault and joint swarms have been mapped in the Sydney Basin by Shepherd *et al.* (1981).

Multiphase jointing

Even where there are no more than two orthogonal sets of joints there is commonly clear field evidence from butting relationships that each set evolved during a multiphase failure sequence (Fig. 15a). Thus it is rarely possible to state that a certain set contains joints that are universally younger than those in another (e.g. Hancock & Kadhi 1978, Engelder & Geiser 1980), although it may be possible to establish amongst sets the order in which the oldest joints in each set developed. Consider from the perspective of fracture sequence the 10 imaginary joint sets illustrated in Fig. 14. On the basis of butting relationships exposed in plan on the top of the lower bed it can be concluded that NNW- and NNE-striking vertical joints and N–S striking steep joints are younger than WNW- and ENE-striking vertical joints and E–W striking steep joints. N–S vertical joints generally abut E–W joints, but near the subarea marked by a 'C' (Fig. 14) an E–W joint abuts a N–S one, itself abutting an E–W joint. In the upper bed all but one of the E–W striking joints is butted or cut by joints in the other sets. Thus with the exception of both the N–S and E–W vertical joints, that appear to have been initiated throughout the failure sequence (the E–W ones mainly earlier), the remainder of the sets are related to an early phase of N–S stretching and a late phase of E–W stretching.

A general inference following from the widespread occurrence in many platforms of multiphase, but single generation sets of orthogonal joints (e.g. Fig. 15a) is that in the past there have been many extensional failure sequences related to bulk strains involving approximately synchronous horizontal elongation in two directions (cf. the multifault sets of Reches 1983) (Fig. 2).

The imaginary sets shown in Fig. 14 are symmetrical about N–S and E–W axes in both beds, but such a high degree of symmetry in adjacent beds or a region, although common, is not universal because with time the principal stress plane can rotate (Engelder 1985). The evolution of a complete joint pattern involves the progressive infilling of increasingly smaller blocks and in some settings the attitudes of older joints control those of younger ones.

Timing of jointing with respect to tectonism

Because the horizontal rocks of platforms, including their youngest formations (e.g. Babcock 1973, Hancock *et al.* 1984), are often cut by joints of uniform orientation throughout the sequence it is clear that joint initiation can precede folding and that the time gap between sedimentation and jointing can be relatively short (e.g. Cook & Johnson 1970). Joints in a platform that is subsequently incorporated in a deformation belt may, if they are suitably orientated, become the sites of later slip, dilation or pressure solution (e.g. Marshak *et al.* 1982, Winslow 1983) and thus cease to be joints as defined here. Where former extension joints are transformed into shear planes or solution surfaces their initial character will be difficult to establish.

The timing of jointing with reference to related faulting is uncertain, many authors claiming that jointing generally precedes faulting (e.g. Shepherd & Huntington 1981, Segall & Pollard 1983a). Although many fault and joint sets share the same strike it is less common for joint planes to be parallel in both strike and dip to normal or thrust faults, an observation also indicative of a likely age gap between jointing and faulting. Faults can be boundaries between joint domains characterized by contrasting patterns and frequencies (Shepherd & Huntington 1981).

Although some joints are pre-folding structures others are younger than the folds containing them (Hancock 1964, Rixon *et al.* 1983), and a single fold can contain both early- and late-formed joints. The principal field criteria for recognizing joints that were initiated after folding is that they cut associated cleavage surfaces and that bedding planes are not offset where they are intersected by the joints. If some joints in a set pre-date the last episode of fold amplification many of them will become active in shear (Fig. 16d) or the loci for veining or pressure solution. Price (1966) has explained the high degree of symmetry between post-tectonic joint sets and folds by proposing that the orientations of the stored stresses responsible for jointing weakly mirror those of the earlier tectonic phase. Engelder (1985) suggests that extension joints are formed in response to a variety of processes that start early in the history of a basin and continue through uplift and unroofing.

MICROTECTONIC SEQUENCE

As Letouzey & Trémolières (1980) have argued, mesofracture assemblages can be dated by relating directions of shortening inferred from structures within a sequence to its stratigraphy. The critical aspect is not whether a particular structural type or direction is represented within a bed but rather whether there is a change in the number and directions of inferred shortening axes among horizons. Although dating depends on complete stratigraphic sequences they are rarely available and it is generally simpler to establish a sequence of events. The principal criteria for determining microtectonic sequence are abutting and overprinting relationships and the offset of one structure by another. Because within small volumes (10's of m^3) of rock there can be 10 or more sets of brittle mesostructures, it follows that the development of early formed discontinuities does not necessarily inhibit the initiation of new failure surfaces, sometimes at small angles to older ones. By contrast, many macrofaults display evidence of repeated reshearing in a stress field orientated obliquely to that in which the fracture was initiated (McKenzie & Jackson 1983).

CASE STUDIES

Thrust-fold belts

The geometrical relationships between mesofractures and folds have attracted the interest of numerous workers, notable contributions being those of Stearns (1964), Muecke & Charlesworth (1966), N. J. Price (1966), R. A. Price (1967), Groshong (1975), Nickelsen (1979) and Marshak *et al.* (1982). Mesofracture sets in many multilayer sequences are symmetrically arranged with respect to the orientation of the layer containing them and the plunge of an adjacent fold hinge line. Small folds and rocks containing a well-developed axial-plane cleavage are commonly cut by mesofractures symmetrical about axial planes and axes (Price 1966). Winsor (1979) has shown that sedimentary anisotropy controls the arrangement of joints in some folds and likewise a cleavage fabric can also control the orientation of some late-formed joints (Engelder 1985).

Where a system is symmetrical with respect to folds the orientations of sets will change from locality to locality according to variations in plunge, layer dip or axial-plane attitude (also see Dunne in press). Thus in order to describe fracture geometry it is useful to employ a notation not dependent on absolute orientation. A commonly used method refers the geometry of a set or system to an orthogonal fabric cross, the axes of which are labelled *a*, *b* and *c* without kinematic or dynamic implications (Turner & Weiss 1963). Planes perpendicular to one axis but containing the other two are said to be in *ab*, *ac* or *bc* after the axes they contain, whereas surfaces oblique to two axes but containing one are in 0*kl*, *h*0*l* and *hk*0, and those oblique to all three axes are in *hkl*; *h*, *k* and *l* referring to notional intercepts on the *a*, *b* and *c* axes, respectively and O indicating parallelism to an axis (e.g. Price 1967). In the examples illustrated in Fig. 17 the fractures at each point throughout the fold are symmetrically arranged about the dip of layers and the hinge line, and hence bedding is defined as the *ab* fabric plane with *b* parallel to the hinge. As Fig. 17 shows there may be shear, dilation or shortening on surfaces in *ab*, whereas across those in *bc* or *ac* evidence of shear is less common. Many 0*kl*, *h*0*l* and *hk*0 surfaces belong to conjugate systems of shear or hybrid fractures. The development of mesofracture sets and systems symmetrically orientated about sedimentary layering and hinge lines leads to shortening or elongation parallel or normal to those directions. Some late-stage contraction and extension faults are asymmetrical about layering although they intersect each other parallel to nearby hinge lines. Hancock & Atiya (1979) reported that in the Mount Lebanon anticlinorium the acute bisector between contraction faults is refracted away from bedding and towards the horizontal, while the acute bisector between extension faults is refracted towards the vertical.

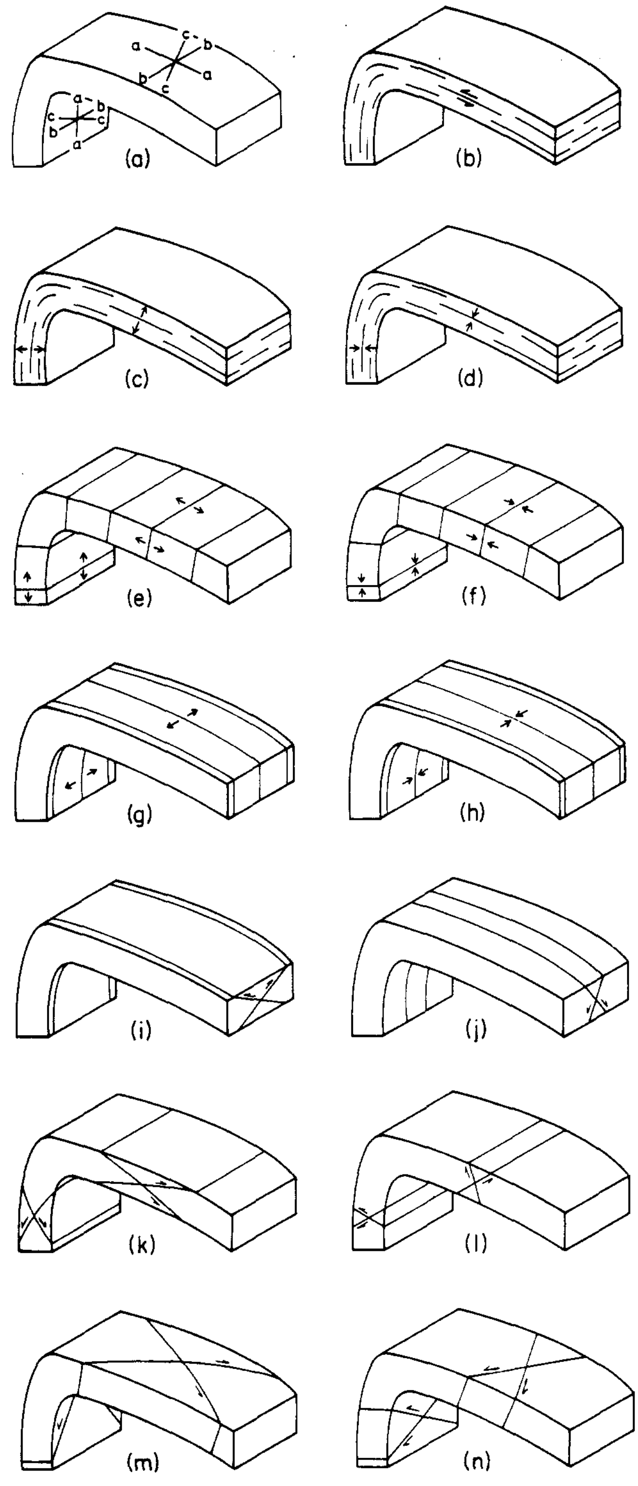

Fig. 17. Block diagrams illustrating mesofracture sets and systems symmetrically arranged with respect to sedimentary layering and fold hinge lines. (a) Definition of the fabric axial cross. (b) Shear surfaces in *ab*. (c) Extension fractures in *ab*. (d) Stylolites in *ab*. (e) Extension fractures in *bc*. (f) Stylolites in *bc*. (g) Extension fractures in *ac*. (h) Stylolites in *ac*. (i) Conjugate 0*kl* fractures enclosing an acute angle about *b*. (j) Conjugate 0*kl* fractures enclosing an acute angle about *c*. (k) Conjugate *h*0*l* fractures enclosing an acute angle about *a*. (l) Conjugate *h*0*l* fractures enclosing an acute angle about *c*. (m) Conjugate *hk*0 fractures enclosing an acute angle about *a*. (n) Conjugate *hk*0 fractures enclosing an acute angle about *b*.

The value of investigating mesofractures in thrust-fold belts is that they can supply additional evidence about the bulk strains experienced during deformation. Mesofractures restricted to certain parts of a fold may be related to local responses, for example elongation or

Table 2. Common mesofractures and allied structures in large amplitude folds within the thrust-fold zone of the Variscan externides in southwest Wales (Pembrokeshire)

Geometry	Styles
bc set	disjunctive solution cleavage
ac set	extension joints, veins
conjugate sets in *h0l* enclosing an acute angle about *a*	mesofaults, shear zones, kinks, hybrid and shear joints
conjugate sets in *h0l* enclosing an acute angle about *c*	mesofaults, shear zones, kinks, hybrid and shear joints
conjugate sets in *hk0* enclosing an acute angle about *a*	shear zones, shear veins, hybrid joints
conjugate sets in *0kl* enclosing an acute angle about *c*	mesofaults, shear veins, hybrid joints

Table 3. Common mesofractures in the Guarga synclinorium, external southwestern Pyrenees

Geometry	Styles
bc set	pervasive extension joints
conjugate sets in *hk0* enclosing an acute angle about *b*	hybrid joints, rare mesofaults
conjugate sets in *h0l* enclosing an acute angle about *c*	mesofaults, hybrid and shear joints
ac sets	extension joints

shortening above or below a neutral surface, or in a leading or trailing limb. As layers pass over frontal, oblique and/or lateral thrust-ramps they experience complex strains and hence mesofractures related to passage over one ramp may be superimposed on those related to movement over another of different orientation.

Two regions investigated by the author are used to illustrate how contrasting mesofracture assemblages associated with folds reflect different bulk strain histories and hence deformation in different environments. In both settings the sets and systems are symmetrical with reference to the attitude of the layers containing them and the plunge of an adjacent fold hinge line (Fig. 17).

(1) *Variscan southwest Wales*. In this contractional tectonic setting the structures cut Palaeozoic rocks in the outermost thrust-fold zone of the Northwest European Variscides (Hancock *et al.* 1982, 1983). Table 2 summarizes the mesofracture sets present in some large-amplitude folds. Reference to Fig. 17 shows that with the exception of the sets in *h0l* at an acute angle to *c*, the assemblage is one which indicates there was layer-parallel shortening normal to fold hinge lines and/or axial elongation. Thus the bulk strains inferred from the majority of mesofractures also point to deformation having occurred in an essentially contractional environment. Extension faults and allied structures in *h0l* are interpreted as local products of stretched fold limbs. Some mesofractures in smaller amplitude folds are symmetrical about axial-planes, and systems on opposed fold limbs are not everywhere identical (Hancock *et al.* 1983).

(2) *Alpine external southwestern Pyrenees*. The Guarga synclinorium is a composite downfold containing 4000 m of late Eocene–Oligocene molasse within the Jaca basin of the southwestern Spanish Pyrenees (Puigdefabregas 1975). During thrusting the basin was deformed into nearly upright folds during the Oligocene to earliest Miocene. Table 3 catalogues the commonest mesofractures. The overwhelming proportion are pervasive *bc* extension joints (Fig. 16d) which within synclines decrease in abundance as limbs decrease in dip towards hinge zones. Of the remaining structures, conjugate *hk0* fractures enclosing an acute angle about *b* are locally common and a few of these are mesofaults displacing older, but rare, systematic *ac* joints (Fig. 16c). The first three sets or systems listed in Table 3 are the most abundant and their collective dominance indicates a bulk strain regime involving layer-parallel elongation normal to hinge lines (Fig. 17), an interpretation consistent with the observation that the folds are growth folds (Puigdefabregas 1975). Older, and now more steeply inclined, layers will have been most stretched.

Platforms

Microtectonic investigations of platforms are especially important because such terrains commonly lack kinematically significant large-scale structures, and their mesofracture suites are more likely to be related to far-field stresses than those of thrust-fold belts, transcurrent fault zones or graben fields. As Engelder (1982a) and Holst (1982) have shown, some extension joint sets in the northeastern U.S.A. strike parallel to the present-day direction of σ_1 and hence elsewhere the analysis of equivalent sets could provide a cheap method of assessing directional variations in the contemporary stress field.

Notable examples of brittle microtectonic studies in platforms are those of Hodgson (1961a), Norris (1967), Babcock (1973, 1974), Roberts (1974), Choukroune (1976), Reches (1976), Engelder & Geiser (1980), Letouzey & Trémolières (1980), Holst & Foote (1981), Shepherd & Huntington (1981) and Eyal & Reches (1983). In this account, mesofractures in the central and eastern parts of the Arabian platform are described to demonstrate how they reflect the influence of both intraplate and plate boundary processes. Where fully developed the complete mesofracture pattern, comprising 95% joints, consists of ten sets (Hancock & Kadhi 1978, 1982) but of these five are dominant (Fig. 18) and occur throughout most of central Arabia (Fig. 19). They can be classified into one set of extension fractures and two systems of conjugate surfaces, the majority of which are hybrid fractures (Fig. 18). From the dominant sets at any given station an identical direction of σ'_3 can be inferred. As Fig. 19 shows it is possible to recognize three overlapping joint domains. (1) In the domain of the central Arabian arch either the joints strike normal to the local trend of beds or the acute bisector between conjugate vertical sets is normal to that trend. Thus the inferred extension direction swings parallel to the curve of outcrops in the arch. (2) In the domain of the central Arabian graben system the strike of the joints or the horizontal direction of the acute bisector is everywhere parallel to the trend of

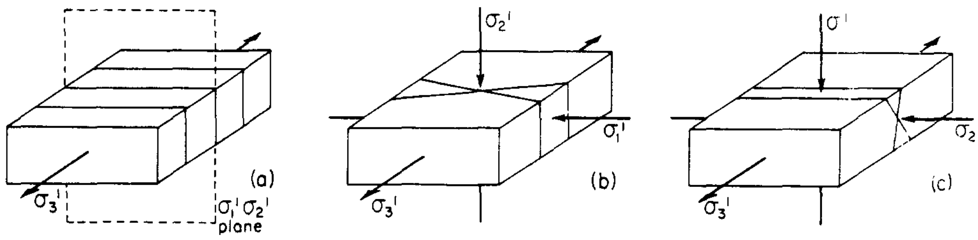

Fig. 18. Block diagrams illustrating geometry of dominant joint sets in the central and eastern parts of the Arabian platform. The front of each block is taken to be parallel to the controlling direction within each of the three joint domains. (a) Single set of extension fractures. (b) Vertical sets of conjugate hybrid fractures. (c) Steeply inclined sets of conjugate hybrid fractures. After Hancock & Kadhi (1978, fig. 4). See text for details.

the nearest graben. Hence in the eastern part of the graben system the inferred extension direction is N–S, whereas in the western part of the system it is generally NE–SW. (3) The third domain is defined by uniformly orientated sets either striking NE–SW or enclosing an acute bisector about that direction, which is parallel to Wadi Al Batin, a major lineament trend (Hancock *et al.* 1984). The joints in the Batin domain cut both the Miocene–Pliocene sediments in the east of the region and they are superimposed on the older pattern within the Mesozoic–Palaeogene rocks in the southern sector of the arch domain. Because there is no overlap between joints in the arch and graben system domains they are probably of the same age and coeval with the late Cretaceous–Palaeogene megastructures to which they are geometrically related.

Although a comprehensive account of how the joint domains are related to the tectonic evolution of Arabia is beyond the scope of this paper, the following summary from Hancock *et al.* (1984) outlines the principal conclusions. Joints in the arch domain developed due to strike-parallel stretching when the arch amplified as a consequence of the development of a peripheral bulge triggered by emplacement of the Oman/Zagros ophiolite nappes. Joints in the graben domain are related to local directions of stretching generated during evolution of the fault zone which is coincident with the crest of the arch along part of its length. Joints in the Batin domain are interpreted as being an expression of the lateral extension of the Arabian foreland at about the same time as there was complementary NE–SW shortening in the Zagros thrust-fold belt.

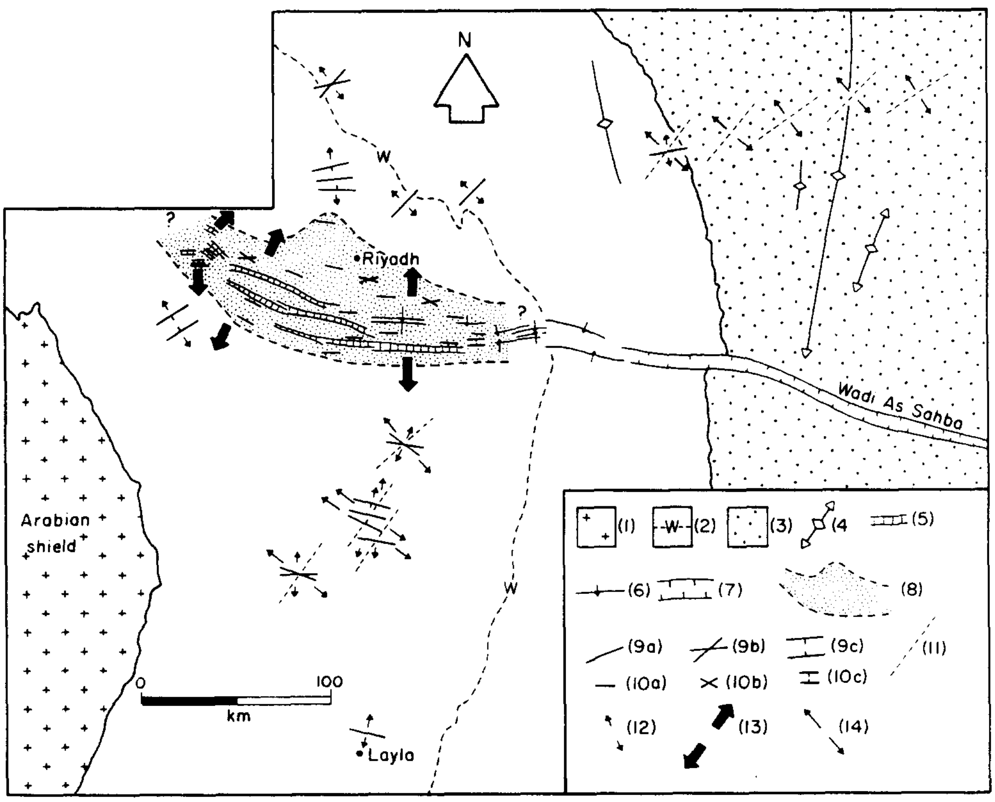

Fig. 19. Mean strikes of dominant joint sets and directions of inferred extension in the central and eastern parts of the Arabian platform. Legend: 1, Precambrian; 2, Permian–Eocene sedimentary rocks (W-sub-Cenomanian unconformity); 3, Neogene sedimentary rocks; 4, anticline; 5, graben; 6, monocline; 7, Wadi As Sahba; 8, joint domain of the central Arabian graben system; 9, strikes of fractures in the joint domain of the central Arabian arch; 10, strikes of fractures in the joint domain of the central Arabian graben system [a, b and c for 9 and 10 refer to the classes illustrated in a, b and c of Fig. 18]; 11, strike of fractures in the Batin joint domain; 12, extension axes in the domain of the arch; 13, generalized directions of extension axes in the domain of the graben system; 14, extension axes in the Batin domain. After Hancock *et al.* (1984, fig. 4).

CONCLUSIONS

(1) The analysis of brittle and semi-brittle mesofractures and allied structures can provide evidence about past and contemporary directions of principal stresses.

(2) Brittle microtectonic investigations are most rewarding in platforms where macrostructures are rare and there have not been significant intraterrain rotations.

(3) The ideal (but rarely realized) suite of mesofractures and allied structures comprises kinematic indicators such as mesofaults, shear zones, arrays of en échelon cracks, kink bands, pressure-solution seams and gaping or filled fissures.

(4) The abundance of joints in platforms makes their analysis essential, especially where kinematic indicators are absent.

(5) An assessment of whether a joint system comprises a set of extension fractures or conjugate sets of hybrid or shear fractures can be made by integrating information concerning: (a) microscopic characteristics, (b) surface markings, (c) parallelism, continuity or symmetry with allied kinematic indicators, (d) lithologically controlled fracture refraction, (e) curviplanar geometry, (f) fracture-system architecture and (g) dihedral angles between conjugate sets. The architectural style of a joint system is a way of describing the spatial relationships of neighbouring surfaces, and is readily visualized in plan or profile from the pattern of their traces which can be characterized by reference to the shapes of capital letters in the Latin alphabet.

(6) In many tectonic settings, especially platforms, the majority of joints are vertical or steeply inclined and comprise sets of extension or hybrid fractures, indicating that during failure the effective minimum stress is generally tensile and that stress differences are usually small.

(7) Jointing is a multiphase process involving the progressive 'filling in' of previously intact blocks. Within a single joint set not all the surfaces are of the same age.

(8) Allocating individual joint planes to sets is an exercise best undertaken in the field on the basis of visible angular relationships.

(9) Joints can be initiated before, during and after folding. Some joints initiated in horizontal rocks that become incorporated in a fold belt and some of those initiated during folding become slip planes, veins or pressure-solution seams.

Acknowledgements—I am most grateful to Dick Glen, Chris Powell, Mike Rickard and other members of the S.G.T.S.G. of the Geological Society of Australia who invited me to address the Bermagui conference. John Creasey and Mike Rickard showed me many fine examples of unusual joints in New South Wales. Some of the ideas expressed in this paper were refined during preliminary presentations at E.T.H. in Zürich and the University of Florence. John Creasey, Terry Engelder, Mike Etheridge and Celâl Şengör critically reviewed the manuscript. Much stimulus, information and invaluable field experience has come from contact with former and present research students, in particular M. S. Atiya, A. A. Barka, T. G. Bevan, W. M. Dunne, T. Erkal, A. Al Kadhi, N. A. Sha'ath, M. E. Tringham and J. P. Turner. General encouragement or financial support have been forthcoming from Sir Kingsley Dunham, Martin Bott, David Dineley, the University of Bristol and the Royal Society of London. The drafting of the figures and the typing of the text for this paper were done with great care and patience by Jean Bees, Alma Gregory and Joyce Rowland. Finally, and most importantly, I should like to thank Neville Price of University College, London, for having been my 'fracture guru' for many years.

REFERENCES

Al Kadhi, A. & Hancock, P. L. 1980. Structure of the Durma–Nisah segment of the central Arabian graben system. *Miner. Resour. Bull. Saudi Arabia* **16**, 1–40.

Anderson, E. M. 1942. *The Dynamics of Faulting*. (1st ed.) Oliver & Boyd, Edinburgh.

Angelier, J. 1984. Tectonic analysis of fault slip data sets. *J. geophys. Res.* **89**, 5835–5848.

Angelier, J. & Colletta, B. 1983. Tension fractures and extensional tectonics. *Nature, Lond.* **301**, 49–51.

Arthaud, F. 1969. Méthode de détermination graphique des directions de raccourcissement d'allongement et intermédiare d'une population de failles. *Bull. Soc. geól. Fr.* **11**, 739–737.

Aydin, A. & Johnson, A. M. 1983. Analysis of faulting in porous sandstones. *J. Struct. Geol.* **5**, 19–31.

Aydin, A. & Reches, Z. 1982. The number and orientation of fault sets in the field and in experiments. *Geology* **10**, 107–112.

Babcock, E. A. 1973. Regional jointing in Southern Alberta. *Can. J. Earth Sci.* **10**, 1769–1781.

Babcock, E. A. 1974. Jointing in central Alberta. *Can. J. Earth Sci.* **11**, 1181–1186.

Bahat, D. 1979. Theoretical considerations on mechanical parameters of joint surfaces based on studies on ceramics. *Geol. Mag.* **116**, 81–92.

Bahat, D. & Engelder, T. 1984. Surface morphology on cross-fold joints of the Appalachian Plateau, New York and Pennsylvania. *Tectonophysics* **104**, 299–313.

Bally, A. W., Bernoulli, D., Davis, G. A. & Montadert, L. 1980. Listric normal faults. *Oceanol. Acta* (*Proc. 26th Int. Geol. Congr. Geology of Continental Margins Symp.*, Paris), 87–101.

Barka, A. A. & Hancock, P. L. 1984. Neotectonic deformation patterns in the convex-northwards arc of the North Anatolian fault zone. In: *The Geological Evolution of the Eastern Mediterranean* (edited by Dixon, J. E. & Robertson, A. H. F.). *Spec. Publs geol. Soc. Lond.* **17**, 763–774.

Bartlett, W. L., Friedman, M. & Logan, J. M. 1981. Experimental folding and faulting of rocks under confining pressure: Part IX: wrench faults in limestone layers. *Tectonophysics* **79**, 255–277.

Beach, A. 1975. The geometry of en-échelon vein arrays. *Tectonophysics* **28**, 245–263.

Bevan, T. G. in press a. A reinterpretation of fault systems in the Upper Cretaceous rocks of the Dorset coast. *Proc. Geol. Ass.*

Bevan, T. G. in press b. Tectonic evolution of the Isle of Wight: a Cenozoic stress history based on mesofractures. *Proc. Geol. Ass.*

Blès, J-L. & Feuga, B. 1981. *La Fracturation des Roches*. B.R.G.M., Orléans.

Brace, W. F. 1964. Brittle fracture of rocks. In: *State of Stress in the Earth's Crust* (edited by Judd, R.). Elsevier, New York, 111–180.

Buchner, F. 1981. Rhinegraben: horizontal stylolites indicating stress regimes of earlier states of rifting. *Tectonophysics* **73**, 113–118.

Carson, B., von Huene, R. & Arthur, M. 1982. Small-scale deformation structures and physical properties related to convergence in Japan Trench slope sediments. *Tectonics* **1**, 277–302.

Choukroune, P. 1969. Un exemple d'analyse microtectonique d'une série calcaire affecté de plis isopaques ("concentriques"). *Tectonophysics* **7**, 57–70.

Choukroune, P. 1976. Strain patterns in the Pyrenean chain. *Phil. Trans. R. Soc.* **A283**, 271–280.

Cook, A. C. & Johnson, K. R. 1970. Early joint formation in sediments. *Geol. Mag.* **107**, 361–368.

Cox, S. F. & Etheridge, M. A. 1983. Crack–seal fibre growth mechanisms and their significance in the development of orientated layer silicate microstructures, *Tectonophysics* **92**, 147–170.

Dennis, J. G. 1972. *Structural Geology*. Ronald Press, New York.

Donath, F. A. 1961. Experimental study of shear failure in anisotropic rocks. *Bull. geol. Soc. Am.* **72**, 985–990.

Donath, F. A. 1964. Strength variation and deformational behaviour in anisotropic rock. In: *State of Stress in the Earth's Crust* (edited by Judd, W. R.). Elsevier, New York, 281–297.

Dunne, W. M. in press. Geometric analysis of brittle and semi-brittle mesostructures in folds. *J. Struct. Geol.*

Durney, D. W. & Ramsay, J. G. 1973. Incremental strains measured by syntectonic crystal growths. In: *Gravity and Tectonics* (edited by De Jong, K. A. & Scholten, R.). Wiley, New York, 67–96.

Elliott, D. 1976. The energy balance and deformation mechanisms of thrust sheets. *Phil. Trans. R. Soc.* **133**, 311–327.

Engelder, J. T. 1974. Cataclasis and the generation of fault gouge. *Bull. geol. Soc. Am.* **85**, 1515–1522.

Engelder, J. T. 1976. Effect of scratch hardness on frictional wear and stick-slip of Westerly Granite and Cheshire Quartzite. In: *The Physics and Chemistry of Minerals and Rocks* (edited by Strens, R. G. J.). Wiley, London, 139–150.

Engelder, T. 1982a. Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America? *Tectonics* **1**, 161–177.

Engelder, T. 1982b. Reply to a comment by A. E. Scheidegger on 'Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America? by T. Engelder. *Tectonics* **1**, 465–470.

Engelder, T. 1985. Loading paths to joint propagation during a tectonic cycle: an example from the Appalachian Plateau, U.S.A. *J. Struct. Geol.* **7**, 459–476.

Engelder, T. & Geiser, P. 1980. On the use of regional joint sets as trajectories of paleostress fields during the development of the Appalachian Plateau, New York. *J. geophys. Res.* **85**, 6319–6341.

Etheridge, M. A. 1983. Differential stress magnitudes during regional deformation and metamorphism: upper bound imposed by tensile fracturing. *Geology* **11**, 231–234.

Eyal, Y. & Reches, Z. 1983. Tectonic analysis of the Dead Sea Rift regions since the late-Cretaceous based on mesostructures. *Tectonics* **2**, 167–185.

Farrell, S. G. 1984. A dislocation model applied to slump structures, Ainsa basin, South Central Pyrenees *J. Struct. Geol.* , 727–736.

Fletcher, R. C. & Pollard, D. D. 1981. Anticrack model for pressure solution surfaces. *Geology* **9**, 419–424.

Friedman, M. & Logan, J. M. 1970. Microscopic feather fractures. *Bull. geol. Soc. Am.* **81**, 3417–3420.

Gamond, J. F. 1983. Displacement features associated with fault zones: a comparison between observed examples and experimental models. *J. Struct. Geol.* **5**, 33–45.

Gay, N. C. 1970. The formation of step structures on slickensided shear surfaces. *J. Geol.* **78**, 523–532.

Gay, N. C. & Weiss, L. E. 1974. The relationship between principal stress direction and the geometry of kinks in foliated rocks. *Tectonophysics* **21**, 287–300.

Gibbs, A. D. 1983. Balanced cross-section construction from seismic sections in areas of extensional tectonics. *J. Struct. Geol.* **5**, 153–160.

Gibbs, A. D. 1984. Structural evolution of extensional basin margins. *J. geol. Soc. Lond.* **141**, 609–620.

Gill, W. D. 1979. Syndepositional sliding and slumping in the West Clare Namurian basin, Ireland. *Spec. Pap. geol. Surv. Ir.* **4**, 1–31.

Groshong, R. H. 1975. Strain, fractures and pressure solution in natural single layer folds. *Bull. geol. Soc. Am.* **86**, 1363–1376.

Hancock, P. L. 1964. The relations between folds and late-formed joints in South Pembrokeshire. *Geol. Mag.* **101**, 174–184.

Hancock, P. L. 1969. Jointing in the Jurassic limestones of the Cotswold Hills. *Proc. Geol. Ass.* **80**, 219–241.

Hancock, P. L. 1972. The analysis of en-échelon veins. *Geol. Mag.* **109**, 269–276.

Hancock, P. L. 1973. Shear zones and veins in the Carboniferous Limestone near the Observatory, Clifton, Briston. *Proc. Bristol Nat. Soc.* **32**, 297–306.

Hancock, P. L. & Atiya, M. S. 1975. The development of en-échelon vein segments by the pressure solution of formerly continuous veins. *Proc. Geol. Ass.* **86**, 281–286.

Hancock, P. L. & Atiya, M. S. 1979. Tectonic significance of mesofracture systems associated with the Lebanese segment of the Dead Sea fault. *J. Struct. Geol.* **1**, 143–153.

Hancock, P. L. & Barka, A. A. 1981. Opposed shear senses inferred from neotectonic mesofracture systems in the North Anatolian fault zone. *J. Struct. Geol.* **3**, 383–392.

Hancock, P. L. & Kadhi, A. 1978. Analysis of mesoscopic fractures in the Dhruma-Nisah segment of the central Arabian graben system. *J. geol. Soc. Lond.* **135**, 339–347.

Hancock, P. L. & Al Kadhi, A. 1982. Significance of arcuate joint sets connecting oblique grabens in central Arabia. *Mitt. Geol. Inst. ETH Univ. Zürich, Neue Folge* **239a**, 128–131.

Hancock, P. L., Al Kadhi, A. & Sha'at, N. A. 1984. Regional joint sets in the Arabian platform as indicators of intraplate processes. *Tectonics* **3**, 27–43.

Hancock, P. L., Dunne, W. M. & Tringham, M. E. 1982. Variscan structures in Southwest Dyfed. In: *Geological Excursions in Dyfed, South-west Wales* (edited by Bassett, M. G.). National Museum of Wales, Cardiff, 214–248.

Hancock, P. L., Dunne, W. M. & Tringham, M. E. 1983. Variscan deformation in south-west Wales. In: *The Variscan Fold Belt in the British Isles* (edited by Hancock, P. L.). Adam Hilger, Bristol, 47–73.

Harding, T. P. 1974. Petroleum traps associated with wrench faults. *Bull. Am. Ass. Petrol. Geol.* **60**, 365–378.

Hobbs, B. E., Means, W. D. & Williams, P. F. 1976. *An Outline of Structural Geology*. Wiley, New York.

Hodgson, R. A. 1961a. Regional study of jointing in Comb Ridge–Navajo Mountain area, Arizona and Utah. *Bull. Am. Ass. Petrol. Geol.* **45**, 11–38.

Hodgson, R. A. 1961b. Classification of structures on joint surfaces. *Am. J. Sci.* **259**, 439–502.

Holst, T. B. 1982. Regional jointing in the northern Michigan Basin. *Geology* **10**, 273–277.

Holst, T. B. & Foote, G. R. 1981. Joint orientation in Devonian rocks in the northern portion of the lower peninsula of Michigan. *Bull. geol. Soc. Am.* **92**, 85–93.

Jackson, J. & McKenzie, D. 1983. The geometrical evolution of normal fault systems. *J. Struct. Geol.* **5**, 471–482.

Jackson, J. A., King, G. & Vita-Finzi, C. 1982. The neotectonics of the Aegean: an alternative view. *Earth Planet. Sci. Lett.* **61**, 303–318.

Jamison, W. R. & Stearns, D. W. 1982. Tectonic deformation of Wingate Sandstone, Colorado National Monument. *Bull. Am. Ass. Petrol. Geol.* **66**, 2584–2608.

Jaeger, J. C. & Cook, N. G. W. 1976. *Fundamentals of Rock Mechanics*. (2nd ed.). Chapman & Hall, London.

Jaroszewski, W. 1972. Microscopic structural criteria of tectonics of non-orogenic areas: an example from the northeastern Mesozoic margin of the Swietokrzyski Mountains. *Stud. Geol. Polon.* **38**, 1–215 [in Polish with English summary].

Knipe, R. J. & White, S. H. 1979. Deformation in low grade shear zones in the Old Red Sandstone, S.W. Wales. *J. Struct. Geol.* **1**, 53–66.

Letouzey, J. & Trémolières, P. 1980. Paleo-stress fields around the Mediterranean since the Mesozoic derived from microtectonics: comparison with plate tectonic data. *Mem. Bur. Rech. Geol. Min.* **115**, 261–273.

Lundberg, N. & Moore, J. G. 1981. Structural features of the Middle America trench slope off southern Mexico, Deep Sea Drilling Project Leg 66. *Initial Rep. Deep Sea Drilling Proj.* **66**, 793–814.

McEwen, T. J. 1981. Brittle deformation in pitted pebble conglomerates. *J. Struct. Geol.* **3**, 25–37.

McKenzie, D. & Jackson, J. 1983. The relationship between strain rates, crustal thickening, palaeomagnetism, finite strain and fault movements within a deforming zone. *Earth Planet Sci. Lett.* **65**, 182–202.

Marshak, S., Geiser, P. A., Alvarez, W. & Engelder, T. 1982. Mesoscopic fault array of the northern Umbrian Apennine fold belt, Italy: geometry of conjugate shear by pressure-solution slip. *Bull. geol. Soc. Am.* **93**, 1013–1022.

Mattauer, M. 1973. *Les Déformations des Matériaux de L'Écorce Terrestre*. Hermann, Paris.

Muecke, G. K. & Charlesworth, H. A. K. 1966. Jointing in folded Cardium Sandstones along the Bow River, Alberta. *Can. J. Earth Sci.* **3**, 579–596.

Muehlberger, W. R. 1961. Conjugate joint sets of small dihedral angle. *J. Geol.* **69**, 211–219.

Mullenax, A. C. & Gray, D. R. 1984. Interaction of bed-parallel stylolites and extension veins in boudinage. *J. Struct. Geol.* **6**, 63–71.

Nelson, R. A. 1981. Significance of fracture sets associated with stylolite zones. *Bull. Am. Ass. Petrol. Geol.* **65**, 2417–2415.

Nickelsen, R. P. 1979. Sequence of structural stages of the Alleghany orogeny, at the Bear Valley Strip Mine, Shamokin, Pennsylvania. *Am. J. Sci.* **279**, 225–271.

Norris, D. K. 1958. Structural conditions in Canadian coal mines. *Bull. geol. Surv. Can.* **44**, 1–53.

Norris, D. K. 1967. Structural analysis of the Queensway folds, Ottawa, Canada. *Can. J. Earth Sci.* **4**, 299–321.

Paterson, M. S. 1978. *Experimental Rock Deformation—The Brittle Field*. Springer, Heidelberg.

Philip, H. & Meghraoui, M. 1983. Structural analysis and interpretation of the surface deformations of the El Asnam earthquake of October 10, 1980. *Tectonics* **2**, 17–49.

Phillips, W. J. 1972. Hydraulic fracturing and mineralisation. *J. geol. Soc. Lond.* **128**, 337–359.

Pickering, K. T. 1983. Small scale syn-sedimentary faults in the Upper Jurassic 'Boulder Beds'. *Scott. J. Geol.* **19**, 169–181.

Pittman, E. D. 1981. Effect of fault-related granulation on porosity and permeability of quartz sandstones, Simpson Group (Ordovician), Oklahoma. *Bull. Am. Ass. Petrol. Geol.* **63**, 2381–2387.

Pohn, H. A. 1981. Joint spacing as a method of locating faults. *Geology* **9**, 258–261.

Price, N. J. 1966. *Fault and Joint Development in Brittle and Semi-Brittle Rock*. Pergamon Press, Oxford.

Price, N. J. 1977. Aspects of gravity tectonics and the development of listric faults. *J. geol. Soc. Lond.* **133**, 311–327.

Price, R. A. 1967. The tectonic significance of mesoscopic subfabrics in the southern Rocky Mountains of Alberta and British Columbia. *Can. J. Earth Sci.* **4**, 39–70.

Puigdefabregas, C. 1975. La sedimentación molásica en la Cuenca de Jaca. *Mon. Inst. Est. Pirenaicos* **104**, 1–188.

Ramsay, J. G. 1967. *Folding and Fracturing of Rocks*. McGraw-Hill, New York.

Ramsay, J. G. 1980a. Shear zone geometry: a review. *J. Struct. Geol.* **2**, 83–99.

Ramsay, J. G. 1980b. The crack–seal mechanism of rock deformation. *Nature, Lond.* **284**, 135–139.

Ramsay, J. G. 1981. Tectonics of the Helvetic Nappes. In: *Thrust and Nappe Tectonics* (edited by McClay, K. R. & Price, N. J.). *Spec. Publs geol. Soc. Lond.* **9**, 293–309.

Ramsay, J. G. 1982. Rock ductility and its influence on the development of tectonic structures in Mountain Belts. In: *Mountain Building Processes* (edited by Hsü, K.). Academic Press, London, 111–127.

Ramsay, J. G. & Graham, R. H. 1970. Strain variation in shear belts. *Can. J. Earth Sci.* **7**, 786–813.

Ramsay, J. G. & Huber, M. I. 1983. *The Techniques of Modern Structural Geology, Volume 1, Strain Analysis*. Academic Press, London.

Reading, H. G. (editor) 1978. *Sedimentary Environments and Facies*. Blackwell, Oxford.

Reches, Z. 1976. Analysis of joints in two monoclines in Israel. *Bull. geol. Soc. Am.* **87**, 1654–1662.

Reches, Z. 1978. Analysis of faulting in three-dimensional strain field. *Tectonophysics* **47**, 109–129.

Reches, Z. 1983. Faulting of rocks in three-dimensional strain fields—II. Theoretical analysis. *Tectonophysics* **95**, 133–156.

Reches, Z. & Dieterich, J. 1983. Faulting of rocks in a three-dimensional strain field—I. Failure of rocks in polyaxial, servo-control experiments. *Tectonophysics* **95**, 111–132.

Rickard, M. J. & Rixon, L. K. 1983. Stress configurations in conjugate quartz vein arrays. *J. Struct. Geol.* **5**, 573–578.

Rixon, L. K., Bucknell, W. R. & Rickard, M. J. 1983. Megakink folds and related structures in the Upper Devonian Merrimbula Group, south coast of New South Wales. *J. geol. Soc. Aust.* **30**, 277–293.

Roberts, J. C. 1961. Feather-fractures and the mechanics of rock jointing. *Am. J. Sci.* **259**, 481–492.

Roberts, J. C. 1974. Jointing and minor tectonics of the Vale of Glamorgan between Ogmore-By-Sea and Lavernock Point, South Wales. *Geol. J.* **9**, 97–114.

Sanderson, D. J. & Marchini, W. R. D. 1984. Transpression. *J. Struct. Geol.* **6**, 449–458.

Scheidegger, A. E. 1982. Comment on 'Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of north America?' by T. Engelder. *Tectonics* **1**, 463–464.

Scheidegger, A. E. 1983. Interpretation of fracture and physiographic patterns in Alberta, Canada. *J. Struct. Geol.* **5**, 53–59.

Segall, P. & Pollard, D. D. 1983a. Nucleation and growth of strike slip faults in granite. *J. geophys. Res.* **88**, 555–568.

Segall, P. & Pollard, D. D. 1983b. Joint formation in granitic rock of the Sierra Nevada. *Bull. geol. Soc. Am.* **94**, 563–575.

Shepherd, J. & Huntington, J. F. 1981. Geological fracture mapping in coalfields and the stress fields of the Sydney Basin. *J. geol. Soc. Aust.* **28**, 299–309.

Shepherd, J., Creasey, J. W. & Huntington, J. 1981. Geometric and dynamic analysis of some faults in the Western Coalfield of New South Wales. *Proc. Australas. Inst. Min. Metall.* **279**, 19–32.

Simpson, C. 1983. Displacement and strain patterns from naturally occurring shear zone terminations. *J. Struct. Geol.* **5**, 497–506.

Stearns, D. W. 1964. Macrofracture patterns on Teton anticline, northwestern Montana. *Trans. Am. geophys. Un.* **45**, 107.

Syme Gash, P. J. 1971. A study of surface features relating to brittle and semi-brittle fractures. *Tectonophysics* **12**, 349–391.

Tchalenko, J. S. 1970. Similarities between shear zones of different magnitudes. *Bull. geol. Soc. Am.* **81**, 1625–1640.

Tchalenko, J. S. & Ambraseys, N. N. 1970. Structural analysis of the Dast-e-Bayaz (Iran) earthquake fractures. *Bull. geol. Soc. Am.* **81**, 41–60.

Tjia, H. D. 1972. Fault movement, reoriented stress field and subsidiary structures. *Pacific Geol.* **5**, 49–70.

Turner, F. J. & Weiss, L. E. 1963. *Structural Analysis of Metamorphic Tectonites*. McGraw-Hill, New York.

Van Loon, A. J., Brodzikowski, K. & Gotowala, R. 1984. Structural analysis of kink bands in unconsolidated sands. *Tectonophysics* **104**, 351–374.

Wernicke, B. & Burchfiel, B. C. 1982. Modes of extensional tectonics. *J. Struct. Geol.* **4**, 105–115.

White, S. H., Burrows, S. E., Carreras, J., Shaw, N. D. & Humphreys, F. J. 1980. On mylonites in ductile shear zones. *J. Struct. Geol.* **2**, 175–187.

Winslow, M. A. 1983. Clastic dike swarms and the structural evolution of the foreland fold and thrust belt of the southern Andes. *Bull. geol. Soc. Am.* **94**, 1073–1080.

Winsor, C. N. 1979. The correlation of fracture directions with sediment anisotropy in folded rocks of the Delamerian fold belt at Port Germein gorge, South Australia. *J. Struct. Geol.* **1**, 245–254.

Yielding, G., Jackson, J. A., King, G. C. P., Sinvhal, H., Vita-Finzi, C. & Wood, R. M. 1981. Relations between surface deformation, fault geometry, seismicity and rupture characteristics during the El Asnam (Algeria) earthquake of 10 October 1980. *Earth Planet. Sci. Lett.* **56**, 287–304.