Journal of Structural Geology, Vol. 12, No. 1, pp. 69 to 81, 1990
Printed in Great Britain

0191-8141/90 $03.00+0.00

The role of pressure solution and intermicrolithon-slip in the development of disjunctive cleavage domains: a study from Helvick Head in the Irish Variscides

FRANCIS X. MURPHY

Department of Geology, University College, Cork, Ireland

(*Received* 20 *November* 1988; *accepted in revised form* 5 *June* 1989)

Abstract—The offset of sedimentary laminae, bedding surfaces, veins and other passive markers across disjunctive cleavage domains has commonly been attributed solely to pressure solution. A detailed study of disjunctive cleavage within siltstones and sandstones from Helvick Head, Co. Waterford, on the eastern margin of the Irish Variscides indicates that the cleavage domains were initially formed by pressure solution processes. However, the study also revealed features which cannot be explained by pressure solution alone. The cleavage domains represent planes of weakness within otherwise relatively homogeneous rock. During fold tightening they were rotated out of parallelism with the principal (XY) plane of the finite strain ellipsoid, thus enabling shear movement between the microlithons. It is considered that this intermicrolithon-slip caused a rotation of the elongate quartz grains within the cleavage domains resulting in an asymmetry of their long axes with respect to the cleavage domain margin, grain microfracturing and occasionally slickensides on cleavage surfaces. Intermicrolithon-slip is unlikely to be a localized phenomenon but is probably widespread throughout the Irish Variscides and other orogenic belts.

INTRODUCTION

THE relationship of cleavage to the finite strain ellipsoid has been a subject of much debate, and is one which has yet to be satisfactorily resolved. It is commonly accepted that cleavage develops parallel to the XY plane of the finite strain ellipsoid (Sharpe 1849, Sorby 1853, Harker 1886, Cloos 1947, Siddans 1972, Wood 1974). Since this is a plane of no finite shear strain it was generally held that slip along cleavage was impossible. Consequently the offset of sedimentary laminae, bedding surfaces, veins and other passive markers across cleavage is generally attributed to the loss of material due to pressure solution alone (Plessman 1964, Nickelsen 1972, Groshong 1976, Gray 1977).

However, during a progressive deformation event it is unlikely that disjunctive cleavage will remain in parallelism with the XY plane. While cleavage may initiate parallel to the XY plane during layer-parallel shortening or in the early stages of folding, as the folds develop, cleavage will behave as a material plane and be rotated out of parallelism with the XY plane. An exception to this may occur within mudrocks in which it may be possible for slaty cleavage to continuously track the XY plane (Siddans 1972, Wood 1974). Hoeppener (1956) noted that when the cleavage orientation is close to a plane of maximum shear stress it is likely that slip can occur along the cleavage at stresses too low to give rise to a second cleavage. Dieterich (1969) also observed that shear could occur parallel to cleavage since resistance to shear along such surfaces is low. In addition Hobbs *et al.* (1976, p. 237) noted that cleavage is commonly a plane of shear strain, and that offsets across it sometimes have a sense of movement which cannot be explained by pressure solution. Ghosh (1982) proposed that large shear strains could be accommodated along foliations which were slightly oblique to the XY plane, while Engelder & Marshak (1985) suggested that if cleavage domains were to behave as material planes, small amounts of slip might occur along them during folding since they would no longer maintain parallelism with the XY plane. They also noted that as yet there has been little documentation of intermicrolithon-slip.

This paper reports the results of a study of disjunctive cleavage from the Irish Variscides. Intermicrolithon-slip has been briefly reported from this fold belt (Coe & Selwood 1963, 1964, Cooper *et al.* 1986, O'Sullivan *et al.* 1986) but it has not been examined in detail. An area on Helvick Head, Co. Waterford, was chosen for this study, with the aim of obtaining evidence for the presence or absence of intermicrolithon-slip, and if present, its origin and role in the evolution of the cleavage.

THE GEOLOGY OF HELVICK HEAD

Lithostratigraphy

Helvick Head in Co. Waterford, southern Ireland, forms a small peninsula which extends eastwards into the Celtic Sea (Fig. 1). It is located close to the eastern margin of the Munster Basin, a thick accumulation of Devonian–Carboniferous sedimentary rocks. A condensed sequence of Old Red Sandstone is well exposed on Helvick Head (Fig. 2). The lower part of the Old Red Sandstone is predominantly composed of strongly cleaved siltstones and fine sandstones interbedded with conglomerates (Comeragh Conglomerate Formation). A sequence dominated by siltstones and mudstones (Ballytrasna Formation) occurs above this. The top of the Old Red Sandstone is composed of coarse sandstones interbedded with fine siltstone (Kiltorcan Formation).

Structure

The Helvick Peninsula lies on the southern limb of the Dungarvan Syncline, a broad, open, upright Variscan fold (Murphy 1985). Its steeply-dipping southern limb is disrupted by two imbricate forethrusts which are considered to root into an underlying sole thrust (Cooper *et al.* 1984, 1986). The southern thrust dies out eastwards and is not present on the peninsula (Fig. 1). The northern thrust, estimated from borehole evidence to have a minimum displacement of 500 m in the west, disappears offshore close to Mweelahorna on the northern side of the peninsula. A number of second-order, N-verging, W-plunging, close, asymmetrical anticline–syncline pairs with half-wavelengths of 10–30 m and trends of 270–280° (Fig. 3a) occur along the northern side of the peninsula. However, to the south second-order folds are absent apart from one anticline–syncline pair at Muggort's Bay (Fig. 2). A N–S cross-section through Helvick Head is shown in Fig. 4.

A weak to strong, rough to smooth disjunctive cleavage occurs within the fine sandstones and siltstones, while the mudstones contain a coarse continuous or very closely spaced smooth disjunctive cleavage (in the terminology of Powell 1979 and Borradaile *et al.* 1982). Engelder & Marshak (1985) considered that a minimum of 10% clay was required to enable the initiation of cleavage. The importance of clay is well illustrated in the lithologies on Helvick Head. The conglomerates of the Comeragh Conglomerate Formation have a clay-rich matrix and have developed a weak to moderate, rough disjunctive cleavage (Fig. 5a). However the coarse sandstones of the Kiltorcan Formation are uncleaved due to the low proportion of clay within their matrix (5–10%). Although cleavage is generally axial-planar to the Dungarvan Syncline (Fig. 3b) the transection of minor folds occurs locally and dextral Δ values (Borradaile 1978) up to 30° have been recorded between Ballynagaul and Crow's Point (Fig. 2).

The limbs of the Dungarvan Syncline are cross-cut by numerous approximately N–S-trending dextral wrench faults and N–S-trending dextral tension-gash arrays. These structures are considered to have been produced by a horizontal N–S dextral shear stress which is attributed to sticking on the sole thrust in the east due to the resistance of the Leinster Massif to thrust propagation (Murphy 1985, 1988).

ZONAL CLEAVAGE MORPHOLOGY

On Helvick Head, and elsewhere throughout the Dungarvan Syncline, the disjunctive cleavage developed within the coarse siltstones and sandstones is widely spaced, particularly within the Ballytrasna Formation and the Ballyquin Member of the Kiltorcan Formation. It resembles the rough cleavage of Dennis (1972) and Gray (1978) and the cleavage zones of Nickelsen (1972). Similar structures have been termed deformation zones by Boyer (1984) and *P–Q* fabrics by Waldron & Sandiford (1988). This zonal cleavage is considered to be primarily a product of pressure solution processes (Borradaile 1982). However, some of its features cannot be explained by pressure solution alone.

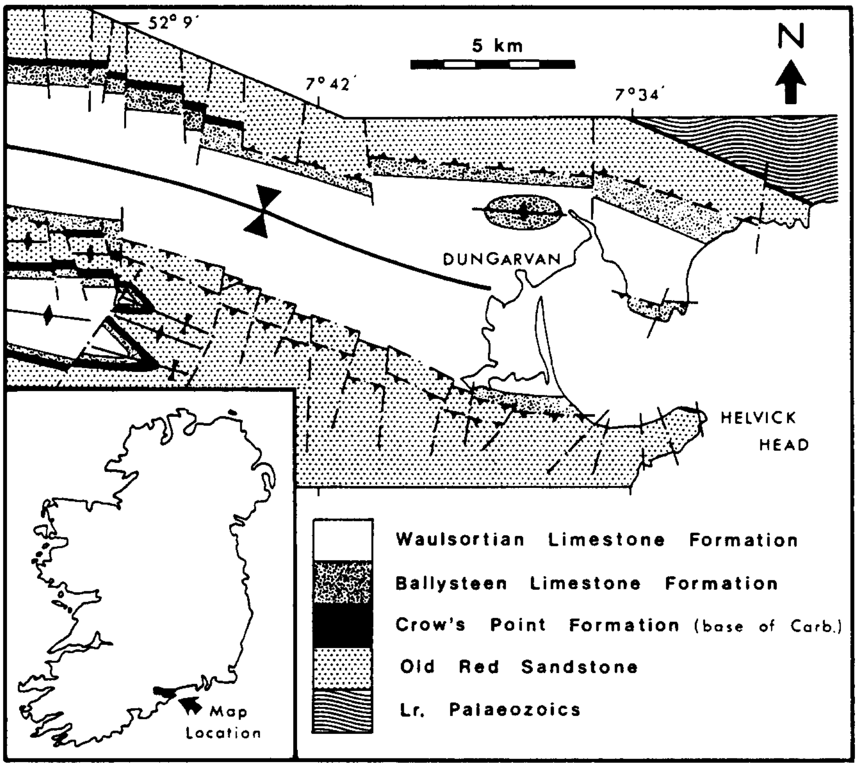

Fig. 1. Geological map of the eastern part of the Dungarvan Syncline, showing the location of Helvick Head.

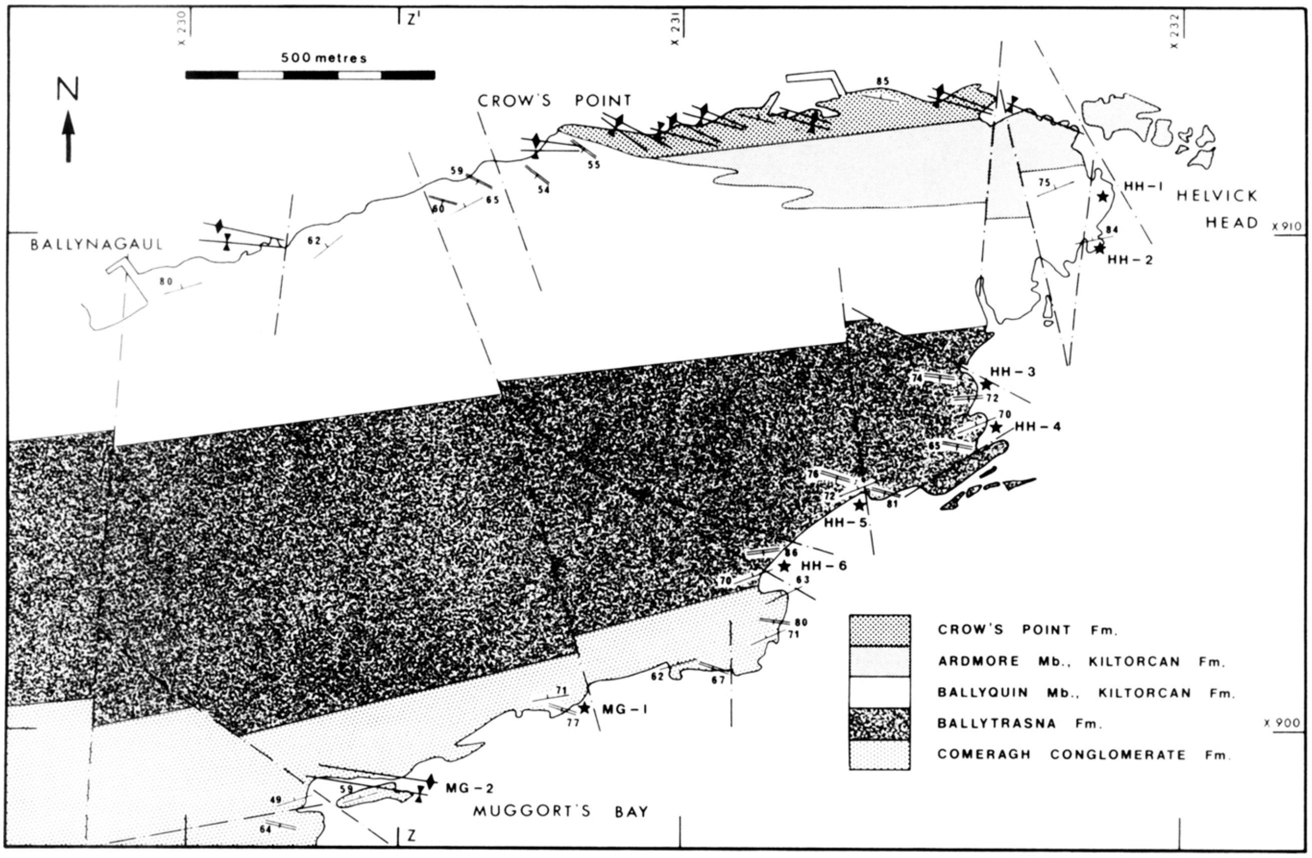

Fig. 2. Geological map of the Helvick Peninsula. Key: single line with tick, bedding; double line with tick, cleavage; stars, thin-section localities.

Mesoscopic features

The cleavage zones are generally preferentially eroded on weathered surfaces. There is often a colour contrast between the cleavage domains, which are frequently pale green to greenish grey, and the purple or deep red microlithons. The domains range in width from 0.5 to 10 mm and in length from a few cm to a few tens of cm. They are generally parallel and smooth to sinuous in shape (Fig. 5a). In general, cleavage domains are continuous through an entire bed only where the strata are thinly bedded. They seldom extend from the top to the base of beds thicker than 1 m. Some bifurcate, especially close to their terminations (Fig. 6). 'Horse-tails', diffuse zones of solution seams, similar to those described by Boyer (1984) commonly occur at their terminations. En échelon patterns are sometimes developed, where one zone terminates another commences close by (Fig. 5b). These are often linked by 'horse-tails'. However, some cleavage zones are linked by microfaults. Their surfaces are commonly lined by minute quartz veins. The microlithons are also highly variable in width (5–50 mm) and

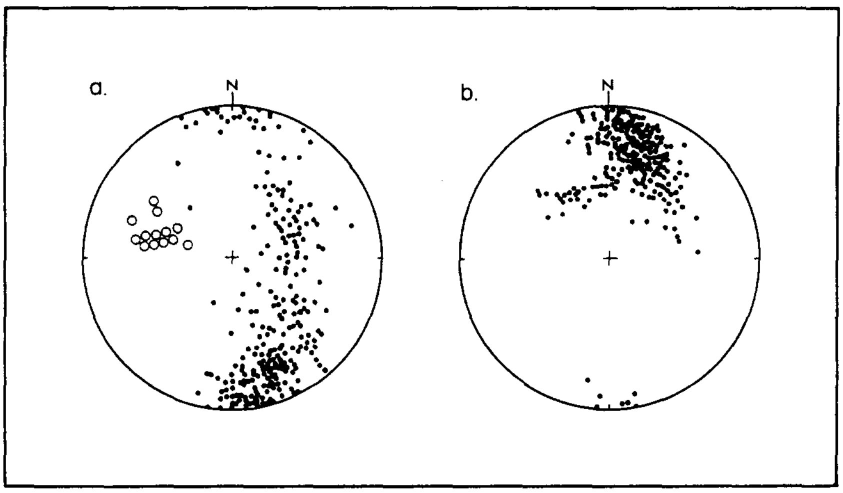

Fig. 3. Equal-area stereographic projections of: (a) fold axes (open circles) and poles to bedding (dots); (b) poles to cleavage (dots) from the Helvick Peninsula.

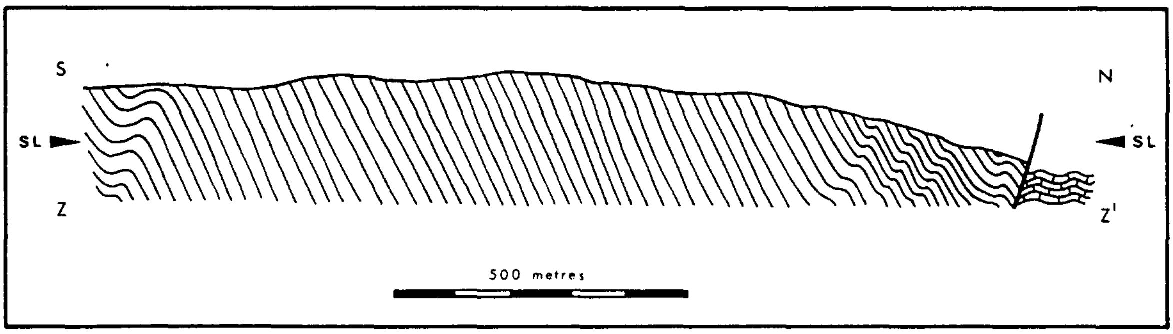

Fig. 4. Structural cross-section through the Helvick Peninsula. The line of section ZZ' is shown in Fig. 2. SL, sea level.

in general, the coarser the grain size, the greater the separation between the cleavage domains. Thin (1–2 mm) laminae are commonly offset across the cleavage domains by up to 5 mm. However, occasionally the displacements are of the order of several cm. Sometimes the laminae can be traced through them (Fig. 6); upon approaching the zone margin they begin to curve, becoming strongly reorientated within the zone, and return to their original orientation on the opposite side. The sense of curvature or offset is consistent from one zone to another, although the degree of curvature or amount of offset varies with zone width and abundance of solution seams. Rarely, very thin 'shear' veins with down-dip slickenside fibres occur along cleavage surfaces. These are usually less than 1 mm thick.

Bedding surfaces are often stepped due to the offset across successive cleavages (Fig. 5c). The magnitude of offset is similar to that displayed by adjacent laminae (Fig. 5d). Where coarse siltstones or fine sandstones containing a zonal cleavage are interbedded with finer siltstones or mudstones, the cleavage tends to splay outwards at the contact into the finer lithology (Fig. 5e), producing structures which resemble the bundled slaty cleavage of Southwick (1987). The offset across a single cleavage domain in the former is distributed across numerous cleavage surfaces in the latter. A refraction of cleavage normally occurs across such contacts.

However, in many cases the strain incompatibility produced by differential shortening between the strongly cleaved mudstones or fine siltstones and the moderately to weakly cleaved coarse siltstones or sandstones is accommodated by bedding surface slip. Such surfaces are characterized by thin quartz shear veins with down-dip slickenside lineations. Similar features have been observed by Cooper *et al.* (1986).

Microscopic features

Microscopic features have been examined in specimens collected from several localities on Helvick Head (Fig. 2). A number of thin sections were cut in the *ac*-plane, the profile plane of the regional fold structure, and are hence perpendicular to cleavage. Some sections were also obtained in other orientations such as the *ab*-plane, perpendicular to cleavage but parallel to the regional fold axis. The cleavage domains normally have discrete boundaries (Fig. 7a) and are composed of a number of very thin, anastomosing or discontinuous solution seams (Fig. 7d) and thin mica films (Voll 1960, Gregg 1985) which are commonly aligned along truncated grain boundaries (Fig. 7b). Sometimes solution seams are less numerous close to the zone margins, and grain-size reduction less pronounced, although in general the transition between microlithon and cleavage domain is sharp, occurring within a fraction of a mm. There is a marked reduction in the proportion of quartz within the zones, coupled with a dramatic decrease in the quartz grain size. The mineral compositions were determined by point counting. A minimum of 300 counts per cleavage domain were measured for a number of samples and from the adjacent microlithons. Quartz decreased from about 55 to 60% in the latter to between 30 and 40% within the cleavage domains. There is a corresponding increase in the phyllosilicate and opaque content (Table 1).

Aspect ratios of quartz grains within the cleavage domains are much higher than within the microlithons. A strong alignment of quartz long axes occurs (Fig. 7b). This is best developed in those thin sections cut in the *ac*-plane, although aspect ratios from sections in the *ab*-plane are also high. There is a consistent clockwise

Fig. 5. Disjunctive cleavage domains from Helvick Head. (a) Rough disjunctive cleavage within a steeply northward-dipping conglomerate overlain by a more intensely cleaved siltstone (locality MG-1). Scale bar = 50 cm. (b) Lenticular cleavage zones within a fine sandstone. A—cleavage zones show an en échelon arrangement. Each zone is linked to the adjacent zones by 'horse tails'. B and C—Parallel laminae are offset across cleavage by more than 1 cm (locality HH-2). Scale bar = 10 cm. (c) Cleavage within coarse siltstone (left) and fine sandstone (right). A—Bedding surface offset across cleavage (locality HH-2). Scale bar = 10 cm. (d) Widely spaced disjunctive cleavage within a fine sandstone. A—Small offsets of sedimentary laminae (locality HH-2). Scale bar = 10 cm. (e) Widely spaced cleavage within a thin coarse sandstone (right) and a more closely spaced cleavage within a coarse siltstone (left). A—Single cleavage domain within the sandstone splays outwards into a more diffuse zone rich in weaker cleavage domains within the siltstone (locality HH-2). Scale bar = 10 cm.

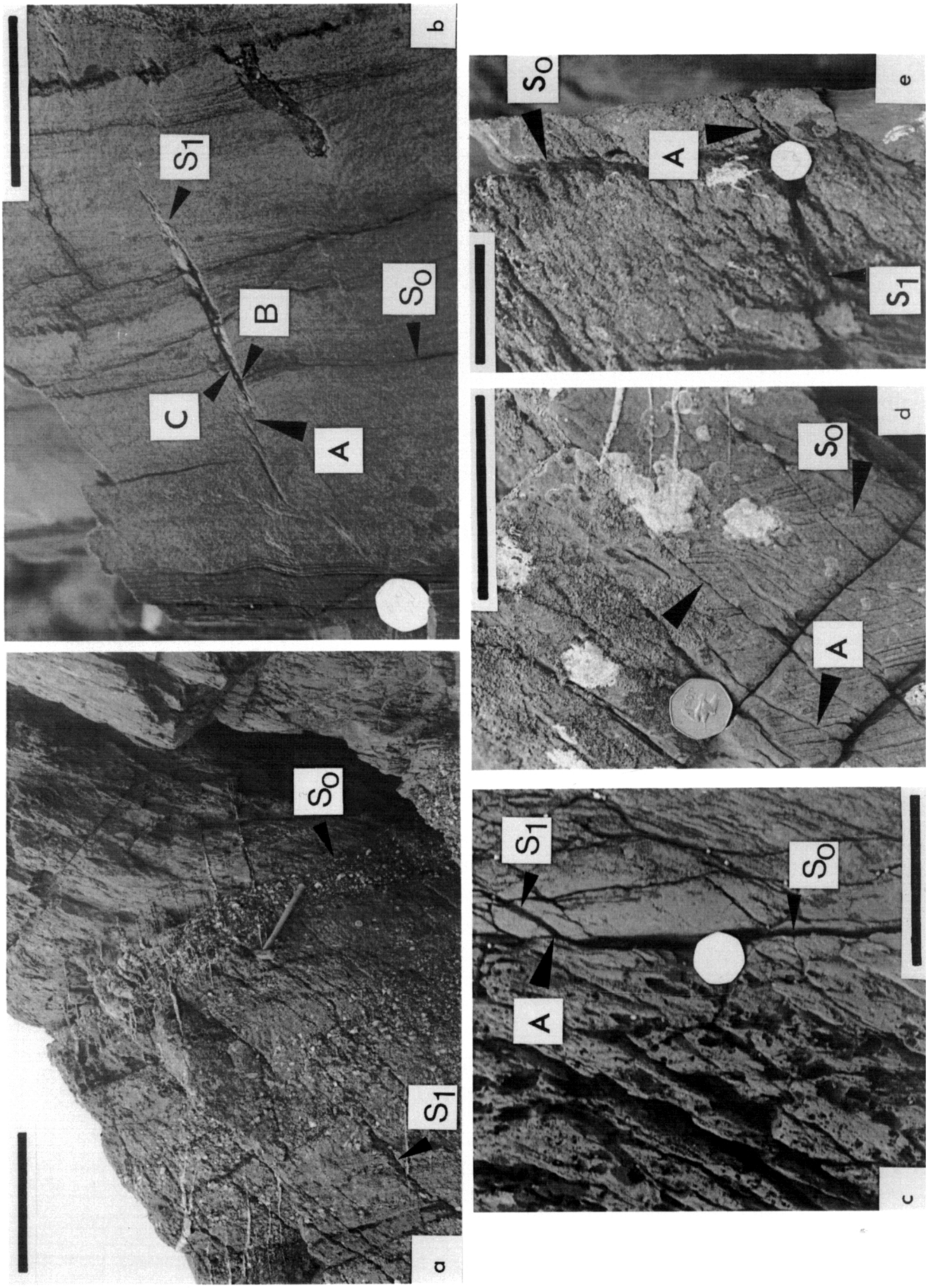

Fig. 5.

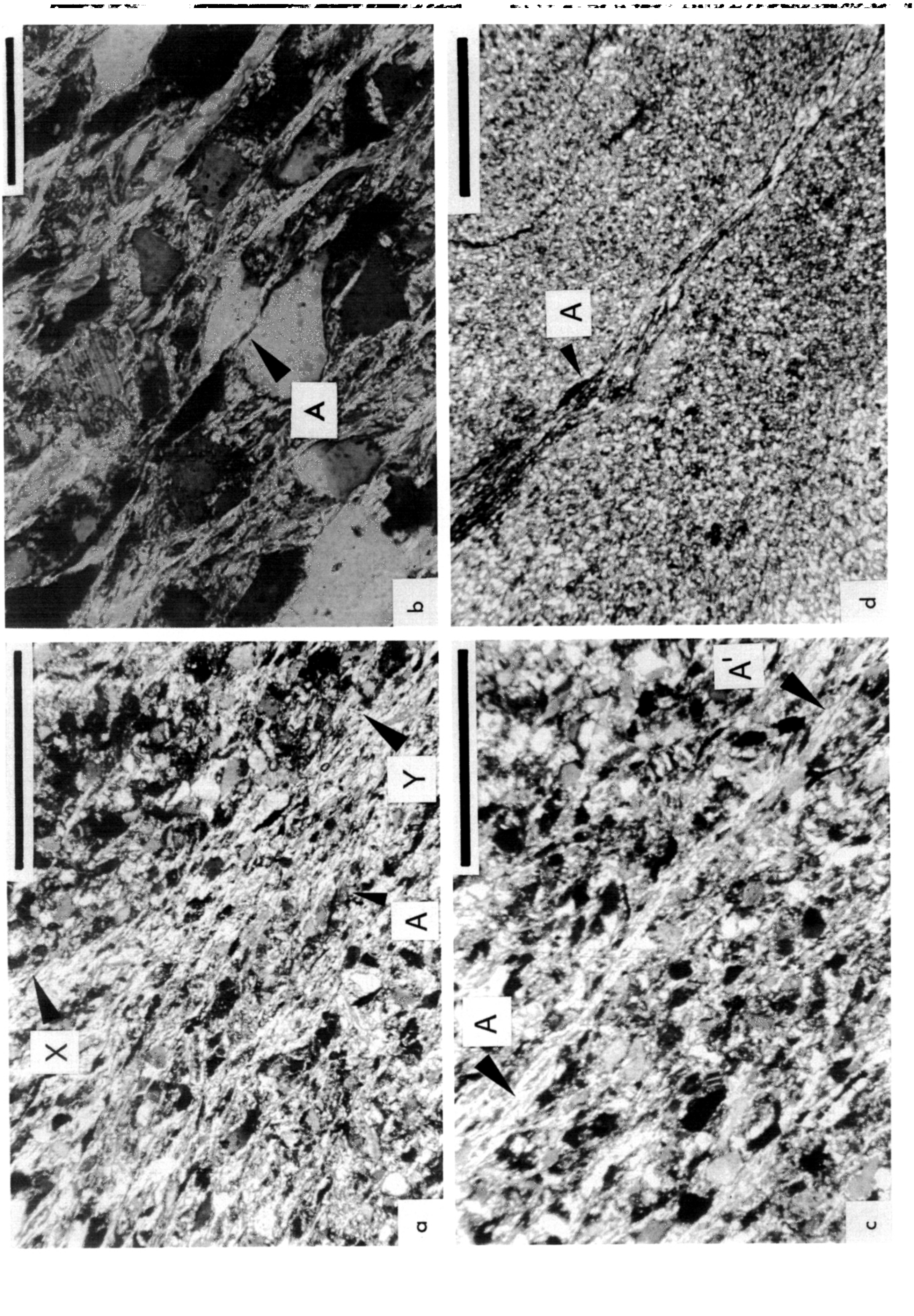

Fig. 7.

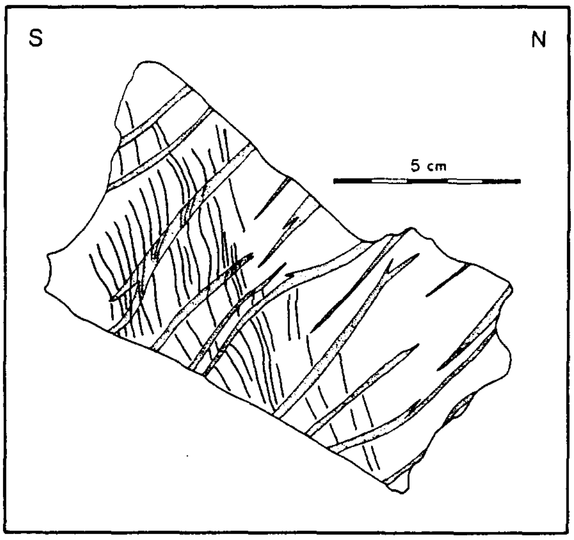

Fig. 6. Line drawing of a polished slab of coarse siltstone showing typical features of zonal cleavage (stippled) with offset sedimentary laminae (thick lines). The laminae can also be seen to change orientation within some cleavage zones (locality HH-3).

(looking west) asymmetry of the long axes with respect to the zone margins within the former. However, within the latter this is not readily apparent.

The fine-grained phyllosilicates within the cleavage domains are aligned approximately parallel to the quartz grain long axes (Fig. 7b) and often appear to be wrapped around the grains. The larger detrital flakes are commonly kinked, bent or frayed, and show strong undulose extinction. The phyllosilicates within the microlithons are less deformed. They show weak undulose extinction and are occasionally kinked, possibly due to compaction rather than tectonic deformation. Within the cleavage domains the phyllosilicates occur as either aligned single grain structures or as mica film segments, although occasionally lengthened mica films (Gregg 1985) occur (Fig. 7c). Solution seams, absent from the microlithons, are abundant within the cleavage domains. They normally occur along the planar boundaries of quartz grains and are aligned approximately parallel to the cleavage domain boundaries. Overgrowths commonly occur on the short margins of these grains, consisting of phyllosilicate beards, syntaxial quartz overgrowths and intergrowths of fibrous quartz and phyllosilicates aligned parallel to the quartz long axes and contributing further to the enhancement of their aspect ratios (Fig. 7b). Sedimentary laminae consisting of opaque heavy mineral placer horizons or phyllosilicate-rich bands are often obliterated within the cleavage zones. However, sometimes they can be traced through them. When oblique to the zone margins they show an apparent orientation change within the zone. A tiny offset occurs across each of the numerous closely spaced pressure solution seams within the cleavage zone.

Rarely very thin ribbons of quartz occur within the cleavage domains (Fig. 7). The quartz ribbons are

Table 1. Point counting results, volume loss estimates, pressure solution offsets and intermicrolithon-slip calculations for cleavage zones from Helvick Head

Specimen No.	CZ-1		CZ-3		CZ-4		CZ-5		CZ-12		CZ-14		G6-1		G6-2			
Locality	HH-2		HH-2		HH-5		HH-6		HH-4		HH-5		HH-3		HH-3			
	C	M	C	M	C	M	C	M	C	M	C	M	C	M	C	C	C	M
Quartz	33.8	54.8	36.8	59.3	34.5	56.4	35.2	55.3	32.4	58.2	34.0	56.7	38.2	62.6	35.2	36.4	38.4	52.8
Feldspar	0.5	0.7	0.4	0.6	0.4	0.6	0.5	0.8	0.3	0.7	0.6	0.7	0.2	0.2	0.8	0.6	0.4	0.6
Detrital mica	4.2	3.3	2.4	5.1	4.6	3.8	5.8	4.2	4.1	3.3	5.2	4.4	2.4	1.0	8.0	4.8	5.4	3.1
Opaques	3.8	2.6	3.5	2.3	3.6	2.4	3.2	2.0	3.8	2.0	4.1	2.8	5.2	5.4	3.8	2.2	3.6	2.5
Chlorite	1.1	1.0	1.0	0.9	1.0	0.8	0.8	0.7	0.9	0.5	0.8	0.5	1.2	1.6	0.6	—	—	0.9
Phyllosilicates	56.6	37.6	55.9	31.8	55.9	36.0	54.5	37.0	58.5	35.3	55.3	34.8	52.8	29.2	46.2	56.0	61.2	36.3
n	300	300	300	300	300	300	300	300	300	300	300	300	500	500	500	500	500	1000
I	65.7	44.5	62.8	40.1	65.1	43.0	64.3	43.9	67.3	41.1	65.4	42.5	61.6	37.2	58.6	63.0	61.2	42.8
$\Delta V = (I_m/I_c) - 1$	−0.322		−0.361		−0.339		−0.317		−0.389		−0.371		−0.396		−0.269	−0.320	−0.301	
Volume loss (%)	32		36		34		32		39		37		40		27	32	30	
W_1 (mm)	2.25		1.5		3.0		2.0		2.0		2.5		4.0		2.5	3.5	1.5	
W_o (mm)	3.30		2.3		4.6		2.9		3.3		4.0		6.7		3.4	5.2	2.1	
S_o, S_1	70°N, 72°S		70°N, 50°S		80°S, 60°N		70°N, 72°S		70°N, 50°S		80°N, 60°S		70°N, 65°S		70°N, 65°S			
θ (°)	38		60		40		38		60		40		45		45			
X_{ps} (mm)	1.30		0.5		1.9		1.2		0.8		1.7		2.7		0.9	1.7	0.6	
X_t (mm)	2.25		2.0		2.5		2.0		2.0		2.0		4.0		1.5	3.0	2.0	
X_{mm}	0.95		1.5		0.6		0.8		1.2		0.3		1.3		0.6	1.3	1.4	

Key: C, cleavage domain; M, microlithon; n, number of point counts per sample; I, proportion of insoluble material; I_m, insolubles within microlithons; I_c, insolubles within cleavage domains; ΔV, volume loss within cleavage domains; W_1, present width of cleavage zone; W_o, original width of the segment of rock from which the cleavage zone was derived; S_o, dip of bedding; S_1, dip of cleavage; θ, angle between bedding and cleavage; X_{ps}, offset of bedding across the cleavage zone due to pressure solution; X_t, total offset of bedding across cleavage; X_{ms}, offset of bedding across cleavage due to intermicrolithon-slip.

Fig. 7. Photomicrographs of disjunctive cleavage domains from Helvick Head. (a) Phyllosilicate-rich cleavage domain (lower left) is separated from the quartz-rich microlithons (upper right) by a sharp boundary (X–Y). A—Phyllosilicate flakes and elongate quartz grains are orientated obliquely to the cleavage-domain boundary (specimen G6-1, locality HH-3). Scale bar = 1 mm. (b) Strong grain shape fabric within a cleavage zone. The quartz grain margins parallel to the fabric appear truncated while those normal to the fabric display phyllosilicate beards and overgrowths. A—Quartz grain fractured and offset along a phyllosilicate seam (specimen G6-1, locality HH-3). Scale bar = 0.1 mm. (c) A lengthened mica film developed within a cleavage zone (A–A') due to the amalgamation of small phyllosilicate flakes as a consequence of the removal of the intervening quartz grains by pressure solution (specimen CZ-2, locality HH-2). Scale bar = 1 mm. (d) Thin cleavage domain within a fine siltstone. A cluster of solution seams occur in the top left. These pass into a thinner cleavage zone through which runs a thin quartz-lined fracture. This produces a small pull-apart structure at A (specimen CZ-7, locality HH-6). Scale bar = 1 mm.

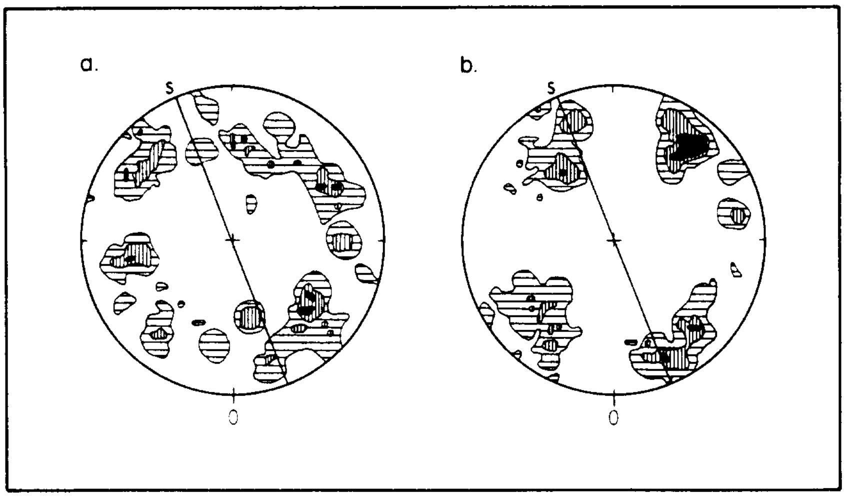

Fig. 8. Equal-area stereographic projections of quartz c-axes within (a) the cleavage zones (n = 300) and (b) the microlithons (n = 300). Measurements obtained from specimens G6-1, G6-2 and G6-3 (locality HH-3). Contour intervals: 1%, horizontal lines; 3%, vertical lines; 5%, black. S, cleavage zone margin.

usually quite long and sometimes extend across an entire thin section. Offsets across these are often visible. Small steps along the ribbons have also formed and resemble microscopic pull-apart structures (Fig. 7d).

Occasionally quartz grains within the cleavage domains are fractured. This is most obvious when there has been only a minor degree of relative movement between the fragments (Fig. 7b). Similar features have been recognized within cleaved sandstones by White & Johnston (1981) and Onasch (1983).

Three hundred quartz c-axis measurements were obtained from several of the largest zones (Fig. 8a) and from the microlithons (Fig. 8b). Both distributions were random and there is no obvious relationship of the c-axis to the cleavage zone margins. In addition there is no evidence for increased polygonal subgrain development, indicative of crystal recovery (Hobbs *et al.* 1976), within the domains since the proportion of quartz showing subgrains is approximately equal to that within the microlithons.

Cleavage zone terminations are generally very diffuse. The quartz grain aspect ratios decrease while the quartz size and content increases. The solution seams become more widely spaced and tend to fan out, producing features similar to the 'horse-tails' described by Boyer (1984).

INTERPRETATION AND DISCUSSION

Origin

Cleavage domains similar to those under discussion have not been commonly described within the literature. They are most analogous to the type-c 'rough' cleavage of Gray (1978) and Onasch (1983), and the P–Q fabrics of Waldron & Sandiford (1988). However, Gray's (1978) suggestion that they did not develop within an initially homogeneous rock, but in fact formed along a primary sedimentary anisotropy, possibly produced by dewatering channels during sediment compaction or by bioturbation, is inapplicable in this case. The utilization of burrows as favoured sites of pressure solution has also been evoked by Nickelsen (1972) to account for the formation of similar zones. However, those on Helvick Head are planar structures and it seems unlikely that they could have originated from linear features such as burrows. In addition, sedimentary laminae can often be traced through the cleavage domains and show an orientation change within them. If these structures originated as burrows then all traces of primary sedimentary structures should have been obliterated by bioturbation. Their origin as dyke-like dewatering channels can similarly be ruled out. If they were dewatering structures there should be some continuity between them and the overlying or underlying strata. However, many of these zones are discontinuous or en échelon, and are confined to the interiors of beds.

In thin section these zones show features typical of pressure solution. Solution seams, absent from the microlithons, are abundant within the cleavage domains and are often aligned parallel to the quartz grain long axes. Grain-boundary truncation is also important, producing elongate grains with planar truncated margins parallel to the solution seams. Phyllosilicate beards and intergrown quartz–phyllosilicate pressure fringes are frequently developed on quartz grain boundaries at right angles to the solution seams. These features provide ample evidence that the cleavage zones have undergone considerable volume loss due to pressure solution. Similar conclusions concerning the origin of disjunctive cleavage in the Irish Variscides have been drawn by Beach & King (1978), Sanderson (1984), Cooper *et al.* (1986) and O'Sullivan *et al.* (1986). Consequently their reduced grain size and increased phyllosilicate content is almost entirely attributed to pressure solution. However, the presence of fractured quartz grains indicates that microfracturing has also made a very minor contribution to grain-size diminution.

The interbedded coarse sandstones and conglomerates are intensely veined, indicating that much of the dissolved material was locally redeposited within hydraulic fractures. By considering the proportion of insoluble material to be constant (Gratier 1983), volume losses of up to 40% have been determined for the largest cleavage zones (Table 1).

The mica film segments within the cleavage zones have formed primarily due to the dissolution of adjacent and intervening quartz grains in contrast to the origin of mica films proposed by Gregg (1985) in which recrystallization and reorientation of phyllosilicates were considered important. The presence of phyllosilicates enhances pressure solution (Engelder & Marshak 1985, Marshak & Engelder 1985). Thus, as the proportion of mica passively increases due to the removal of quartz the rate of pressure solution will increase, given a constant stress.

Strain within the cleavage zones

The area is devoid of reliable strain markers. Even reduction spots and quartz fibre pressure shadows associated with pyrite cubes which are common throughout the Irish Variscides (Hanna 1983) are absent from the Helvick Peninsula. However, an estimate of the strain within the cleavage domains can be obtained by treating the quartz grains as strain markers. The strain due to pressure solution can be analysed using R_f/ϕ techniques, although the R_f/ϕ plots will be characterized by greater fluctuations than those of passive strain markers (Lisle 1985). This analysis would be invalid if the distances envisaged by Waldron & Sandiford (1988) of the order of 5–10 mm for solute transfer within psammites were applicable to this area, since a proportion of the material would be locally redeposited within the domains. However, it is obvious from vein relationships that solute migration occurred over a much wider scale. Calcrete horizons occur at certain intervals within the mudstones of the Ballytrasna and Kiltorcan Formations. These mudstones have undergone intense pressure solution during cleavage formation. Consequently much of the carbonate was removed from the calcretes and locally redeposited as carbonate veins, as composite quartz veins with carbonate cores or as vein intergrowths of quartz and carbonate. These veins may occur up to 20 m stratigraphically above or below the nearest calcrete, indicating that material removed by pressure solution could migrate across distances much greater than those suggested by Waldron & Sandiford (1988).

The axial ratios of quartz grains within a number of cleavage zones and their orientations relative to cleavage are plotted on hyperbolic nets shown in Fig. 9 (De Paor 1981a,b, 1988). Using this net, the parameters of the strain ellipse (R_s/ϕ_s) can be determined. An overlay is rotated about the axis of the net until the N–S R-axis subdivides the data equally. This gives the orientation, ϕ_s, of the maximum principal stretch axis to cleavage. Then the hyperbola which splits the data into two equal parts is found. This intersects the R-axis at a distance R_s from the origin. Thus the parameters of the strain ellipse have been determined. This is valid only on the assumption that the fabric is entirely tectonic in origin. In the cleavage zones examined the strain due to pressure solution, R_s, ranges from 2 to 3.

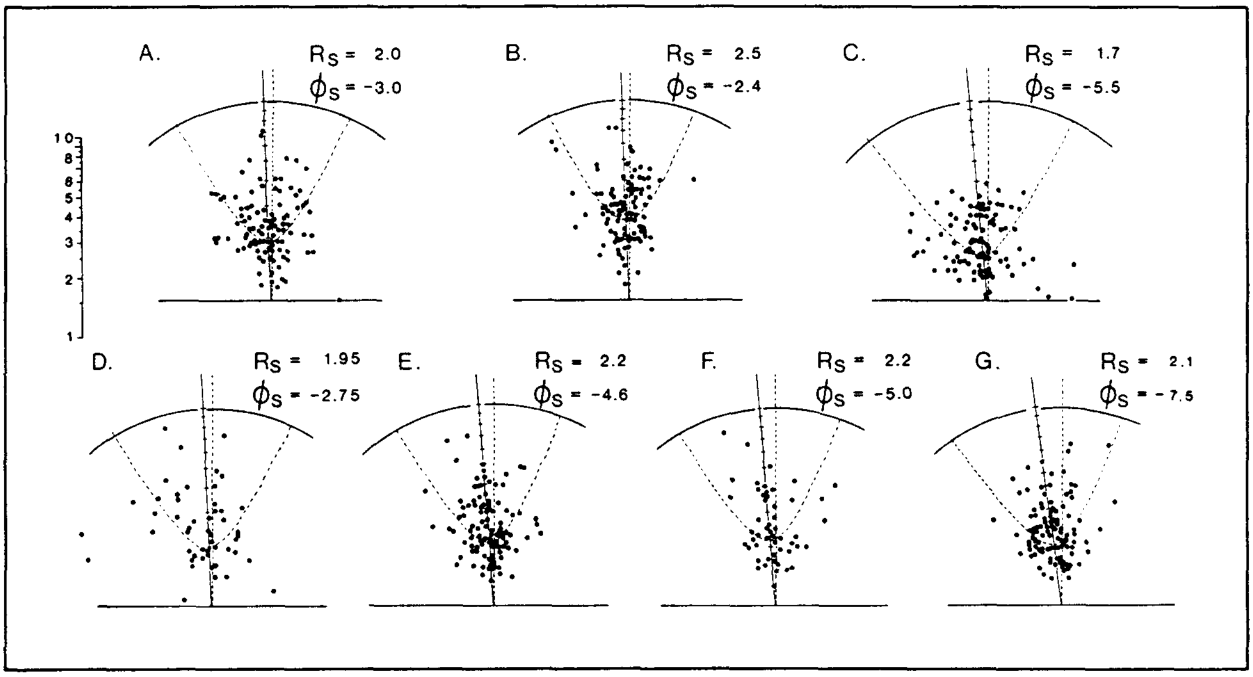

Fig. 9. R_s/ϕ plots on the hyperbolic net (De Paor 1988) for deformed quartz grains within cleavage zones. Dashed line, cleavage trace (zero ϕ position); dashed hyperbola, 50% of data curve; solid line, principal stretch axis. A, Specimen G6-1, locality HH-3, n = 100; B, specimen G6-1, locality HH-3, n = 100; C, specimen G6-2, locality HH-3, n = 100; D, specimen G6-2, locality HH-3, n = 50; E, specimen CZ-1, locality HH-2, n = 100; F, specimen CZ-3, locality HH-2, n = 50; G, specimen CZ-12, locality HH-4, n = 100.

The shape factor, ε (De Paor 1988), within the cleavage domains can be determined from the equation:

$$\varepsilon = 1/2 \ln (R),$$

where R, the axial ratio, is equal to R_s. Thus ε ranges from 0.35 to 0.55. These strain values are typical of those recorded from disjunctive pressure solution cleavage within psammites (Lisle 1977, Onasch 1983, Norris & Rupke 1986).

If these cleavage zones were the product of pressure solution alone, ϕ_s should be approximately zero (parallel to cleavage). However, each zone shows a small deviation of ϕ_s from cleavage in a sense which is consistent from one zone to another as a result of the asymmetry of the quartz grain long axes with respect to the boundaries of the cleavage domains. If their shape and orientation were solely a function of pressure solution acting perpendicular to the cleavage their long axes should be parallel or symmetrically disposed about the zone margin. Quartz grain axial ratios within the microlithons were plotted against ϕ (Fig. 10). A random pattern was obtained, indicating that any significant grain shape modification has been confined to the cleavage domains.

It is considered that this asymmetry was produced at a fairly late stage in the deformation history, after the main intensity of pressure solution had ceased, probably during late fold tightening. It can be attributed to two possible mechanisms: (1) shearing parallel to cleavage; or (2) changing direction of maximum dissolution.

Model 1. The cleavage was formed early in the deformation event (Cooper *et al.* 1986) by pressure solution processes (Fig. 11a). However, during folding the cleavage was then rotated out of parallelism with the XY plane of the bulk rock finite strain ellipsoid (a plane along which shearing is impossible) so that it became geometrically possible for shear to take place (Ghosh 1982). Thus as the folds tightened and flexural-slip became more difficult there were two possible ways in which further deformation could take place: flexural-flow could occur, deforming and producing a sigmoidal cleavage (Trayner & Cooper 1984); or slip along the cleavage could take place. These cleavage zones, containing fine-grained quartz and a high proportion of clay or fine silt grade phyllosilicates, probably represent zones of weakness and shear-strain localization which enabled the more competent microlithons to slide past each other. This movement would have rotated the grains within the cleavage domains (Fig. 11b), especially those with high aspect ratios, resulting in the observed clockwise (looking west) asymmetry (Fig. 11c). In general the movement was minute. However, where the slip was greater, from a few mm to several cm, thin quartz slickenside fibres sometimes developed along the cleav-

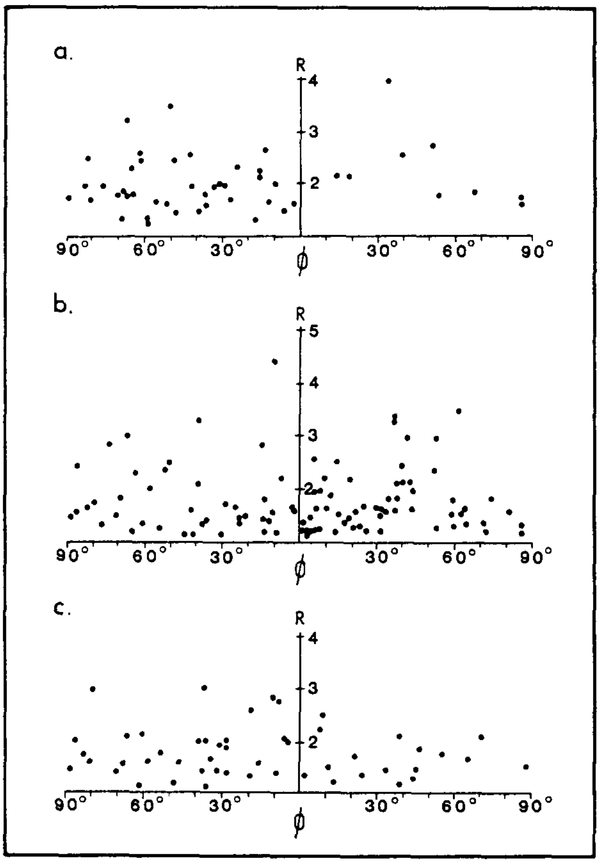

Fig 10. Plots of the axial ratios, R, of quartz grains within the microlithons against ϕ, the angle of deviation of the quartz grain long axes from cleavage-zone boundaries. (a) Specimen G6-1, locality HH-3, $n = 50$. (b) Specimen G6-2, locality HH-3, $n = 100$. (c) Specimen CZ-3, locality HH-2, $n = 50$.

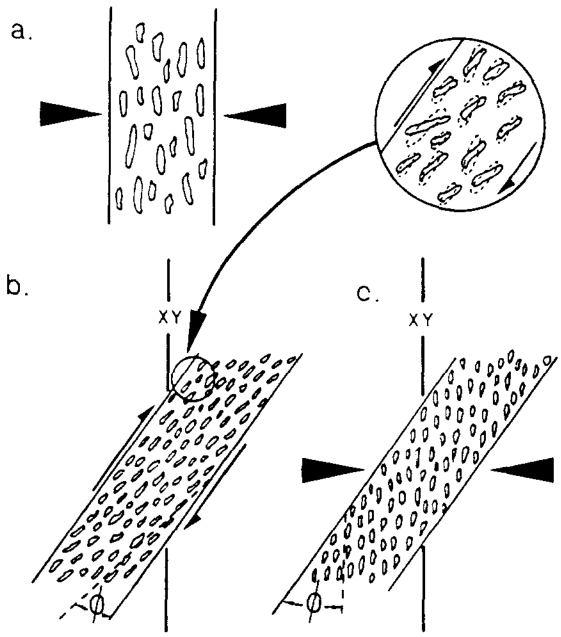

Fig. 11. Sketches to illustrate the effect of intermicrolithon-slip upon quartz grain orientation within cleavage domains. (a) Fabric produced within a cleavage zone by pressure solution. An alignment of quartz grain long axes parallel to the margins of the cleavage zones is produced as a consequence of grain-boundary dissolution. (b) Model 1: as the cleavage is rotated out of parallelism with the XY plane of the finite strain ellipsoid intermicrolithon-slip becomes geometrically possible, and the elongate grains (see inset) are rotated by the shearing. This results in a clockwise obliquity of their long axes with respect to the cleavage-zone boundary. (c) Model 2: alternatively as the cleavage is rotated out of parallelism with the XY plane of the finite strain ellipsoid the direction of maximum dissolution within the cleavage zone may change producing an anticlockwise obliquity of the quartz grain long axes with respect to the cleavage zone boundary. The sense of obliquity in model 1 is similar to that observed within the cleavage zones studied.

age surfaces. In thin section these appear as quartz ribbons. Offsets across them are visible, and microfracturing of adjacent quartz grains has occurred. Slip along these surfaces has also occasionally resulted in the formation of tiny pull-apart structures.

Model 2. An alternative mechanism to account for the long axis asymmetry without evoking shear strain involves a similar early history with the initiation and intensification of cleavage by pressure solution processes during early buckling. As the folds tightened and the cleavage was rotated out of parallelism with the XY plane of the finite strain ellipsoid, pressure solution was restricted to the pre-existing cleavage zones (Gratier 1987) since these provided excellent channel-ways for the migration of fluid necessary during pressure solution (Engelder & Marshak 1985). In addition the initiation of new cleavage domains in strata which were rapidly changing position with respect to the orientation of the XY plane of the finite strain ellipsoid would be difficult. The direction of maximum dissolution within the cleavage domains changed as the fold limbs rotated in order to remain orthogonal to the XY plane. Soper (1986) noted that it is improbable that pressure solution would cease immediately upon rotation of the cleavage domains out of parallelism with the XY plane since the presence of residues along the cleavage seams would enhance solution. Consequently pressure solution could produce elongation of grains at angles oblique to the cleavage zone margins (Fig. 11d). However, the sense of asymmetry expected in such a situation is opposite to that which has actually developed. In addition, a change in the direction of maximum dissolution cannot account for grain microfracturing within the cleavage domains.

Therefore it is considered that intermicrolithon-slip (model 1) is the most likely mechanism for the generation of the observed features. The rare occurrence of slickensides with down-dip lineations on cleavage surfaces is also consistent with this interpretation. Gratier (1987) noted that dissolution can continue along cleavages oblique to σ_1 without slip. By experimentation he found that low deviatoric stresses (10–25 MPa) are needed to avoid slippage on the cleavage when the angle between the normal to cleavage and σ_1 ranges from 0 to 30°. At angles greater than 30° a second cleavage tends to form (Gratier & Vialon 1980). In the Helvick region the cleavage domains under investigation typically dip at angles in excess of 60°S. At the time of deformation σ_1 is considered to have been essentially horizontal (Cooper *et al.* 1986). Thus the angle between σ_1 and the normal to cleavage was invariably less than 30°. Consequently it can be inferred from Gratier's (1987) experimental evidence that the cleavage domains deformed under relatively high deviatoric stress, producing slip rather than further pressure solutions as the cleavage was rotated away from the XY plane.

Displacement along cleavage

Many cleavage zones are discontinuous and do not extend from the top to the base of the bed. The displacement on these is accommodated by 'horse-tails' at their terminations. Sometimes where the zones have an en échelon arrangement slip is transferred from one zone to another by means of a 'horse-tail' as described by Boyer (1984). Displacement along most cannot have been greater than a few mm since passive markers are rarely offset by more than 5 mm. The bulk of this is a consequence of volume loss due to pressure solution with only a minor contribution from intermicrolithon-slip. The absolute values of slip cannot be easily determined. However, if the volume loss is derived for an individual zone, then the amount of offset due to pressure solution can be calculated. If this is found to be smaller than the true offset, then the difference is equal to the amount of slip across that zone. The offset due to pressure solution, X_{ps}, was calculated from the equation:

$$X_{ps} = (W_o - W_1)/\tan\theta.$$

where W_o = original width before pressure solution, W_1 = present width of the cleavage zone and θ = angle between bedding and cleavage (Fig. 12). This was determined for some of the more pronounced cleavage domains (Table 1), and it was found that 0.3–1.5 mm of slip were required to explain their observed offsets. This may seem insignificant, but when all of the cleavage domains on the southern limb of the Dungarvan Syncline are taken into consideration the total displacement must be quite large.

A very rough calculation can be made by assuming an average cleavage dip of 65°, a microlithon width of 10 mm and an intermicrolithon-slip of 1 mm on a fold limb

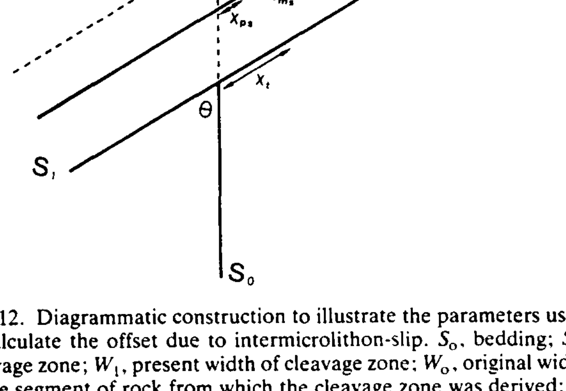

Fig. 12. Diagrammatic construction to illustrate the parameters used to calculate the offset due to intermicrolithon-slip. S_0, bedding; S_1, cleavage zone; W_1, present width of cleavage zone; W_o, original width of the segment of rock from which the cleavage zone was derived; θ, angle between bedding and cleavage; X_{ps}, offset of bedding across the cleavage zone due to pressure solution; X_{ms}, offset of bedding across cleavage due to intermicrolithon-slip; X_t, total offset of bedding across cleavage. W_o can be calculated once the volume loss within the cleavage zone is known, enabling the calculation of X_{ps} by simple trigonometry.

with a down-dip length of 5 km, a typical value for first-order folds within the Irish Variscides. The total displacement due to intermicrolithon-slip on the limb is 500 m, which represents a bedding-parallel shortening of about 210 m or an intermicrolithon-slip natural strain (ε_{ims}) of -0.042. This very generalized model gives a rough estimate of the shortening due to intermicrolithon-slip and indicates that it can be quite significant.

CONCLUSIONS

The offset of bedding surfaces, sedimentary laminae, veins and other passive markers across disjunctive cleavage domains has commonly been attributed solely to the loss of material due to pressure solution. However, Engelder & Marshak (1985) noted that intermicrolithon-slip may occur once the cleavage is rotated out of parallelism with the XY plane of the finite strain ellipsoid. This phenomenon has not been widely recorded although it has been observed with the Irish Variscides (Coe & Selwood 1963, 1964, Cooper *et al.* 1986, O'Sullivan *et al.* 1986).

The study of widely spaced disjunctive cleavage domains on Helvick Head reveals that pressure solution was not the only process involved in their formation. It was determined that there was a consistent clockwise asymmetry of quartz grain long axes with respect to the cleavage zone margin and that microfracturing of quartz grains within the zones had occurred. The sense of asymmetry suggests that shearing took place parallel to cleavage. This is believed to be a consequence of fold tightening, after the main intensity of cleavage formation had taken place, which rotated the cleavage out of parallelism with the XY plane of the finite strain ellipsoid. Slip could then take place along the cleavage zones which represented weaknesses within an otherwise homogeneous rock. This movement was another means of accommodating shortening since further folding by flexural-slip was difficult.

The offset due to dissolution (X_{ps}) can be derived from the calculation of the amount of volume loss. The difference between X_{ps} and the actual offset across the cleavage (X_t) is equal to the displacement due to slip (X_{ms}). The magnitude of intermicrolithon-slip along each zone is very small, rarely exceeding 1.5 mm. However, when all of the cleavage domains on the limb of a major fold such as the Dungarvan Syncline are considered it is apparent that the overall displacement is quite large.

Simon & Gray (1982) gave the total natural shortening strain (ε_T) for a deformed sedimentary sequence as:

$$\varepsilon_T = \varepsilon_{lps} + \varepsilon_b + \varepsilon_f + \varepsilon_{ps},$$

where ε_{lps} = layer-parallel shortening strain, ε_b = buckling strain, ε_f = fold flattening strain and ε_{ps} = pressure solution strain. However if intermicrolithon-slip is indeed widespread this equation must be modified to become:

$$\varepsilon_T = \varepsilon_{lps} + \varepsilon_b + \varepsilon_f + \varepsilon_{ps} + \varepsilon_{ims}.$$

Consequently it is unwise to attribute offsets across cleavage entirely to volume loss without careful examination. A more detailed study may reveal that this is not a localized feature but occurs throughout the Irish Variscides and may also be important in other orogenic belts.

Acknowledgements—The field work which provided the basis for this paper was financed by a Geological Survey of Ireland Contract Mapping Studentship. Dr M. A. Cooper provided helpful advice during the initial stages of this work. The paper was greatly improved by the comments and constructive criticism of Dr S. H. Treagus and the anonymous referees.

REFERENCES

Beach, A. & King, M. 1978. Discussion on pressure solution. *J. geol. Soc. Lond.* **135**, 649–651.

Borradaile, G. J. 1978. Transected folds: a study illustrated with examples from Canada and Scotland. *Bull. geol. Soc. Am.* **89**, 481–493.

Borradaile, G. J. 1982. Anastomosing, discrete seams: solution cleavage. In: *Atlas of Deformational and Metamorphic Rock Fabrics* (edited by Borradaile, G. J., Bayly, M. B. & Powell, C. McA.). Springer, Berlin.

Borradaile, G. J., Bayly, M. B. & Powell, C. McA. (editors) 1982. *Atlas of Deformational and Metamorphic Rock Fabrics.* Springer, Berlin.

Boyer, S. E. 1984. Origin and significance of compositional layering in Late Precambrian sediments, Blue Ridge Province, North Carolina, U.S.A. *J. Struct. Geol.* **6**, 121–133.

Cloos, E. 1947. Oolite deformation in the South Mountain fold, Maryland. *Bull. geol. Soc. Am.* **58**, 843–918.

Coe, K. & Selwood, E. B. 1963. Stratigraphy and structure of part of the Beara Peninsula, Co. Cork. *Proc. R. Ir. Acad.* **63B**, 33–59.

Coe, K. & Selwood, E. B. 1964. Some features of folding and faulting in the Sheeps Head Peninsula, Co. Cork. *Sci. Proc. R. Dubl. Soc.* **2A**, 29–41.

Cooper, M. A., Collins, D., Ford, M., Murphy, F. X. & Trayner, P. M. 1984. Structural style, shortening estimates and the thrust front of the Irish Variscides. In: *Variscan Tectonics of the North Atlantic Region* (edited by Hutton, D. H. W. & Sanderson, D. J.). *Spec. Publs geol. Soc. Lond.* **14**, 167–175.

Cooper, M. A., Collins, D. A., Ford, M., Murphy, F. X., Trayner, P. M. & O'Sullivan, M. 1986. Structural evolution of the Irish Variscides. *J. geol. Soc. Lond.* **143**, 53–61.

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

De Paor, D. G. 1981a. Geological strain analysis. Unpublished Ph.D thesis, National University of Ireland.

De Paor, D. G. 1981b. Orthographic analysis of geological structures. A workshop of the Irish Tectonic Studies Group. University College Galway.

De Paor, D. D. 1988. R_f/ϕ_f strain analysis using an orientation net. *J. Struct. Geol.* **10**, 323–333.

Dieterich, J. H. 1969. Origin of cleavage in folded rocks. *Am. J. Sci.* **267**, 155–165.

Engelder, T. & Marshak, S. 1985. Disjunctive cleavage formed at shallow depths in sedimentary rocks. *J. Struct. Geol.* **7**, 327–343.

Ghosh, S. K. 1982. The problem of shearing along axial plane foliations. *J. Struct. Geol.* **4**, 63–67.

Gratier, J. P. 1983. Estimation of volume changes by comparative chemical analyses in heterogeneously deformed rocks (folds with mass transfer). *J. Struct. Geol.* **5**, 329–339.

Gratier, J. P. 1987. Pressure solution—deposition creep and associated tectonic differentiation in sedimentary rocks. In: *Deformation of Sediments and Sedimentary Rocks* (edited by Jones, M. E. & Preston, R. M. F.). *Spec. Publs geol. Soc. Lond.* **29**, 25–38.

Gratier, J. P. & Vialon, P. 1980. Deformation patterns in a heterogeneous material: folded and cleaved sedimentary cover immediately overlying a crystalline basement (Oisans, French Alps). *Tectonophysics* **65**, 151–180.

Gray, D. R. 1977. Differentiation associated with discrete crenulation cleavages. *Lithos* **10**, 89–101.

Gray, D. R. 1978. Cleavages in psammitic rocks from southeastern Australia: their nature and origin. *Bull. geol. Soc. Am.* **89**, 577–590.

Gregg, W. J. 1985. Microscopic deformation mechanisms associated with mica film formation in cleaved psammitic rocks. *J. Struct. Geol.* **7**, 45–56.
Groshong, R. H. 1976. Strain and pressure solution in the Martinsburg Slate, Delaware Water Gap, New Jersey. *Am. J. Sci.* **276**, 1131–1146.
Hanna, S. S. 1983. Estimates of stratal shortening in South Wales and south-west Ireland. Unpublished Ph.D thesis, University of Wales.
Harker, A. 1886. On slaty cleavage and allied rock structures with special reference to the mechanical theories of their origin. *Rep. Br. Ass. Adv. Sci.* (1885), 813–852.
Hobbs, B. E., Means, W. D. & Williams, P. F. 1976. *An Outline of Structural Geology*. John Wiley & Sons, New York.
Hoeppener, R. 1956. Zum problem der Bruchbildung, Schieferung und Faltung. *Geol. Rdsch.* **45**, 247–283.
Lisle, R. J. 1977. Clastic grain shape and orientation in relation to cleavage from the Aberystwyth Grits, Wales. *Tectonophysics* **39**, 381–395.
Lisle, R. J. 1985. *Geological Strain Analysis. A Manual for the* R_f/ϕ *Technique*. Pergamon Press, Oxford.
Marshak, S. & Engelder, T. 1985. Development of cleavage in limestones of a fold-thrust belt in eastern New York. *J. Struct. Geol.* **7**, 345–359.
Murphy, F. X. 1985. The lithostratigraphy and structural geology of the Dungarvan Syncline and adjacent areas, Co. Waterford, Southern Ireland. Unpublished Ph.D thesis, National University of Ireland.
Murphy, F. X. 1988. The origin of Variscan kink bands: a study from the Dungarvan Syncline, southern Ireland. *Geol. Mag.* **125**, 641–650.
Nickelsen, R. P. 1972. Attributes of rock cleavage in some mudstones and limestones of the Valley and Ridge Province, Pennsylvania. *Proc. Penn. Acad. Sci.* **46**, 107–112.
Norris, R. J. & Rupke, N. A. 1986. Development of slaty cleavage in a mudstone unit from the Cantabrian Mountains, northern Spain. *J. Struct. Geol.* **8**, 871–878.
Onasch, C. M. 1983. Origin and significance of microstructures in sandstones of the Martinsburg Formation, Maryland. *Am. J. Sci.* **283**, 936–966.
O'Sullivan, M. J., Cooper, M. A., MacCarthy, I. A. J. & Forbes, W. H. 1986. The palaeoenvironment and deformation of Beaconites-like burrows in the Old Red Sandstone at Gortnabinna, SW Ireland. *J. geol. Soc. Lond.* **143**, 897-906.
Plessmann, W. 1964. Gesteinslösung, ein Hauptfaktor beim Schieferungsprozess. *Geol. Mitt.* **4**, 69–82.
Powell, C. McA. 1979. A morphological classification of rock cleavage. *Tectonophysics* **58**, 21–34.
Sanderson, D. J. 1984. Structural variation across the northern margin of the Variscides in NW Europe. In: *Variscan Tectonics of the North Atlantic Region* (edited by Hutton, D. H. W. & Sanderson, J. D.). *Spec. Publs geol. Soc. Lond.* **14**, 149–165.
Sharpe, D. 1849. On slaty cleavage. *Q Jl. geol. Soc. Lond.* **5**, 111–115.
Siddans, A. W. B. 1972. Slaty cleavage—a review of research since 1815. *Earth Sci. Rev.* **8**, 205–232.
Simon, R. I. & Gray, D. R. 1982. Interrelations of mesoscopic structures and strain across a small regional fold, Virginia Appalachians. *J. Struct. Geol.* **4**, 271–289.
Soper, N. J. 1986. Geometry of transecting, anastomosing solution cleavage in transpression zones. *J. Struct. Geol.* **8**, 937–940.
Sorby, H. C. 1853. On the origin of slaty cleavage. *Edinb. New Philos. J.* **55**, 137–148.
Southwick, D. L. 1987. Bundled slaty cleavage in laminated argillite, north-central Minnesota. *J. Struct. Geol.* **9**, 985–993.
Trayner, P. M. & Cooper, M. A. 1984. Cleavage geometry and the development of the Church Bay anticline. *J. Struct. Geol.* **6**, 83–87.
Voll, G. 1960. New work on petrofabrics. *Geol. J.* **2**, 503–567.
Waldron, H. M. & Sandiford, M. 1988. Deformation volume and cleavage development in metasedimentary rocks from the Ballarat slate belt. *J. Struct. Geol.* **10**, 53–62.
White, S. H. & Johnston, D. C. 1981. A microstructural and microchemical study of cleavage lamellae in a slate. *J. Struct. Geol.* **3**, 279–290.
Wood, D. S. 1974. Current views on the development of slaty cleavage. *Annu. Rev. Earth & Planet. Sci.* **2**, 1–37.