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Disjunctive cleavage formed at shallow depths in sedimentary rocks

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Abstract—Work of the past decade concerning tectonic cleavage has focused on nine topics: (1) appropriate terminology for cleavage classification, (2) processes involved in cleavage formation, (3) controls on cleavage distribution and morphology, (4) depth dependence of cleavage formation, (5) controls and causes of cleavage initiation, (6) nature of lithologic changes during cleavage development, (7) relation of cleavage to strain, (8) relation of cleavage to stress and (9) chronology of cleavage development. This paper reviews these issues as they apply to the formation of disjunctive cleavage in sedimentary rocks. Discussion is limited to cleavage that formed at shallow crustal conditions in which rock–water interaction, rather than plastic flow, recrystallization, or grain growth, is the dominant deformation process.

INTRODUCTION

DISJUNCTIVE cleavage is a manifestation of deformation of sedimentary rocks in the upper crust of the Earth. Sorby (1853, 1856, 1863, 1908) was the first to establish that interaction of water with rock was necessary for the development of such cleavage. It is now recognized that in rocks that were never buried deeply disjunctive cleavage formed without significant grain-shape change by intracrystalline plastic flow (i.e. less than 15% strain due to dislocation creep), dynamic recrystallization (resulting in grain-size reduction, as occurs during mylonitization), metamorphic phase transitions (including growth of new phyllosilicate species), or total destruction of original sedimentary fabrics. Understanding of disjunctive cleavage has improved significantly during the past 15 years because of revealing theoretical and experimental investigations concerning the process of rock–water interaction (e.g. Durney 1972, 1976, Rutter 1976, 1983, DeBoer 1977, Robin 1978, Etheridge *et al.* 1984) and because of an accumulation of observational data on rock fabrics and strain, using field, electron beam, and X-ray techniques (e.g. Borradaile *et al.* 1982).

The purpose of this paper is to present an overview of nine current issues concerning disjunctive cleavage in sedimentary rocks. We focus on cleavage that formed under shallow crustal conditions (<6 km) in which rock–water interaction is the dominant deformation process. By limiting our scope to this end-member of the cleavage 'family', we will not cover some topics, such as grain growth and recrystallization, that have been discussed by previous reviewers (e.g. Wood 1974, Williams 1977). The present overview relies in part on our own studies of cleavage within strata of the Appalachian Plateau and Hudson Valley regions of the Appalachian Mountains, U.S.A., and of the Umbrian Apennines, Italy, both regions which have never been subjected to more than 6–8 km overburden. The refinement of concepts concerning cleavage formation has required some modification of the terminology used to classify cleavage; thus, the discussion that follows commences with a consideration of current cleavage classification schemes.

CLASSIFICATION

The term 'cleavage', as used here, refers to "... a set of closely spaced secondary, planar, fabric elements that impart mechanical anisotropy to a rock without apparent loss of cohesion" (Dennis, in Bayly *et al.* 1977). The fabric elements, which are commonly called 'domains', are zones in which the original fabric of the rock has been altered by deformation processes responsible for creation of the cleavage. Domains have been variously referred to as bands, films, folia, selvages and stripes. Uncleaved, or less strongly cleaved, bands of rock that lie between adjacent domains are called microlithons (see Borradaile *et al.* 1982). 'Disjunctive' cleavage domains, in contrast to 'crenulation' domains, cut across pre-existing (sedimentary) layering in the rock without reorienting the layering (Powell 1979). The principal characteristics that have been used to classify disjunctive cleavage are; (1) the spacing of cleavage domains and (2) the morphology of the domain (Borradaile *et al.* 1982).

Disjunctive cleavage has been subdivided into two principal types based on domain spacing. Spaced cleavage is recognized where domains and microlithons can be distinguished in hand specimen or in outcrop, and slaty cleavage where adjacent domains cannot be distinguished without the aid of optical microscopy; Powell

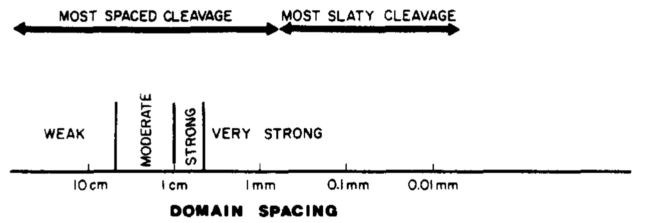

Fig. 1. A classification of cleavage based on spacing between domains. Powell (1979) subdivided cleavage into spaced and slaty, whereas Alvarez *et al.* (1978) subdivided spaced cleavages based on the spacing between domains.

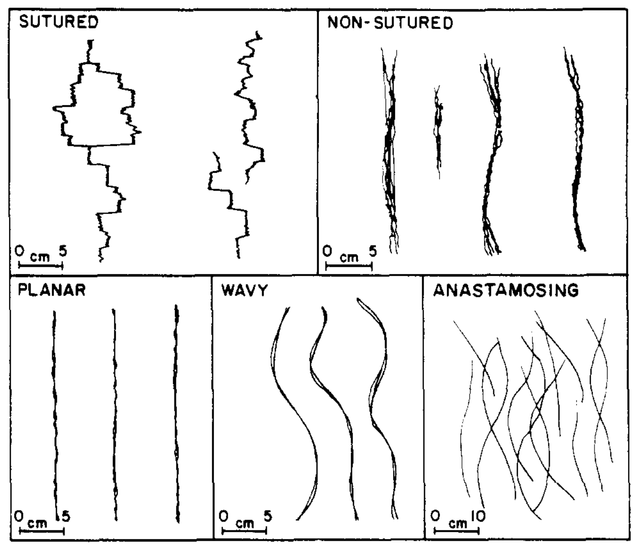

Fig. 2. Patterns of disjunctive cleavage cutting beds.

(1979) noted that most slaty cleavage has a domain spacing of less than 1 mm. Spaced cleavages can be further labeled, based on domain spacing, as weak, moderate, strong and very strong (Alvarez *et al.* 1978). A combination of Powell's (1979) and Alvarez *et al.*'s (1978) classification schemes is presented as Fig. 1. The term 'slaty', as used in Fig. 1, refers only to domain spacing. Spaced cleavage has been variously referred to as tectonic striping, solution cleavage, fracture cleavage, stylolitic cleavage, zonal cleavage, fissuring, reticulate cleavage, false cleavage, and spaced-solution cleavage. Very widely spaced domains, which commonly have pitted surfaces, are also called stylolites. The variety of terms listed above can be confusing, for some of these terms imply a mechanism of origin and others indicate morphological characteristics. The term 'fracture cleavage', for example, implies that disjunctive cleavage is a consequence of brittle failure of rock, which is generally not the case. Usually, the non-genetic term 'spaced cleavage' can be used in preference to the others.

The morphology of cleavage domains is defined by three parameters; (1) small-scale topography of the domain, (2) trace length of the domain as measured on the outcrop and (3) topography of the enveloping surface of the domain. The first and third parameters proved to be the most definitive in describing cleavage in our field areas. We suggest using the term 'sutured' for pitted (tooth-like in cross section) domains and 'non-sutured' for smoother non-pitted domains, following Wanless (1979), and the terms 'planar' and 'wavy' to describe the overall topography of the enveloping surface (Fig. 2). If adjacent domains are intertwined, the cleavage is anastomosing. The terminology used in Fig. 2 differs slightly from that of Powell (1979) and Borradaile *et al.* (1982).

PROCESSES INVOLVED IN CLEAVAGE FORMATION

Evidence from both field and laboratory studies (Weyl 1959, Plessman 1964, Williams 1972, Nickelsen 1972, Carannante & Guzzetta 1972, Geiser 1974, Holeywell & Tullis 1975, Groshong 1975, 1976, Alvarez *et al.* 1976, Platt 1976, Rutter 1976, 1983, DeBoer *et al.* 1977, McClay 1977, Engelder *et al.* 1981, Gray 1981a, Borradaile *et al.* 1982, Engelder 1982, 1984, Nelson 1983) demonstrate that in sedimentary rocks deformed at shallow depths development of disjunctive cleavage is dominantly a result of rock–water interaction, but may also involve grain rotation and grain-boundary sliding. The following consideration of cleavage formation emphasizes four types of rock–water interaction: pressure solution, free-face dissolution, diffusion in fluid films and advection by bulk fluid circulation. These processes depend in part on the chemical effect of water (i.e. the capability of water to dissolve mineral grains; Sorby 1863, Stockdale 1926) and in part on the mechanical effect of water (i.e. the involvement of fluid in controlling effective stress; Terzaghi 1943, Handin *et al.* 1963, Engelder & Scholz 1981).

Pressure solution

The association of cleavage with tectonic deformation has long been considered evidence that cleavage forms under conditions of non-hydrostatic stress (e.g. Sorby 1853, 1856, 1863). Under such conditions, pressure solution is thought to be the dominant mode of rock–water interaction (e.g. Rutter 1976, Paterson 1973), and, as a consequence, studies of rock–water interaction pertaining to cleavage development have focused on pressure solution. Innumerable examples of truncated grains and fossils in association with cleavage domains of both coarse- and fine-grained rocks (e.g. Nickelsen 1972, Holeywell & Tullis 1975, Beutner 1978, White & Knipe 1978, Engelder 1979, Gray 1981a, Borradaile *et al.* 1982) have been taken as evidence of the activity of pressure solution. The sedimentological literature concerning diagenetic compaction and formation of bedding-parallel stylolites provides similar examples (e.g. Stockdale 1922, 1926, Dunnington 1954).

Pressure solution refers to the process of dissolution at grain-to-grain contacts under conditions that cause the rate of dissolution to be controlled by the magnitude of normal stress across the contact; grains dissolve most rapidly where the normal stress is greatest. Subsequent to removal from the grain surface, dissolved ions then diffuse through fluid films away from the point of dissol-

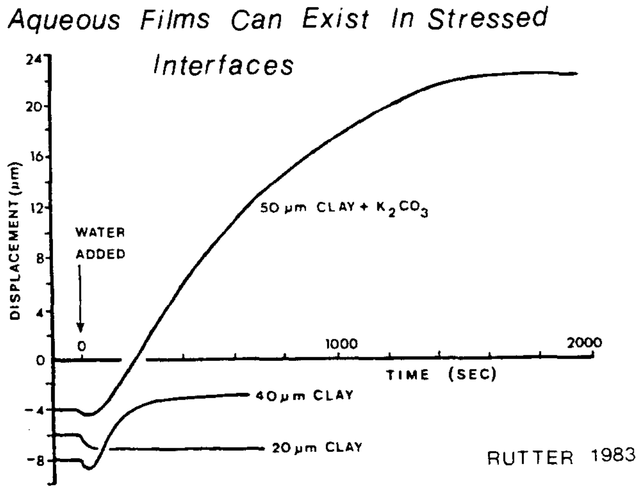

Fig. 3. Behavior of a kaolinite film between loading anvils. On wetting of a 20 μm thick clay-layer expansion against 100 MPa applied pressure is observed. The initial expansion against the same pressure is followed by further compaction on wetting a 40 μm thick layer. 50 μm thick films of zero porosity mixtures of potassium carbonate and kaolinite wetted at 100 MPa pressure also show expansion followed by compaction (after Rutter 1983).

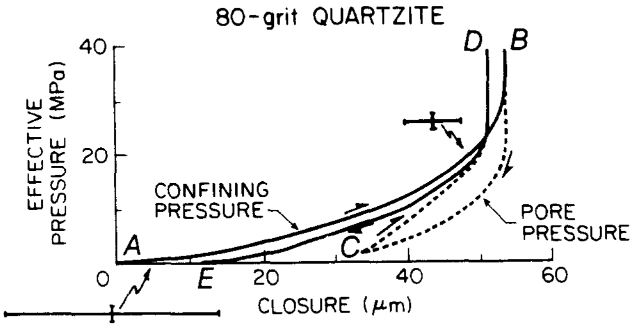

Fig. 4. Experiment on artificial joints in Cheshire quartzite to show the effect of fluid pressure on mechanically opening and closing joints (cleavage planes) (after Engelder & Scholz 1981). Joint aperture is plotted vs effective pressure where the solid curve represents closure of a joint during adjustment of confining pressure (*A–B*; *D–E*) and the dashed curve represents closure during adjustment of pore pressure (*B–C*; *C–D*). The experiment shows that confining pressure is more effective at changing the closure distance than the pore pressure. Error bars shown at two effective pressure levels.

ution and either precipitate at sites of lower normal stress, or pass into the pore fluid of the rock. The rate-controlling step of the pressure-solution process is thought to be this diffusion through fluid films (Rutter 1976), and based on thermodynamic arguments, the driving force for the diffusion accompanying pressure solution is a chemical potential gradient created by variations in the magnitude of normal stress at grain–grain contacts (Durney 1972, Elliott 1973, DeBoer 1977, Robin 1978, Green 1984). Apparently, chemical potential of the solute in solution is affected more by variations in normal stress at grain–grain contacts than by variations in elastic strain energy of the crystal lattice beneath the contact (Robin 1978) or by temperature (Rutter 1976). The above proposals have been summarized by a constitutive flow law for pressure solution in which the rate of deformation resulting from pressure-solution activity is directly proportional to normal stress but is largely independent of temperature (Rutter 1976).

A full understanding of the pressure-solution deformation mechanism requires knowledge of the mechanical behavior of fluid films along grain-to-grain contacts. If fluid along these contacts behaves like free-fluid in pore spaces, it could exert hydrostatic pressure against grain boundaries. Thus, the effective stress across grain-to-grain contacts would obey the laws of Terzaghi (1943) and Handin *et al.* (1963) (simply stated, $P_{\text{effective}} = P_{\text{confining}} - xP_{\text{pore pressure}}$, where $x = 1$) and conceivably approach 0, thereby bringing pressure-solution activity to a halt. To overcome this problem, it has been proposed that fluid along grain-to-grain contacts does not behave like free pore fluid and thus that simple laws of effective stress do not completely describe the behavior of fluid during pressure solution. According to current theories (e.g. Rutter 1983) fluid films at grain contacts have strength, in that they can resist being squeezed from between two surfaces in contact under high normal stress and thus can sustain a shear stress.

Experimental data concerning the strength of grain-boundary fluid films and free pore-fluid are not abundant. Rutter (1983) studies fluid-film strength by wetting a clay layer (20 μm of kaolinite) that had been placed between loading anvils. Upon wetting, the clay layer expanded by more than 1 μm against an applied normal stress of 100 MPa (Fig. 3). Only if the fluid film had strength (of at least 100 MPa) could it have expanded against the applied normal stress. Free pore fluid in Rutter's experimental set up could have exerted hydrostatic pressure equal in magnitude to atmospheric pressure, for the system was open to the atmosphere. In an effective stress law governing this situation, x would be much greater than 1. Results from Engelder & Scholz (1981), who studied closure of artificial joints in Cheshire Quartzite, indicate that under circumstances such that either fluid films already exist between grain contacts or fluid is unable to inject between grain-to-grain contacts, the coefficient x in the effective stress law is less than 1. An incremental increase in pore pressure does not open a joint by as much as the same incremental increase in confining pressure closes the joint (Fig. 4). This result indicates that a change in pore pressure within the pore spaces is not accompanied by an equal change in pore pressure along grain-to-grain contacts. The contrast in values of x that are implied by comparison of Rutter's (1983) results with those of Engelder & Scholz (1981) probably reflects the contrast between mechanical behavior of fluid films and that of free pore fluid. The inability of fluid to enter between grain-to-grain contacts in the Cheshire Quartzite may reflect contrasts in surface charge characteristics of quartz vs clay grains.

At elevated temperatures, fluid films on grain surfaces may dissipate. Experiments of DeBoer *et al.* (1977) showed that pressure-solution deformation of quartz grains in an aggregate under compression at temperatures of 260 and 330°C ceased when the experimental capsule drained (Fig. 5). If a fluid film attached to the quartz grains could sustain a high shear stress at these temperatures, draining of the capsule would not result in removal of this film.

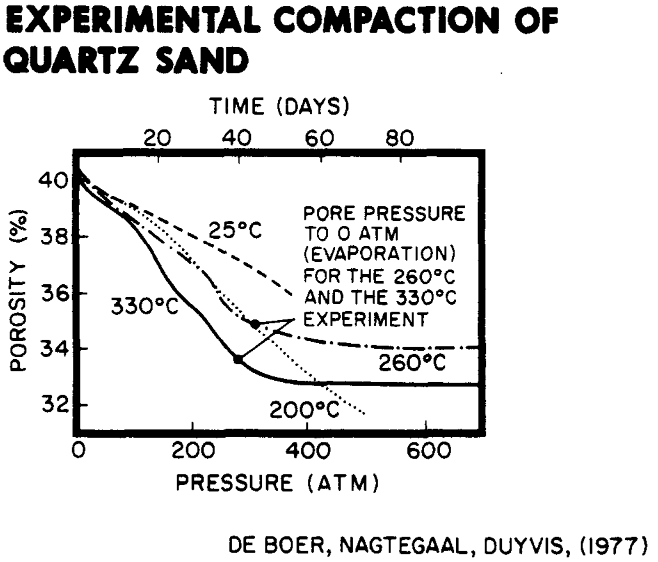

Fig. 5. Compaction curves for sand under increasing load. Water evaporated from experiments at 260 and 330°C after about 45 days at which time compaction stopped (after DeBoer *et al.* 1977).

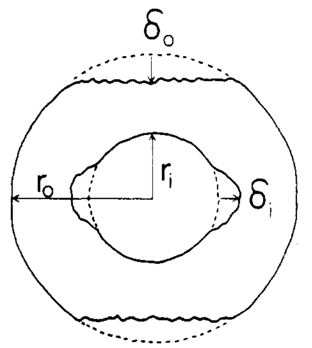

Fig. 6. Shape changes of crinoid columnals by pressure solution (δ_o) and free-face dissolution (δ_i). r_o and r_i are initial external and internal radii (after Engelder 1982).

Free-face dissolution

Pressure solution is not the only type of rock–water interaction to occur during tectonic deformation (e.g. Engelder 1982). Free-face dissolution, the process by which mineral grains dissolve into free fluids without the influence of non-hydrostatic stress at grain–grain contacts (e.g. Weyl 1958, 1959), may also occur in rocks at shallow crustal depths. During free-face dissolution, ions detach from the grain surface and then diffuse through a boundary layer directly into free fluid (e.g. Garrels *et al.* 1949). The rate-controlling step of this process is either the rate of flushing of fluid past the grain surface or the rate of diffusion of ions off the grain surface (e.g. Kaye 1957, Berner 1978, Weyl 1958, Helgeson 1971, Bosworth 1981, Lasaga 1984).

Good evidence for the occurrence of free-face dissolution has come from studies of crinoid columnals in cleaved siltstones in the Appalachian Plateau (Engelder 1982). The axial canal of these columnals dissolved at sites where the crystalline lattice adjacent to the canal surface was most highly strained, as indicated by the presence of mechanical twins. Dissolved segments of the canal surface are at right angles to segments of the outer surface of the columnals subject to pressure solution (Fig. 6). Dissolved segments of the outer surface define micro-cleavage domains (planes) that are approximately perpendicular to the maximum compression direction indicated by dynamic analysis of twinning in the crinoid (Engelder 1979). Dissolution of the outer surface of these crinoids appears to have occurred by pressure solution, but the orientation of dissolved surface segments within the canal, and their association with zones of relatively high twinning strain suggest that dissolution of the canal wall was controlled by strain energy accompanying twinning, not by the magnitude of normal stress across the grain surface. Such behavior is compatible with the occurrence of free-face dissolution, not pressure solution (e.g. Bosworth 1981).

The activity of free-face dissolution is also indicated by distribution of solution pitting on calcite grains of cleaved lime wackestones of the Hudson Valley (Marshak & Engelder 1985). Grain-surface pitting in these wackestones is not restricted to grain boundaries parallel to cleavage domains, suggesting that dissolution was not strictly controlled by stress distribution. Meike (1983), in studies with the transmission electron microscope, showed that dissolution of calcite grains along stylolites does not occur at grain–grain contacts, but rather occurs in the pore space between grains. Additionally, Meike showed that the distribution of dissolution pitting in these stylolites appears to be controlled on a small scale by specific crystallographic planes of calcite grains, and that dissolution can be enhanced by the occurrence of dislocations.

In order for the process of free-face dissolution to occur, soluble grains must be in intimate contact with unbound fluid. The occurrence of such unbound fluid is supported by recent studies. Carbonate rocks commonly contain a microporosity (Heling 1968) and work with transmission electron microscopes shows that tiny pockets of fluid reside along incompatible boundaries between grains in silicic rocks (White & White 1981). Free-face dissolution does not appear to result in the development of large voids in the rock; matrix grains are apparently pressed into dissolution pits by tectonic compaction.

Diffusion and advection

Removal of ions from grain boundaries by either pressure solution or free-face dissolution involves diffusion. Calculations of diffusion rates in shallow crustal rocks suggest that movement of ions by diffusion is too slow to permit removal of dissolved ions from the local rock system (Fletcher & Hoffman 1974). Therefore, in a rock system closed to advection, cleaved rocks should contain abundant mineral overgrowths, veins and late-stage cements.

At some localities, there is evidence that material dissolved by pressure solution is precipitated locally. For example, Heald (1956) observed that the amount of

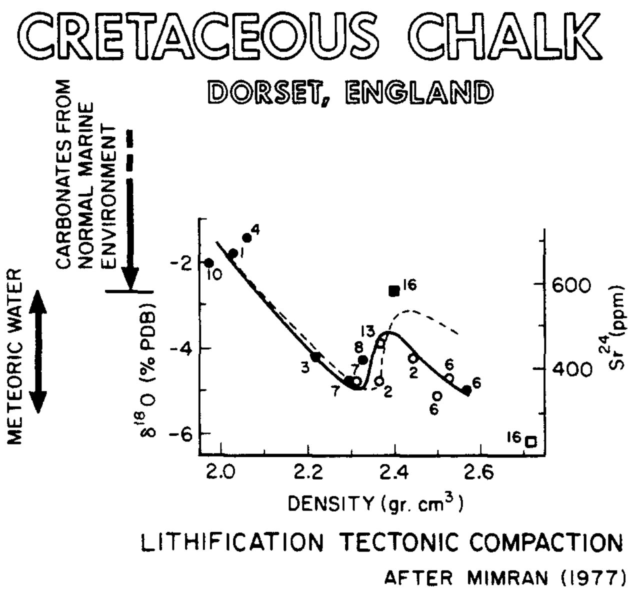

Fig. 7. Relationship between rock density and δ^{18} values for Dorset chalks and calcite veins from exposures of Corfe Castle chalk (square). Dashed line indicates the relationships between density and Sr^{2+} concentration for Dorset chalk (after Mimran 1977).

silica occurring as secondary quartz or cement in the St. Peter Sandstone was nearly equal to that dissolved from the original sand as a result of pressure solution (see also Manus & Coogan 1974). Mitra *et al.* (1984) present an example where material removed from cleavage domains apparently precipitated in contemporaneous veins in the same rock. However, many cleaved units, such as the Devonian limestones of the Hudson Valley (Marshak 1983) do not contain sufficient 'sinks' within diffusion range. In such localities, material dissolved in cleavage domains must have been removed from the local rock system. The only process that could act fast enough to accomplish such removal is advection by bulk circulation of large volumes of pore fluid (Engelder 1984). The concept that fluid circulation is involved in solution deformation of rocks is implicit in Stockdale's (1922) discussion of stylolites, for he proposed that stylolites initiated along fractures or bedding planes where circulation of CO_2-charged groundwater occurred most freely.

Inferences concerning the source of fluids involved in circulation during deformation and concerning the ultimate sink for the dissolved ions come from study of bulk chemical changes associated with cleavage development (to be discussed later), from estimates of volume-loss strain, and from isotopic analyses. Calcite solubility data suggest that downward circulation of meteoric water was necessary for the 35% volume-loss strain within the limestones of the Apennines, but that circulation of connate or dehydration water derived from the dewatering of shales was sufficient for the development of cleavage and the accompanying volume loss strain in the Appalachian Plateau (Engelder 1984). In both situations the rock–water system involved more than 1 km^3 of rock which means that the circulation path must be of comparable dimensions. Study of oxygen-isotope geochemistry in chalk from southern England (Mimran 1977) indicates that δ^{18}O values decrease with an increase in chalk density (Fig. 7). Uncompacted chalk has initial δ^{18}O values, equivalent to carbonates formed within normal marine environments, but calcite from chalk samples with a density of 2.3 g cm^{-3} has δ^{18}O values approaching those of calcite formed in meteoric water. These results imply that removal of calcite from the chalk during compaction involved interaction with meteoric water.

Circulation of large volumes of fluid from outside the local rock system, as described above, is one way that fluid undersaturation can be maintained during deformation so that large volume-loss strain can occur. Etheridge *et al.* (1984) suggest fluid undersaturation may also be maintained by precipitation of the dissolved mineral species in nearby veins by the crack–seal mechanism. These veins need not be within diffusion range, but can be in adjacent rock units. In concurrence with this proposal, Marshak (1983) observed that abundant veining occurs in uncleaved lime grainstone units adjacent to cleaved lime wackestone units in the Hudson Valley.

Truncation, rotation and growth: development of preferred orientation

It is a common observation that grains within cleaved rocks, even the very low-metamorphic grade sedimentary rocks under discussion in this paper, exhibit preferred orientation in that they are aligned with the plane of cleavage domains. This alignment can be of two types: grain-shape alignment, determined by parallelism of the long axis of grains with cleavage domains, or crystallographic alignment, determined by parallelism of specific mineralogical axes with the domain. Typically, grains with preferred orientation lie within the cleavage domain, but in some rocks, particularly shales, a microlithon fabric also develops.

Three processes can contribute to development of preferred orientation. Grain-shape alignment can result from truncation of grains by pressure solution (i.e. grain surfaces parallel to domains are removed, leaving relics that are elongate in the plane of the domain) (e.g. Holeywell & Tullis 1975, Beutner 1978), by mechanical rotation of inequant grains into the plane of the domain, or by growth of beards (e.g. Bell 1978, Beutner 1978) and optically continuous overgrowths. Crystallographic alignment can result from recrystallization (which does not occur to a large extent in rocks which developed cleavage at very shallow depths), grain growth (i.e. growth of authigenic clays and growth of beards), and mechanical rotation (if the shape of the grain reflects its crystallographic axes, as is the case for phyllosilicates). The relative importance of the mechanisms listed above is highly variable.

Originally, mechanical rotation was considered to be of major importance; it was considered that cleavage formed in rocks during dewatering, such that grains were essentially suspended in fluid and could easily rotate (Maxwell 1962). This idea, in which grain rotation has been tied to dewatering of unlithified sediments, has been generally discounted (e.g. Beutner 1978), however, rotation of grains during volume-loss strain of lithified strata remains a possibility. If framework grains dissolve, either by pressure solution or free-face dissolution, and if the dissolved ions are removed from the local rock system with development of void space in the rock, compaction of the rock matrix must occur. It seems likely that such compaction would result in plastering of inequant grains against the surfaces of either the cleavage domain or of relict framework grains. Thus, the inequant grains would assume an orientation parallel to the domain. This process has been compared to a 'house-of-cards' collapse by Gray (1981a).

In recent years much evidence has accumulated supporting the concept that grain truncation is a very important process in development of preferred orientation. Direct observation of truncated quartz or calcite grains has been reported by Gray (1981a) and in several of the articles in Borradaile *et al.* (1982). Truncation of phyllosilicates during formation of slaty cleavage in shales is more difficult to document, because of small grain size, but has been reported by Holeywell & Tullis (1975), Bell (1978) and Beutner (1978) among others.

Grain-boundary sliding

Grain-boundary sliding may also play a role in the deformation of low-grade rocks when intragranular strain in many rocks cannot account for the total strain observed in the rock (Borradaile 1981). Part of this discrepancy in strain can be accounted for if part of the total strain is accommodated by movement of grains past one another. Such movement could be aided by high fluid pressures which might decrease the effective stress across grain boundaries. If the sliding is restricted to discrete zones, a cleavage could be produced. Direct evidence for sliding on grain boundaries is difficult to detect. Borradaile (1981) presents several examples of circumstantial evidence for the occurrence of particulate flow, but because of the lack of direct evidence, the significance of this process remains controversial.

CONTROLS ON DISTRIBUTION AND MORPHOLOGY

Cleavage is not distributed uniformly in deformed terranes, and, as implied by earlier discussion, can occur with a range of spacing and domain morphology. Cleavage morphology and distribution are controlled by several parameters including: rock composition, structural position and deformational environment (e.g. Marshak 1983). Cleavage distribution probably also depends on pore-fluid chemistry (Beach 1982) and amount of fluid circulation (Engelder 1984).

Lithologic control: the role of clay

At a given location, different lithologies typically develop different structures. For example, pure quartzites or pure limestones deform predominantly by intracrystalline mechanisms (resulting in undulose extinction and deformation bands in quartz, and in twin lamellae in calcite), whereas impure (clay-rich) sandstones and limestones develop cleavage (e.g. Heald 1956, Weyl 1959, DeBoer 1977, Wanless 1979, Morris 1981, Marshak & Engelder 1985). Pressure-solution activity in the St. Peter Sandstone was localized in association with clay in the matrix (Heald 1956) (Fig. 8). Heald (1956) concluded that ". . . pressure solution may occur without clay, but if other conditions are favorable, clay accelerates the process." It was also observed that clay–matrix content in sandstones and limestones controls domain spacing (spacing decreases as clay content increases), and domain morphology (rocks with high-clay contents typically have non-sutured domains) (e.g. Wanless 1979). Shales, which are composed largely of clay, are generally susceptible to cleavage development and typically have closely spaced cleavage domains.

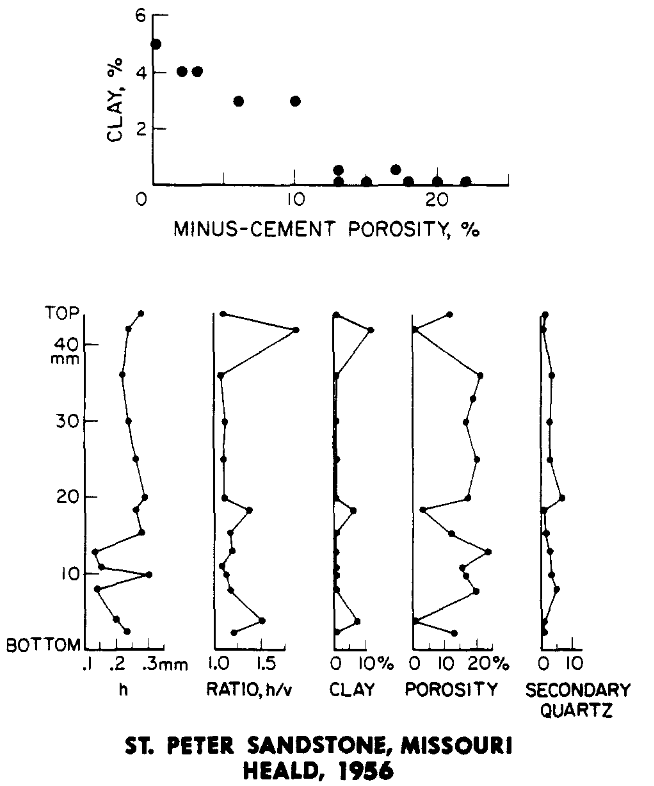

Fig. 8. Variation in grain size and shape, clay content, porosity, and secondary quartz in a single specimen of St. Peter Sandstone, Missouri. h, horizontal intercept of grains; v, vertical intercept of grains. Low porosities and high h/v ratios are due to greater pressure solution in the clay-rich portions of the specimen. Top and bottom of the bed are indicated (after Heald 1956).

Clay increases the susceptibility of a sedimentary rock to cleavage development by rock–water interaction. (Other components such as organic material may also affect this susceptibility; e.g. Nelson 1983). Several ideas have been put forth to explain the manner in which clay is involved in rock–water interaction. There have been proposals that clay acts as a chemical catalyst or that it affects the pH of the fluid in the rock, but many recent studies emphasize that the role of clay is textural, not chemical. Weyl (1959) suggested that clay flakes act as loci for water films which "... afford a path for the diffusion of pressure-dissolved solute . . . ," and further, that any insoluble grain, not just clay, which can attract a water film and is small relative to the soluble clasts of the rock, can enhance solution deformation. Analysis of strain partitioning as a function of the proportion of clay–quartz matrix in a limestone by Marshak & Engelder (1985) indicates that the presence of just a small quantity of clay does not make a limestone susceptible to solution deformation; at least 10% clay–quartz matrix must be present (cf. Wanless 1979). It is suggested that, as proposed by Weyl (1959), the water associated with the clay flakes provides a diffusion pathway away from sites of dissolution (i.e. clay plays a textural not chemical role). Furthermore, an interconnecting network of clay must exist in order to link the sites of dissolution with the free-fluid system of the rock, where advection further enhances cleavage development. At least 10% of clay must be present for the necessary interconnectivity.

Structural control

Rock composition is not the only factor controlling cleavage distribution, for a single rock type can contain cleavage with a range of domain spacing. Alvarez *et al.* (1978), among others, showed that domain spacing in a single unit is a function of strain; the higher the strain, the more closely spaced are the domains. Typically, cleavage domains become more planar, and microlithon fabric becomes stronger in zones of high strain (Marshak 1983). The structural positions at which large cleavage strains occur could have been locations of stress concentrations, for an increase in stress acts to increase the strain rate due to pressure solution (Rutter 1976), but, alternatively, these positions may merely have been zones where deformation continued for a longer period of time. Marshak & Engelder (1985) show that cleavage intensity is particularly strong in certain structural positions of a fold-thrust belt, such as in fault interaction zones, along detachment horizons, and on overturned fold limbs.

DEPTH CONTROL ON DISJUNCTIVE CLEAVAGE DEVELOPMENT

Several different processes can contribute to the total strain that develops during rock deformation. The manner in which strain is partitioned among these processes is a function of many parameters including rock composition, stress, temperature, and rate of fluid circulation. Disjunctive cleavage in sedimentary rocks forms when rock–water interaction is the dominant deformation process, and intracrystalline flow is suppressed. The effect of rock composition on strain partitioning has been discussed above. This section first discusses the effect of variations in stress and temperature, which are essentially proportional to variation in depth, on strain partitioning, then summarizes examples from the literature that relate cleavage development to burial depth.

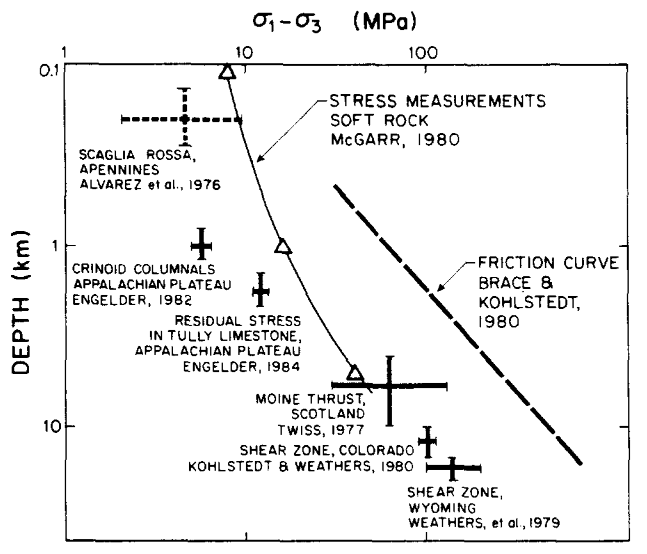

Fig. 9. Plot of differential stress vs depth within the crust of the earth. See text for details.

Many of these examples rely on correlation of cleavage distribution with illite crystallinity.

Stress and temperature effects

Comparison of flow laws (illustrated by deformation maps, e.g. Rutter 1976) shows that intracrystalline deformation mechanisms (such as twin gliding) have a yield strength (Jamison & Spang 1976, Groshong 1975) whereas pressure solution (one type of rock–water interaction responsible for cleavage formation) does not (Rutter 1976, 1983). As a consequence, at low stresses, pressure solution will be operative whereas intracrystalline deformation will not be operative. In addition, deformation mechanisms contributing to intracrystalline flow are thermally activated, so variations in temperature with depth also affect strain partitioning. Both temperature and differential stress increase in magnitude with depth (Fig. 9), thus, partitioning of strain between pressure solution and intracrystalline flow is likely to vary with depth. Kerrich & Allison (1979) suggest that intracrystalline plastic flow becomes a dominant deformation mechanism in calcite at a temperature of 200–300°C, and in quartz at 300–450°C. If a geothermal gradient of about 30°C is assumed, the transition into the realm of intracrystalline flow, depending on lithology, occurs at a depth of 6–15 km. Thus, the formation of disjunctive cleavage is favored at depths less than 6–15 km. As a consequence of mineralogic control on temperatures at which intracrystalline flow becomes the dominant deformation mechanism, strain-partitioning ratios measured in deeply buried quartzites may equal strain-partitioning ratios measured in shallowly buried limestones (Fig. 10).

Within the range of deformation conditions under which rock–water interaction is the dominant deformation process, the functional dependence of this process on temperature is not well known. Rutter (1983) suggested that the rate of pressure solution is largely independent of temperature, but that it is linearly pro-

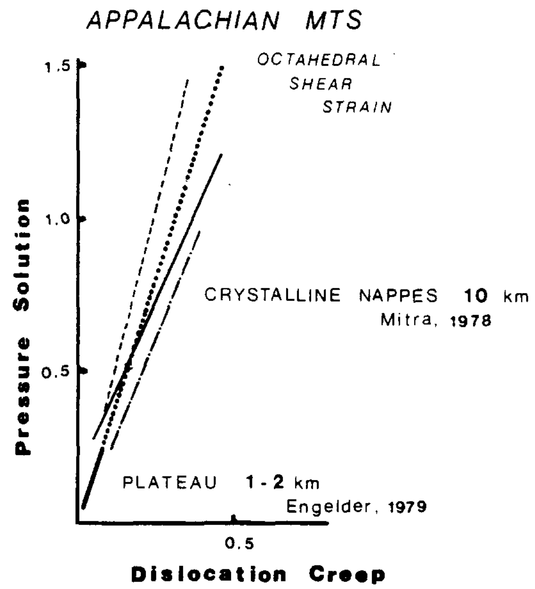

Fig. 10. Plot of the octahedral shear strain for pressure solution vs octahedral shear strain for either twinning of calcite (Engelder 1979) or dislocation creep of quartz (Mitra 1976). Mitra (1976) determined the ratio of pressure solution strain to dislocation creep strain in quartzites of the Blue Ridge region of the Appalachians. The ratio that he found is the same as that found by Engelder (1979) in a comparison of pressure-solution strain to twinning strain in calcite grains from the Appalachian Plateau, a region which was never buried as deeply as the Blue Ridge region. The coincidence of these ratios reflects lithologic control on strain partitioning. Dotted line is an extrapolation of data from the Appalachian Plateau into data from crystalline nappes.

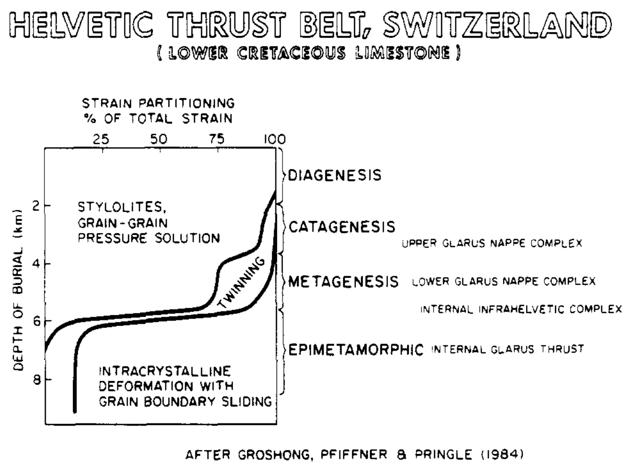

Fig. 11. Plot of strain partitioning vs depth for coarse-grained limestone in the eastern Helvetic zone of the Swiss Alps (after Groshong *et al.* 1984).

portional to magnitude of deviatoric stress. Free-face dissolution may also be affected by variations in pressure and temperature for mineral solubility is affected by these parameters. However, in many cases the effect of pressure and temperature variations on rates of free-face dissolution is overshadowed by the effect of chemical variations in the solvent or by rate of fluid flow.

Field examples of depth dependence

Strain partitioning and cleavage morphology in the Helvetic Nappes of Switzerland have been observed to vary markedly with depths of burial (Groshong *et al.* 1984). At burial depths less than 5 km, pressure solution is the dominant deformation mechanism (Fig. 11), but at burial depths greater than 5 km, other deformation mechanisms, including grain-boundary sliding and intra-crystalline flow, begin to dominante. This transition in deformation mechanism delineates that depth to which disjunctive cleavage is common within the crust of the earth.

Several studies have correlated cleavage distribution and morphology with illite crystallinity, because illite crystallinity provides an estimate of the temperature to which the rock has been subjected, and, with the assumption of a geothermal gradient, an estimate of depth. For example, variations in cleavage development as a function of illite crystallinity in the Wyoming fold-thrust belt (Mitra & Yonkee 1985) indicate that cleavage occurrence corresponds with the areal extent of overriding thrust sheets, and thus reflects the amount of overburden (temperature) at the time of deformation. Siddans (1977) documented the development of cleavage, metamorphic minerals and strain in a Jurassic marl from the external zone of the French Alps. Across the width of its exposure, the depth of burial of this marl, inferred from illite crystallinity, increases from 2.5 to 12 km. The 12 km depth arose from the emplacement of the Embrunias Nappes (Fig. 12). Shales with an illite crystallinity of around 10, roughly corresponding to a burial depth of 5 km, contain spaced cleavage, whereas shales with crystallinities of around 7, corresponding to burial depths greater than 5 km, contain slaty cleavage. The flattening strain in rocks with slaty cleavage was greater than that of the shales with spaced cleavage. In the Viséan shales of Morocco, there is a similar correspondence between cleavage domain spacing and illite crystallinity (Pique 1982). Spaced cleavage occurs in shales with a crystallinity between 6 and 12, corresponding to the diagenetic zone of metamorphism (Kübler *et al.* 1979) and a depth of burial of the order of 5 km (Fig. 13). Shales with an index lower than 6 contain slaty cleavage. Pique's results suggest that spaced cleavage is characteristic of deformation at relatively shallow levels and that a transition to a slaty cleavage occurs at between 5 and 10 km of burial.

Care must be taken in interpreting the above observations. The distribution of cleavage in both the French Alps and Morocco suggest that slaty cleavage characteristically forms at greater depths than disjunctive cleavage, but it is not clear if this effect follows from variations in strain or variations in pressure–temperature conditions. In addition, not all variation in illite crystallinity associated with cleavage development reflects variation in depth. Marshak & Engelder (1985) found variation of illite crystallinity corresponding with degree of cleavage development in a single outcrop, where depth variation could not have exceeded 45 m.

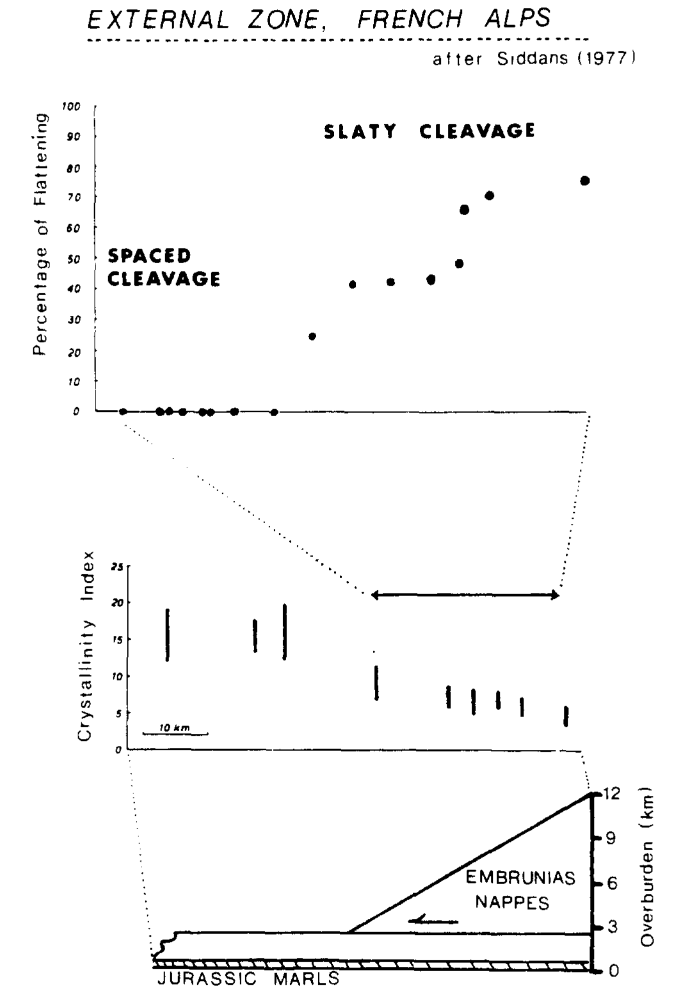

Fig. 12. Development of cleavage within the External Zone of the French Alps (Siddans 1977). The percentage of flattening and the crystallinity index are plotted against distance in a west (left) to east (right) section of Jurassic marls. The relative thickness and position of the Embrunias Nappes are indicated by the lower cross section. Overburden on top of the Jurassic marls varies from less than 3 km in the west to about 12 km in the east. Vertical to horizontal scale is about 2 to 1.

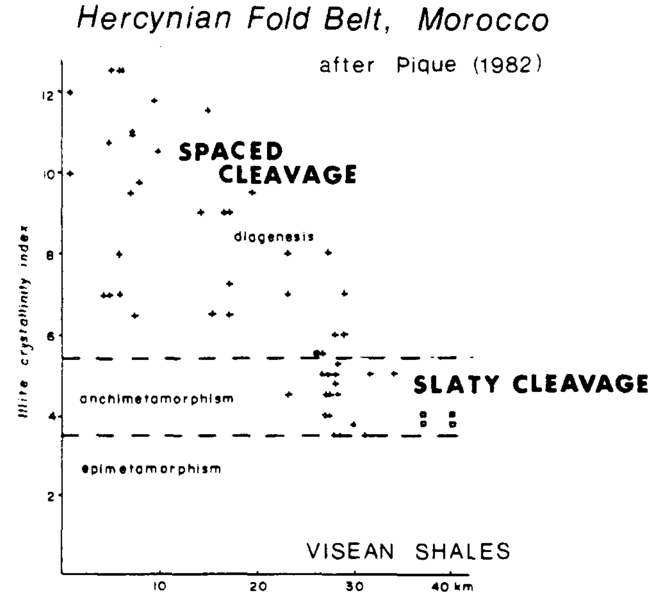

Fig. 13. Crystallinity index across a 40 km section of Viséan shales within the Hercynian Fold Belt of Morocco (Pique 1982). The cleavage associated with various grades of metamorphism are shown.

INITIATION OF CLEAVAGE

The question of how cleavage domains initiate is particularly enigmatic because once cleavage has formed, the original lithology is altered, and the feature of the original lithology that controlled the initiation site of the cleavage domain is destroyed. Two concepts have been put forth concerning the initiation of spaced cleavage domains. First, that domains initiate on pre-existing macroscopic discontinuities, and second, that domain-initiation sites are controlled by the grain-scale fabric of the rock. Slaty cleavage in shales may initiate as a pencil fabric.

Pre-existing discontinuities

In sedimentary rocks there are two types of discontinuity that may control cleavage distribution. Such discontinuities include sedimentary structures, such as mud-cracks or worm burrows, as well as fractures. These structures, in deformed regions, commonly display solution activity along them. For example, stylolites on which the axes of the solution pits are obliquely oriented to the surface are good evidence that the pressure-solution activity responsible for the pits did not control the orientation of the surface itself. Geiser & Sansone (1981) suggest, in fact, that pre-existing joint arrays control some cleavage geometries. Certainly, fractures are domains of enhanced permeability, and thus will initially be loci where dissolution occurs (e.g. Leith 1905, Lammers 1940), however, the spacing of cleavage domains in deformed areas is commonly less than the spacing of pre-existing fractures or sedimentary structures. Alvarez *et al.* (1978) found that, as strain increased, the number of cleavage domains per unit volume of rock increased. Thus, new cleavage domains developed as a deformation progressed; early-formed cleavage domains do not continue to grow in thickness indefinitely. Although some cleavage domains may initiate on pre-existing sedimentary structures, not all do.

Grain-scale fabric

Original grain-scale inhomogeneities in sedimentary rocks can control the sites at which cleavage domains initiate. For example, the truncated edges of crinoids columnals in siltstones of the Appalachian Plateau define microscopic cleavage domains that do not extend into the surrounding matrix (Engelder & Engelder 1977), and in lime wackestones of the Hudson Valley cleavage domains tend to thicken adjacent to larger clasts, indicating that the domain probably initiated adjacent to the clast (Marshak & Engelder 1985). Apparently, grain-scale inhomogeneities either effect fluid movement in the rock, or they may affect stress fields. As discussed above, stress concentrations may affect the rate of solution deformation. The anticrack model of Fletcher& Pollard (1981) suggests that once a cleavage domain initiates at a stress concentration, it propagates outward at its margin, and that the stress and displacement fields associated with the domain tip are identical to, but opposite in sign to, those associated with extensional crack tips. This latter hypothesis is largely supported by the variation of domain thickness (which is proportional to strain) as a function of distance from the domain tip.

The interconnectivity of the clay matrix in a rock appears to play an important role in determining the extent to which the rock is susceptible to the development of cleavage (Marshak & Engelder 1985). If clay interconnectivity is important, and the distribution of clay in rocks is not uniform, it follows that sites at which interconnectivity occurs can be sites at which cleavage initiates. Furthermore, it is possible that cleavage growth is a positive feed-back process; as the carbonate or quartz grains are removed, interconnectivity develops at additional sites, and new cleavage domains develop. Thus, the decrease in spacing between domains as a consequence of strain reflects both the removal of microlithon material and the initiation and growth of new domains.

Pencil cleavage

Weakly deformed shales often exhibit a crude fabric which is manifested by the separation of weathered rock into elongate or pencil-like fragments. Recent work (e.g. Crook 1964, Engelder & Geiser 1979, Reks & Gray 1982, Ramsay & Huber 1983) indicates that such pencil cleavage is a precursor to the development of slaty cleavage in shales. Shales possess bedding-parallel fabric as a consequence of compaction during burial loading. Lateral shortening due to tectonic compression gradually overprints the depositional fabric as progressive deformation occurs. When the fabric due to lateral compaction is approximately equivalent to that due to bedding compaction, the rock develops a pencil fabric. Thus, the pencils are defined by the intersection of bedding and cleavage. Support for this proposal comes from observations that, in some localities, cleavage-domain boundaries of the pencils display evidence that the domains truncate pre-existing bedding fabric (Reks & Gray 1982). Pencil lineations in shales are parallel to planar cleavage domains in adjacent limestones (Engelder & Geiser 1979). Where the fabric due to lateral compaction (and cleavage development) dominates over bedding compaction, spaced to slaty cleavage develops.

LITHOLOGIC CHANGES ACCOMPANYING CLEAVAGE DEVELOPMENT

Though cleavage can form in sedimentary rocks without destroying original bedding fabrics, rock compositions and mineralogy are affected to some degree by cleavage-related deformation. These changes are subtle and often involve the clay matrix of the rock; for clay is particularly susceptible to exchange reactions with ions diffusing through adhered water films. Bulk-rock alteration appears to reflect the occurrence of preferential dissolution and removal of specific mineral phases.

Mineralogical changes

Disjunctive cleavage can form without the occurrence of large-scale metamorphic transformation, but nevertheless mineralogic changes associated with ion-exchange reactions do occur in domains. In limestones, for example, zones of alteration of calcite to dolomite are, in some cases, spatially related to cleavage domains (Wanless 1979). Phyllosilicate sheets in slates are fairly durable (Oertel 1985), but are thought to have been affected by a number of alteration effects, including: (1) changes in illite crystallinity (e.g. Marshak & Engelder 1985); (2) changes in proportion of different clay minerals (Stephens *et al.* 1979, Gray 1981a, Knipe 1981); (3) dewatering of smectite (Marshak & Engelder 1985); (4) growth of authigenic crystals or of overgrowths (Beutner 1978, White & Knipe 1978, Knipe 1981, Meike 1983, Mitra *et al.* 1984); (5) transformation from one compositional type to another compositional type of the same mineral (e.g. Fe-rich chlorite to Fe-poor chlorite) (Knipe 1981); and (6) modification of micas into stable compositions indicative of higher grade (Knipe 1979).

Chemical changes

Several authors have employed electron beam techniques to examine chemical changes that accompany cleavage development (e.g. Gray 1981a, Erslev *et al.* 1983, Schwander *et al.* 1981, Lee *et al.* 1983, Marshak 1983). This work shows that major differences in composition distinguish cleavage domains from adjacent microlithons. For example, in limestones, the ratio between CaO and K_2O is much lower in the cleavage domains than in the microlithons. Such changes are clearly a consequence of preferential removal of calcite from the cleavage domain.

As discussed earlier, there is evidence that ions which dissolve in domains are removed from the local rock system by advection in circulating fluids during cleavage development. If certain minerals preferentially dissolve, and if such advection occurs, the bulk chemistry of the whole rock should change. (If cleavage development involved only redistribution of elements by diffusion in the local rock system, the bulk chemistry of the rock would not change). Marshak & Engelder (1985) examine variations in bulk chemistry resulting from cleavage development and provide preliminary data that suggest that certain elements (e.g. CaO in limestone) are removed from the local rock system during deformation.

The apparent mobility of some elements may be a function of the scale of observation. For example, on the scale of the bulk rock, CaO in limestone appears to be mobile, but SiO_2 appears to be inert (Marshak & Engelder 1985). Slaughter (1981), however, found, by comparing domain composition to microlithon composition in limestones, that, on the scale of the domain, both CaO

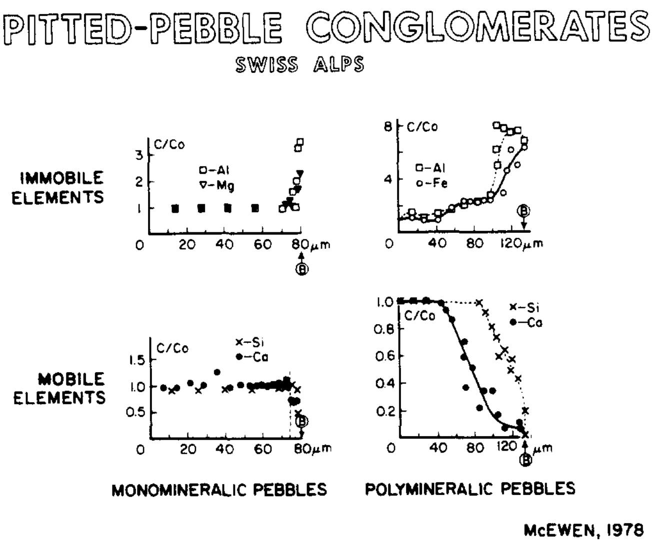

Fig. 14. Normalized element abundances measured using line scans of 50 μm along a traverse normal to a solution contact between two limestone pebbles, within the dissolved pebble. **B** marks the pebble boundary. Both mobile and immobile components are shown for monomineralic and polymineralic pebbles (after McEwen 1978).

and SiO_2 were mobile. Element mobility also appears to be a result of the bulk rock composition. Studies of large-scale volume loss accompanying development of slaty cleavage in the Martinsburg Slate (Wright & Platt 1982) indicated that SiO_2, which was immobile on a large scale in limestones, was mobile on a large scale in slates. Mobility of elements as a function of original rock composition is also indicated by studies of pitted-pebble conglomerates (McEwen 1978, Mosher 1981). In a conglomerate from Spain, changes in relative abundances of elements in the 100 μm wide zones next to pebble surfaces depended on the mineralogy of the pebbles in contact (Fig. 14); mobile components showed little tendency to diffuse from monomineralic pebbles but did diffuse from polymineralic pebbles. In the Purgatory Conglomerate of Rhode Island, Mosher (1981) found that pebble composition controlled the shape of pebble–pebble contacts, thus indicating variations in solubilities of the different pebbles as a function of the composition of neighboring pebbles (see also Trurnit 1968).

RELATION OF CLEAVAGE TO STRAIN

Magnitude of strain due to cleavage

Strain that forms in a rock as a consequence of cleavage development is generally difficult to measure, but a number of techniques have been proposed. Conventional strain measurement techniques can be used in rocks which contain appropriate strain markers. Siddans (1977), for example, documented tectonic flattening strain associated with the development of cleavage in the Alps by measuring deformed ammonites (Fig. 12), and Alvarez *et al.* (1978) were able to document layer-parallel shortening as a function of cleavage spacing in micritic limestones of the Apennines by using imbricated chert beds. In the Apennines, weak cleavage (Fig. 1) corresponded with up to 4% shortening, moderate cleavage with 4–25% shortening, strong cleavage with 25–35% shortening, and very strong cleavage with >35% shortening. Spacing of cleavage domains clearly is a function of strain, which in turn is a function of structural position, but, as emphasized by Marshak & Engelder (1985), spacing is also a function of lithology. Thus, qualitative scales that correlate shortening with domain spacing are lithology dependent.

In the absence of conventional strain markers, other methods to calculate strain due to cleavage have been used. Nickelsen (1972) and Engelder (1979) estimated strain by measuring truncated fossils and center-to-center strain analysis techniques can be applied at the same localities (Ramsay & Huber 1983). Another method involves measurement of selvage thickness and of the concentration of insoluble residue in the microlithons; a simple calculation will tell how much of the bulk rock was removed to produce the selvage thickness observed (assuming no local precipitation) and, therefore, will give shortening strain (e.g. Geiser & Sansone 1981). In practice, however, domain thickness is difficult to measure, and some of the material within the domain may not be insoluble residue or may have been transported into the domain by fluid circulation.

Volume loss associated with cleavage development

According to Wood (1974), Sorby (1853) recognized that significant volume loss can accompany deformation in rocks. Classically, volume loss was considered to be a result of compaction, and, thus, necessitated a density increase. Yet, the compaction mechanism places finite limits on the magnitude of volume loss that can occur; Ramsay & Wood (1973) suggested that only 10% volume loss can accompany the transition of fully lithified shales into slates, unless there is actually material loss. Thus large-scale volume loss associated with cleavage must represent removal of material from the local rock system.

One of the early reports of volume loss was by Cloos (1947) who observed that more highly deformed ooliths in limestones at South Mountain were of consistently smaller volume than lesser deformed ooliths. Recent authors have reported up to 50% volume-loss strains occurring in association with cleavage (Alvarez *et al.* 1978, Wright & Platt 1982, Beutner 1978, Mimran 1977). In most examples, there is no obvious local sink in which the dissolved material precipitated, and the volume loss is too large to represent only compaction. Therefore, in some cases, large-scale volume loss requires that the deforming region acts as an open system, with transport of dissolved material by distances of hundreds of meters to perhaps kilometers. The possibility of large transport distances eliminates the diffusive mass-transfer mechanism of Durney (1972) as a means of developing large-scale volume-loss strain, and requires that ions be transported by advection in a circulating fluid (e.g. Etheridge *et al.* 1984). Volume loss may require three steps: first,

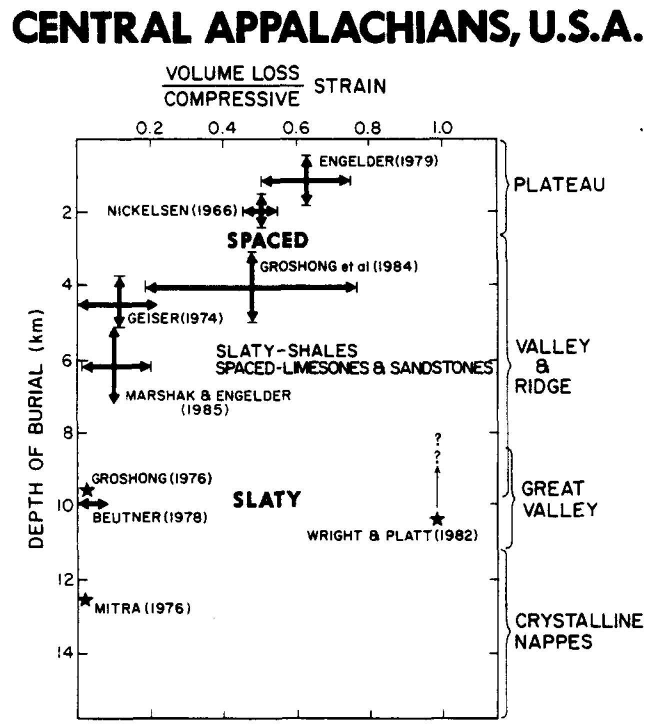

Fig. 15. Plot of ratio of volume loss over compressive strain vs depth for samples from the Central Appalachians, U.S.A. Also shown is the type of cleavage associated with various depths and volume losses.

removal of ions from grain boundaries (either by pressure solution or free-face dissolution), second, diffusion of the ions to pockets of free fluid, and third transport of ions out of the local rock system by advection in circulating free fluid (Marshak & Engelder 1985).

Because the solubility of rock in pore water is finite, large volume-loss strains require circulation of large quantities of undersaturated fluid. In fact, there are indications that the sum of pore water plus water derived from dewatering of hydrous minerals is insufficient for the process, and that meteoric water may be involved as well (e.g. Mimran 1977, Engelder 1984, as discussed above). Large volume circulation of pore fluid is most likely in the upper portion of the crust (Engelder 1984). Figure 15, a plot of ratio of volume loss over compressive strain vs depth of burial in the Appalachians, indicates that volume loss strain occurs dominantly at depths less than 5 km where the development of disjunctive cleavage is most common.

Orientation with respect to the strain ellipsoid

Cleavage fabrics in deformed terranes are widely used as indicators of the distribution and geometry of deformation associated with an orogenic event. Use of cleavage as such an indicator depends on the relation between cleavage and strain magnitude, and between the orientation of cleavage domains and the orientation of the strain ellipsoid. The relation between cleavage and the strain ellipsoid has been the topic of several papers during the past decade (Bayly 1974, Wood 1974, Borradaile 1974, Williams 1976, 1977, Treagus 1983). The results of this work indicate that cleavage may initiate nearly parallel to the XY plane of the finite strain ellipsoid, but that during non-coaxial progressive deformation, cleavage need not maintain parallelism with the XY plane. As an example, domains of weak cleavage of the Appalachian Plateau are parallel to the XY plane in the strain ellipsoid indicated by measurement of calcite twin lamellae (Engelder 1979). If these domains were to behave as material planes and be rotated by movement on faults or by flexural slip of beds during folding, small amounts of slip might occur on the domains and they would not maintain parallelism with the XY plane. The occurrence of such slip on cleavage domains has not been fully documented, and some examples of such slip have been shown to be a result of oblique pressure-solution shortening (Groshong 1975).

The relation of cleavage to the strain ellipsoid has had application most recently in the study of cleavage refraction (Treagus 1983, Helmstaedt & Greggs 1980) and of cleavage transection (Powell 1974, Stringer 1975, Stringer & Treagus 1980, Gray 1981b, Duncan 1985, Sanderson *et al.* 1985, Williams 1985). Cleavage transection, the non-parallelism between cleavage domains and fold axial planes as observed in plan view, appears to result from several processes. The most common explanation is that it occurs where cleavage and folding are not precisely contemporaneous, and thus represent somewhat different strain fields.

Relation of cleavage to rock anisotropy

Slaty cleavage imparts a penetrative, or nearly penetrative anisotropy to rocks. Numerous strategies have been proposed to measure and interpret such anisotropy in terms of strain (e.g. Wood *et al.* 1976, Rathore 1979, Kligfield *et al.* 1981, Oertel 1985). Work has concentrated on use of X-ray goniometry and on use of magnetic susceptibility anisotropy as tools to record the degree of development of cleavage fabrics. Results have been correlated with independent measurements of strain such as elliptical reduction spots, ooliths, and concretions (Rathore 1979, Wood *et al.* 1976, Kligfield *et al.* 1977, 1982). In sedimentary rocks with spaced cleavage, penetrative fabrics do not, by definition, exist, but commonly, weak fabrics are present in the microlithons between cleavage domains. Such fabric can be classified as 'random' to 'complete' (Powell 1979). Recent measurements by Oertel & Engelder (unpub. data) indicate that the fabric in the microlithons of shales with a weak spaced cleavage fabric, indicated by X-ray goniometry, is dominated by the effect of compaction during depositional loading, but does contain a component resulting from tectonic deformation.

Magnetic susceptibility anisotropy studies in cleaved limestones of the Hudson Valley (Kent 1979, Marshak 1983) indicate that in zones of weak to moderate cleavage, the magnetic susceptibility anisotropy ellipsoid is oriented such that its long axis is parallel to the strike of cleavage, its intermediate axis is perpendicular to cleavage strike and is in the plane of bedding, and its short axis is in the plane of cleavage and is perpendicular to bedding. These preliminary results suggest that the magnetic fabric represents a composite of original depositional fabric and tectonic fabric. Measurements from more highly deformed rocks indicate that the intermediate and short axes interchange. In rocks with slaty cleavage, the short axis of the ellipsoid is perpendicular to the plane of cleavage (e.g. Rathore 1979). Kligfield *et al.* (1982) have cautioned that development of magnetic fabric in a rock is a complex process, and thus that direct correlation of the magnitude of susceptibility ellipsoid axes with strain ellipsoid axes is dangerous.

RELATION OF CLEAVAGE TO STRESS

Three issues have evolved concerning the relation of cleavage to stress. First, what is the dependence of the deformation mechanisms responsible for cleavage development on the magnitude of stress? Second, what was the differential stress magnitude observed in terranes where cleavage has formed? Third, is there a geometrical relation between the orientation of cleavage domains and the stress ellipsoid?

Stress dependence of rock–water interaction

Free-face dissolution is a function of differential stress magnitude only in so far as plastic deformation under high differential stress increases the solubility of the crystalline lattice (Bosworth 1981, Engelder 1982). Pressure solution is a direct function of differential stress magnitude and behaves like a viscous process (e.g. Groshong 1975). Rutter (1976, 1983) presented flow laws that describe pressure solution, which are similar in form to those governing linear viscous flow by Coble Creep, and therefore, indicate that there may be no lower threshold at which pressure solution commences. Because the strain rate of pressure solution is proportional to stress, at higher deviatoric stresses the strain rate due to pressure solution may be higher, and thus, cleavage domains may form more rapidly.

Field examples of stress magnitude during cleavage development

Direct measurement *in situ* of stress magnitude during the development of cleavage is difficult and laboratory experiments that model cleavage formation have not yet been perfected. Bathurst (1971) proposed that the occurrence of pressure-solution deformation features in limestones that do not contain twinned calcite indicates that pressure solution can occur at stresses less than those required to cause twinning. This proposal has been confirmed by numerous studies (e.g. Alvarez *et al.* 1976, Marshak & Engelder 1985). The shear stress necessary to cause twinning is 10 MPa (Jamison & Spang 1976); thus, cleavage in sedimentary rocks can develop at stresses less than 10 MPa.

Further evidence concerning this problem comes from studies on the Appalachian Plateau. Engelder (1984) showed that the axial canals of crinoids from this region acted as stress concentrators, and inferred from observation of twinning adjacent to these canals that the crinoids were subjected to a differential stress of 6.2 MPa (Fig. 9). Rocks containing these crinoids are cleaved; thus cleavage in these rocks formed at stress magnitudes less than 6.2 MPa. Residual stress measured in rocks from the more deeply buried portions of the Appalachian Plateau is in the range of 11 MPa (Fig. 9), indicating that the cleavage in these rocks formed at stresses lower than 11 MPa (Engelder & Geiser 1984). Such measurements indicate that in terranes where cleavage is forming, stresses may sometimes be lower than average crustal stresses as estimated from techniques such as measurements *in situ* (McGarr 1980), measurement of fabrics such as dislocation density and subgrain size (Twiss 1977, Weathers *et al.* 1979, Kohlstedt & Weathers 1980), and laboratory estimates (Brace & Kohlstedt 1980) (Fig. 9). This contrast may reflect the lack of a lower threshold stress for the development of cleavage. The active formation of cleavage may relax stresses before they can build to exceed the frictional strength of the upper crust. This stress relaxation may reduce the number of earthquakes in the vicinity of active cleavage development.

Onasch (1983), in a review of studies pertaining to timing of formation of quartz deformation lamellae with respect to folding of sandstone, presented conclusions similar to those described for limestones above. In pure quartzites or calcite-cemented sandstones, deformation lamellae formed early without associated cleavage (e.g. Scott *et al.* 1965, Hansen & Borg 1962, Carter & Friedman 1965). In micaceous sandstones, cleavage formed early with respect to folding, and deformation lamellae formed only after the fold had become tight. These observations indicate that stresses necessary to initiate cleavage in quartz-rich sediments are lower than stresses required to initial deformation lamellae (cf. Heald 1956, Morris 1981). The absolute magnitude of these stresses, however, is not known.

Orientation of cleavage with respect to stress ellipsoid

Determination of the orientation of the stress ellipsoid with respect to cleavage at the time of cleavage development can only be done in areas where strain magnitude is very low. In such areas, dynamic analyses of quartz and calcite (e.g. Friedman 1964) and mapping of joints (e.g. Engelder & Geiser 1980) can be used to determine paleo-stress trajectories. On the Appalachian Plateau, Engelder & Geiser (1960) used the geometrical relation between cleavage domains and joints to show that the enveloping surfaces of cleavage domains formed perpendicularly to maximum principal stress, but, because of the wavy nature of weak cleavages, the orientation of domain segments can diverge from this orientation. If the domain is sutured, it is commonly observed that the axes of the solution pits remain parallel to one another (and presumably to the maximum compressive stress

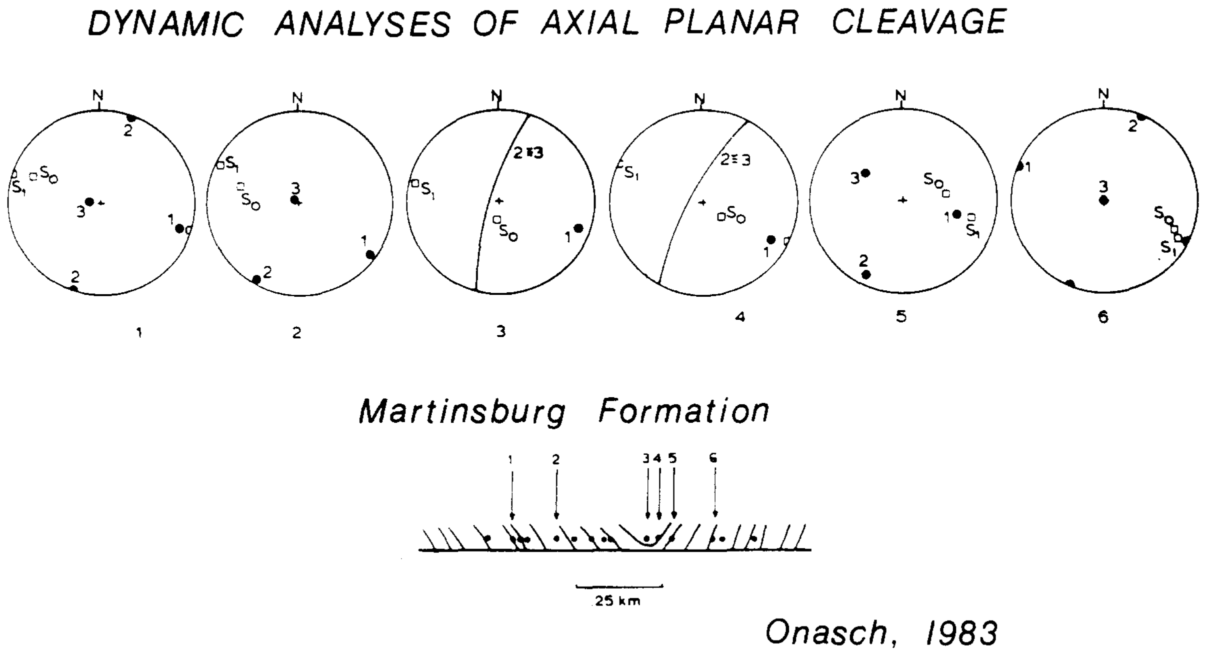

Fig. 16. Fabric diagrams from the dynamic analysis of deformation lamellae in samples of Martinsburg Formation, Maryland. Solid circles are stress axes with 1 the maximum compressive stress, the open squares are poles to bedding, S_0, and cleavage, S_1 (after Onasch 1983).

vector associated with the plane) regardless of variations in the orientation of the domain surface.

Onash (1983), Carter & Friedman (1965), Christie & Raleigh (1959) and Hara (1961) compare dynamic analysis measurements of quartz grains with measurements of cleavage orientations in folded sandstones and conclude that cleavages in folded sandstones form normal to maximum principal compressive stress (Fig. 16). Certainly such simple relationships will not be maintained during progressive deformation, and are complicated by variations in rock composition.

CHRONOLOGY OF CLEAVAGE DEVELOPMENT

The question of the time at which cleavage develops has two aspects: first, when does cleavage form with respect to lithification of rocks, and second, when does cleavage form with respect to other structures in a deformed terrane?

Timing with respect to lithification

The question of timing of cleavage with respect to lithification was hotly debated during the 1970s. Maxwell (1962) and Powell (1969), among others, proposed that the presence of cleavage-parallel sandstone dikes in some slates was evidence that slaty cleavage formed during dewatering of unlithified muds. This proposal was criticized by Geiser (1976) and Beutner (1980), among others, and the majority of evidence now supports the concept that in most localities, cleavage develops subsequent to complete lithification. There are examples, however, of cleavages that developed during soft-sediment slumping (Williams *et al.* 1969), and there have been efforts to document initiation of cleavage in partially lithified rocks of accretionary prisms (Moore & Geigle 1974).

Timing with respect to folds and faults

Cross-cutting relations in fold-thrust belts have demonstrated that cleavage initiates early during the development of such belts, and continues to develop throughout the development of other structures (Marshak & Engelder 1985). Evidence from the Appalachian Plateau is particularly convincing in showing that cleavage can initiate prior to folding. In this region, cleavage development appears to have accommodated layer-parallel shortening above a blind thrust, for cleavage only occurs in strata that lie above a weak salt horizon in which detachment movement is likely; apparently, cleavage tracked the advancing tip of the blind thrust (Geiser in prep.).

Evidence from the Hudson Valley (Marshak 1983) indicates that cleavage is reoriented and intensifies adjacent to faults (suggesting a genetic relation between faults and cleavage) and both cuts, and is cut by, slip lineations on fault surfaces, suggesting that the period of formation of these structures overlapped. Association of intense cleavage with fault intersection zones and overturned fold limbs indicates that cleavage developed contemporaneously with the generation of these structures, some of which formed during final stages of deformation in the belt. Onash (1983) has presented additional examples where cleavage developed subsequent to folding.

As a result of the extended history of cleavage development, geometrical relations between cleavage and other structures can be quite variable. Cleavage quite commonly has an axial-plane orientation, but regionally, cleavage may be inclined toward the direction of regional transport (e.g. Marshak 1983). On a smaller scale, cleavage refraction, reflecting strain adjustments during the progressive evolution of folds, and cleavage transection develop. Late-stage development of cleavage has been proposed as one explanation of cleavage transection (e.g. Stringer 1975).

Of particular interest is the local occurrence of multiple non-coaxial cleavages that appear to have formed during one deformation event (Boulter 1979, Mitra & Elliott 1980, Gray 1981b). Such phenomena have been attributed to variations in timing of cleavage development with respect to other structures and earlier cleavages. Boulter (1979) traced the development of multiple cleavages in the Stirling Range of Australia. In this locality, spaced cleavage initiated during an initial phase of folding dominated by layer-parallel shortening. This cleavage maintained an axial-plane relationship until body rotation of the limbs created a fold-limb dihedral angle of about 140°. The resultant 70° angle between spaced cleavage and bedding on the fold limbs was preserved by flexural slip until the dihedral angle decreased to about 100°, at which time, for mechanical reasons, flattening parallel to the axial plane occurred and a second, mica-film slaty cleavage developed.

CONCLUSIONS

Concepts concerning disjunctive cleavage have evolved rapidly during recent years. Progress has been notable in research on understanding the relation of cleavage to strain, the morphological characteristics of cleavage, the chronology of cleavage development, and controls on the distribution of cleavage. An appropriate terminology for cleavage classification and description is now widely in use. Perhaps the most exciting direction in current cleavage research concerns the nature of deformation mechanisms involved in initiation and growth of cleavage domains. Much of this research is focused on the unique aspects of rock–water interaction at shallow depths, for it appears that cleavage formation occurs in association with pore-water circulation under low differential stresses and relatively low temperatures. Diffusion of dissolved ions alone cannot account for the large volume-loss strain that is commonly associated with cleavage formation. Additional study requiring input from numerous data sources is needed to refine the constitutive and mass-balance equations which characterize rock–water interaction, and to understand better the dependence of such interaction on stress.

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