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Pressure-solution structures in a granite

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Abstract—Syntectonic two-mica granites in northwest Spain have suffered regional Late Variscan shear deformation. The deformed Corcoesto granite shows a marked planar and linear fabric resulting from ductile behaviour and contains quartz veins at about 60° to the foliation. Offsets of these quartz veins along shear planes are commonly contrary to the general sense of shear and thus provide evidence for pressure-solution processes.

INTRODUCTION

A VAST body of literature has accumulated on 'pressure solution' since 1849 when Thomson formulated, and later experimentally confirmed (1861), the principles of solution of crystal aggregates under differential stresses. Sorby (1863) was the first to recognize and describe the significance of pressure-solution phenomena in rocks and his work was followed by many reports of various pressure-solution structures in sedimentary rocks which have undergone low-grade metamorphism (e.g. Borradaile *et al.* 1982). Pressure-solution structures have been recognized only rarely in igneous rocks, for example pegmatites (Bailly 1954) and volcanites (Goldring & Conolly 1962). An historical review and synthesis of research on pressure solution has been given by Kerrich (1977). The purpose of this paper is to describe an example of pressure-solution structures in a granite, a type of rock from which this deformation mechanism has hitherto been poorly documented.

GEOLOGICAL SETTING

The Corcoesto granite is a syntectonic alkaline pluton which crops out in Galicia, northwest Spain (Fig. 1), where the regional structure of the arcuate Variscan Belt has developed from the superposition of two major phases of folding, both associated with cleavage formation (Matte 1968). In the autochthonous rocks, recumbent folds of approximately km-scale amplitude, overturned E and SE towards the interior of the arc, are attributable to the earlier phase. These early folds are associated with E-thrusting of allochthonous high-grade rocks (Bayer & Matte 1979, Iglesias *et al.* 1983). The second phase is represented by upright tight folds whose axial planes may dip steeply W (i.e. towards the exterior of the arc). They deform both the recumbent folds and the thrust contacts. The emplacement of the syntectonic calc-alkaline and alkaline granites is related to this structural sequence with the calc-alkaline granitoids emplaced prior to the second phase of folding. The later alkaline granites are generally considered to be syn- to post-tectonic with respect to the second phase of folding (Capdevila & Floor 1970, den Tex 1977). The Corcoesto granite is one of these later alkaline granites which have been affected by Late-Variscan transcurrent ductile-shear zones (Iglesias & Choukroune 1980) resulting from an approximately E–W compression (Iglesias & Ribeiro 1981). In Corcoesto, arsenopyrite and gold are concentrated within roughly E–W quartz veins, the tectonic setting of which is dealt with in the present study.

DEFORMATION OF THE CORCOESTO GRANITE

Description of the rock—regional observations

The Corcoesto granite is a medium-grained and leucocratic two-mica granite composed of 35–40%

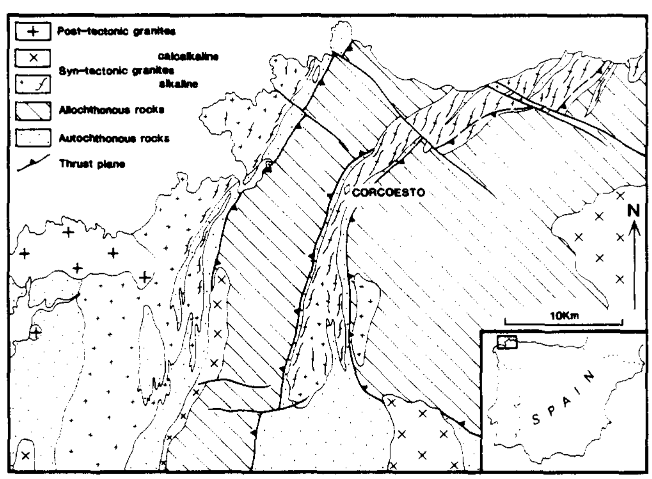

Fig. 1. Location of the Corcoesto Mine on a geological sketch-map of northwest Spain. Crosses in alkaline syntectonic granites indicate undeformed rock. S-shaped symbols follow the foliation trend in deformed bands.

quartz, 15–30% plagioclase (An_{10-20}), alkali feldspar and minor amounts of muscovite and biotite. Accessory minerals include zircon, apatite and arsenopyrite. The deformed granite is mainly located at the borders of the pluton (Fig. 1) in zones varying in width from a kilometre to a metre. From the inner undeformed granite to the country rocks, a subvertical foliation at an angle to the pluton margin becomes progressively closer to parallel to the pluton margin. This corresponds to half of the continuous shear-zone model described by Ramsay & Graham (1970). Locally, within the pluton, discrete conjugate shear zones occur.

Mesoscopic structures—microtectonic analysis

The type of deformation in the granite is comparable to that described from deformed granites associated with the South Armorican Shear Zone in France (Berthé *et al.* 1979); the subvertical foliation (*S* planes) is defined by the average flattening plane of the grains and the preferred orientation of mica grains. The second set of irregularly spaced (10–0.5 cm) shear planes (*C* planes) is also subvertical. The angle between the two sets of planes varies (5–35°) diminishing in the most deformed rocks with a concomitant size reduction of the constituent grains. Stretching lineations marked by elongate feldspar aggregates and long axes of quartz lenses on *S* planes, and slickenside striations on *C* planes, are close to the horizontal, indicating, along with the subvertical *S* and *C* planes, that the deformation areas are transcurrent ductile-shear zones. The intersection between subvertical *S* and *C* planes implies dextral shear (Berthé *et al.* 1979, White *et al.* 1980). Towards the pluton margins, the inferred movement along *C* planes is compatible with the general sense of shear indicated by the overall foliation trajectory patterns (NE–SW dextral shear zones, Figs. 1 and 2) and emphasizes the close association of *S*- and *C*-plane development during progressive deformation. Minor NNW–SSE sinistral shear zones (Fig. 2) are everywhere associated with the major dextral shear zones. From these two conjugate directions, a shortening along approximately 105–285° (st in Fig. 2) can be inferred, which bisects the obtuse angle between the shears as is commonly the case for ductile deformation (Ramsay 1980).

Veining comprised an important element of the deformation and the gold-bearing quartz veins of the Corcoesto mine appear to contain the shortening direction (Fig. 2). They are analogous to the numerous, discrete quartz veins (1–300 mm thick) which can be observed throughout the deformed granite (Fig. 3), generally in a single direction (Fig. 4) oblique to the foliation and to the *C* planes (Figs. 2 and 3). The intersection of the quartz veins with the foliation is almost orthogonal to the stretching lineation (a mean of 88° with a range of 72–119°, Fig. 5). In sections orthogonal to *S* and parallel to the stretching lineation, offsets of the quartz veins on the scale of millimetres along *C* planes have opposite movement senses from the general sense of shear (Figs. 3 and 5).

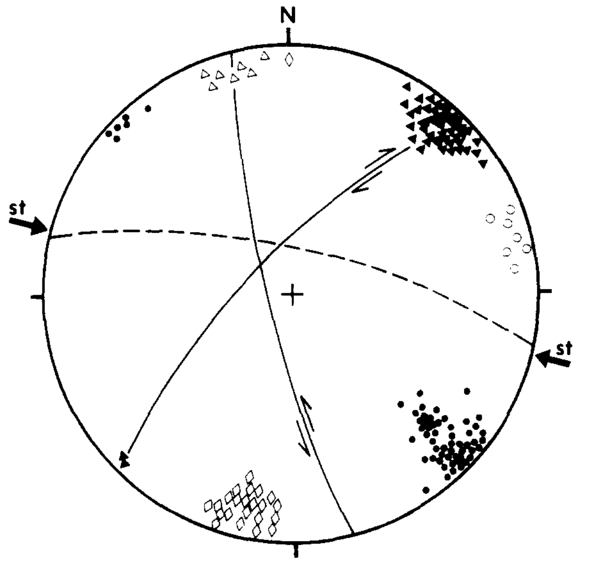

Fig. 2. Equal-area stereographic projection (lower hemisphere) of 59 poles to dextral *C* planes (closed circles) and their slickenside striations (closed triangles), and seven poles to sinistral *C* planes (open circles) and their slickenside striations (open triangles). Diamonds represent 23 poles to the gold-bearing quartz veins at Corcoesto, the cyclographic trace of the mean-vein plane being the dashed circle. st, shortening direction.

Microscopic structures

Microstructures associated with the foliation indicate plastic deformation for micas (undulose extinction, kinks, (001) translation) and quartz (undulose extinction, subgrains, deformation lamellae) whose final grain-size is 100–200 μm. However, individual feldspar grains (up to 1 mm in size) commonly show evidence for both brittle deformation (fracturing with rotation and displacement of component fragments) and plastic flow (deformation bands and deformation-twin lamellae). There is significant grain-size reduction of quartz (20–30 μm), mica (10–100 μm) and feldspar (20–200 μm) along the *C* planes, which are interpreted as resulting from inhomogeneous deformation. Magmatic biotite and muscovite appear to have remained stable during deformation. Accordingly, we think that *S* planes formed early in the history of progressive deformation. The stable mineral assemblage enables us to estimate the metamorphic conditions as medium to low grade ($T \simeq 400°C$ in Winkler 1974). Phyllosilicates which are too

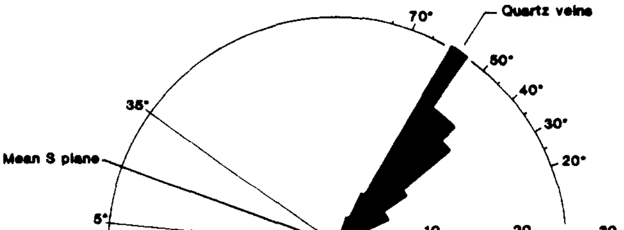

Fig. 4. Histogram of measured angles between quartz veins (106 measurements), *C* planes and *S* planes. 5–35° shows the range of measured angles between *S* and *C* surfaces. Heavy line is mean *S* to *C* angle (20°).

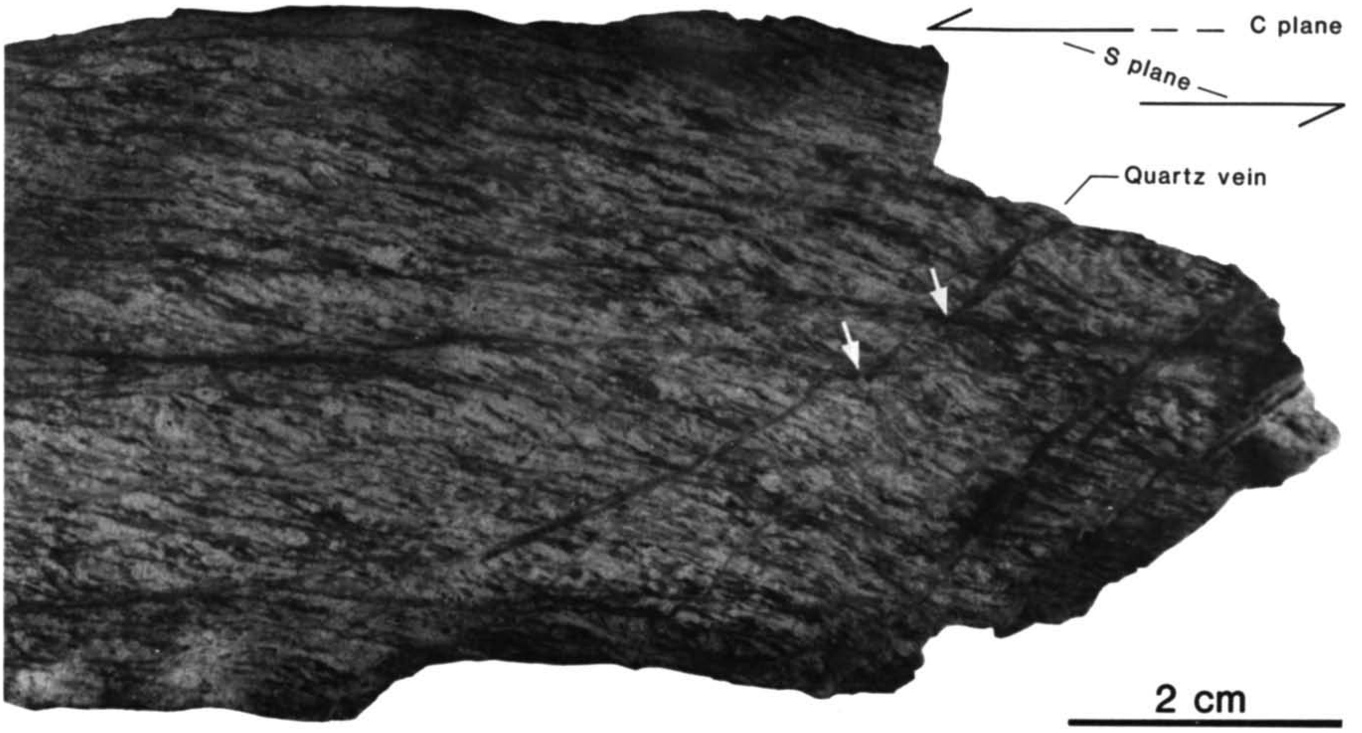

Fig. 3. Typical sample from the Corcoesto Granite. *S*, foliation plane; *C*, shear planes; quartz veins also indicated. *C–S* plane relations show a general sinistral sense of shear. Arrows indicate dextral offset of quartz veins along early *C* planes.

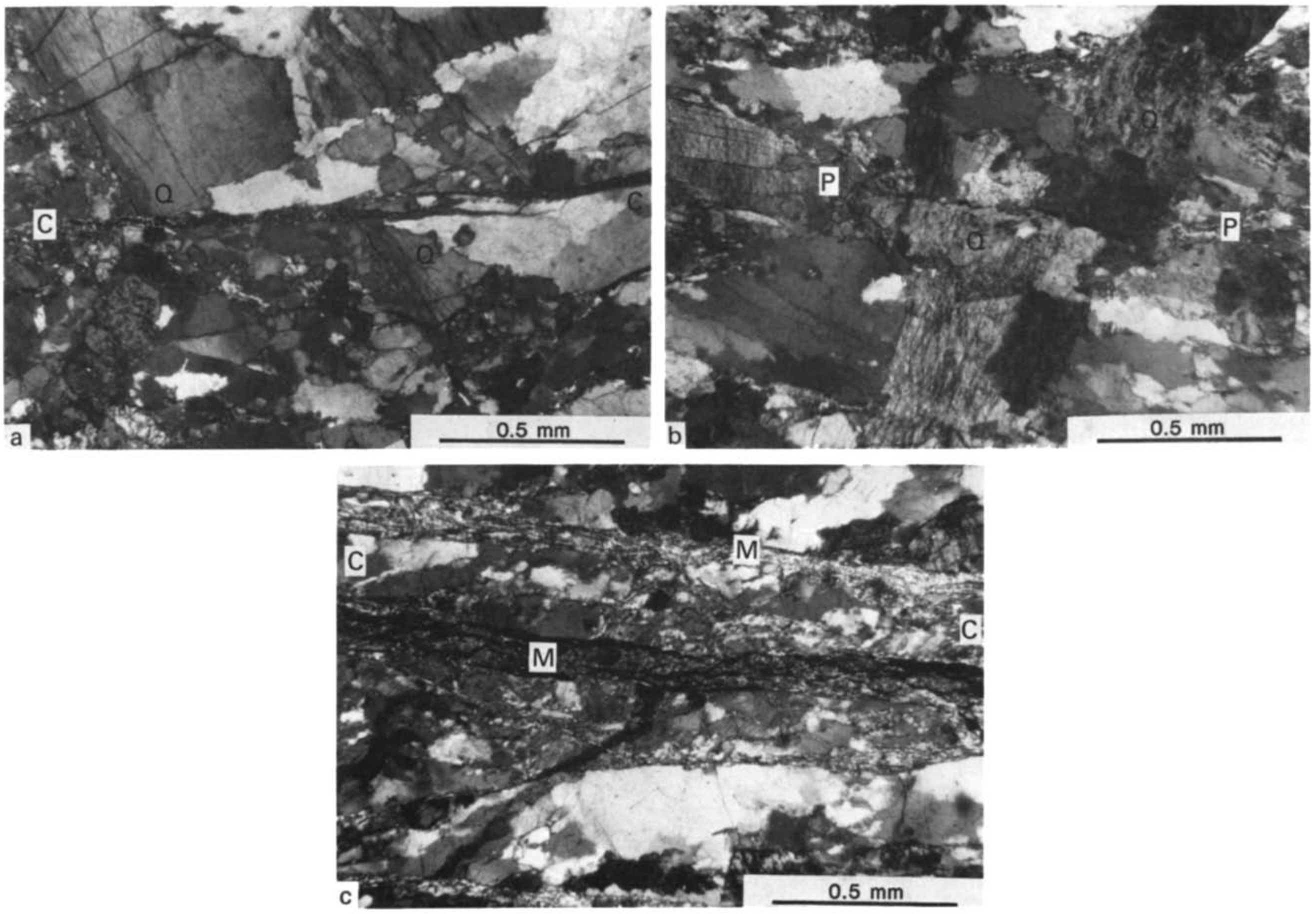

Fig. 6. (a) Photomicrograph of the offset of a quartz vein (Q) along a *C* plane. (b) Photomicrograph of a serrated plane (P–P) containing small arsenopyrite grains along an offset quartz-vein (Q) at the end of a *C* plane. (c) Photomicrograph showing concentration of phyllosilicates (M) and iron oxides along a *C* plane.

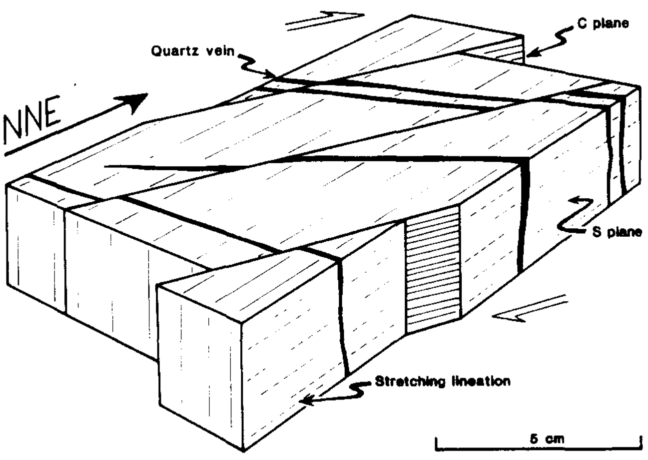

Fig. 5. Block diagram illustrating the attitude of various microstructures in the Corcoesto granite. See text for details.

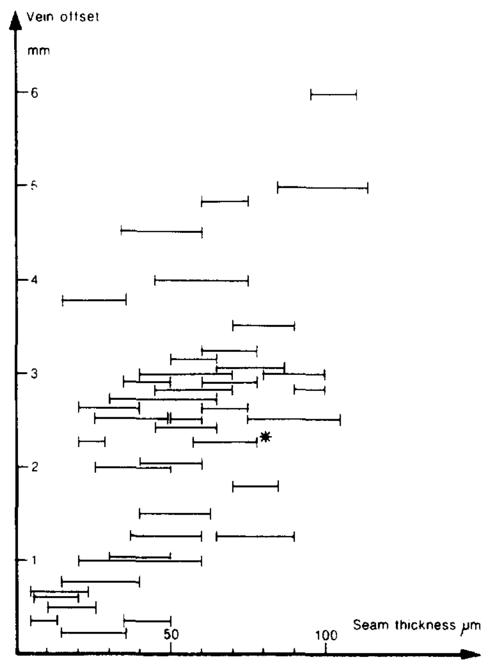

Fig. 7. Plot of 41 seam thicknesses against vein offsets (* indicates that two measurements are identical).

small to identify optically occur in *C* planes. The occurrence of these phyllosilicates suggests similar or lower-grade conditions of metamorphism during formation of the *C* planes.

Veins are commonly filled with quartz which is coarser grained than that observed in the host rock (200–500 μm) (Figs. 6a & b). The quartz displays no significant intracrystalline deformation apart from minor undulose extinction and coarse subgrain development. Quartz veins have sutured and irregular offsets but do not show evidence of localized fracturing and intracrystalline deformation (Figs. 6a & b).

C planes with the deformation features described above at their ends are transitional into zones with a distinct mineralogy. These zones are up to 100 μm in width and show a complete deficiency of quartz and concentration of phyllosilicates and iron oxides (Fig. 6c). Phyllosilicates show no observable shape changes, nor increase in intracrystalline deformation or (001) translation. However, they are more closely packed and have a strong preferred orientation of basal planes parallel to the zone boundaries.

INTERPRETATION

Quartz veins which contain the shortening direction indicated by conjugate *C* planes (Fig. 2) trend perpendicular to the stretching lineation and are oriented at a large angle to the foliation. These veins are interpreted to have initiated as tensile failures normal to the direction of extension, that is, normal to the direction of minimum principal stress responsible for plastic flow. It is well known that tensile failure can occur at depths under conditions of high fluid pressure (Paterson 1978, Nicolas & Jackson 1982). As these quartz veins cross-cut the foliation and some shear planes without any deformation, their development followed the initiation and propagation of both the *S* and *C* planes. Offsets of quartz veins along some *C* planes could be interpreted as implying later slip along strain-induced features. However, several features suggest an absence of such late brittle deformation. (1) The *C* planes show only one direction of slickenside striations parallel to the direction of bulk flow indicated by the stretching lineation. The sense of bulk flow indicated by the $C \wedge S$ relationship is inconsistent with the sense of offset observed for the quartz veins. Finely crushed gouge material is absent, and there is a paucity of fractured grains adjacent to these planes. (2) Micas within *C* planes are relatively undeformed and show no kink or (001) translation which would indicate that movement had taken place along these planes. (3) The locally irregular serrated nature of the planes precludes the possibility of slip (Fig. 6b).

We suggest that the observed quartz-vein offsets are caused by local volume reduction along *C* planes resulting from dissolution and removal of the more soluble minerals, particularly quartz and sulphides. This interpretation explains the increase in phyllosilicate concentrations along these planes. It is also consistent with the apparently proportional relationship between the thickness of the residual seams and the observed offset (Fig. 7) which can be related to the volume of rock removed by dissolution processes. Estimates of the volume of material removed, obtained by projecting missing vein segments, range between 5 and 13%. We believe that the quartz veins and the selective mass transfer implied by this interpretation are penecontemporaneous because they belong to the same regional shear deformation.

CONCLUSION

The lack of cataclastic textures, the mineralogical differentiation along *C* planes and truncation of quartz veins in the Corcoesto granite indicate a regime in which deformation was dominated by pressure solution—at least in the later stages of deformation. Removal of soluble minerals along earlier discontinuities such as *C* planes, and rarely *S* planes, suggests that the solution transfer was preferentially located along zones dominated by small grains, an observation recorded in a number of quartz mylonites (White & Wilson 1978, Kerrich *et al.* 1980). Gold and the associated arsenopyrite in the quartz veins were probably deposited during this late stage of deformation related, in Galicia, to the operation of major transcurrent ductile shear-zones which affected syntectonic granites during a Late Variscan event. Pressure solution then enabled redistribution of gold and sulphides within the granite along *C* planes.

We emphasize that pressure-solution structures may be more common in granites deformed in low-grade metamorphic conditions than would appear from the literature. Without any marker, which in this case is provided by the quartz veins, such processes are difficult to detect in isotropic material such as granite. Consequently the determination of finite strain from the angle $S \wedge C$ planes is hazardous.

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REFERENCES

Bailly, P. A. 1954. Présence de microstylolites dans des pegmatites et des lentilles de quartz. *Bull. Soc. geol. Fr.* **3**, 299–301.

Bayer, R. & Matte, P. 1979. Is the mafic/ultramafic massif of Cabo Ortegal (northwest Spain) a nappe emplaced during a Variscan obduction? A new gravity interpretation. *Tectonophysics* **57**, T9–T18.

Berthé, D., Choukroune, P. & Jégouzo, P. 1979. Orthogneiss, mylonite and non-coaxial deformation of granites: the example of the South Armorican Shear Zone. *J. Struct. Geol.* **1**, 31–42,

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

Capdevila, R. & Floor, P. 1970. Les différents types de granites hercyniens et leur distribution dans le Nord Ouest de l'Espagne. *Bol. Geol. Min. Inst. Geol. Min. España*, **81**, 215–225.

Goldring, H. G. & Conolly, J. R. 1962. Stylolites in volcanic rocks. *J. sedim. Petrol.* **32**, 534–538.

Iglesias Ponce de Leon, M. & Choukroune, P. 1980. Shear zones in the Iberian Arc. *J. Struct. Geol.* **2**, 63–68.

Iglesias, M. & Ribeiro, A. 1981. Zones de cisaillement ductile dans l'arc Ibero-Armoricain. *Comm. Serv. geol. Portugal* **67**, 85–87.

Iglesias, M., Ribeiro, M. L. & Ribeiro, A. 1983. La interpretacion aloctonista de la estructura del noroeste Peninsular. In: *Geologia de España*. Int. Geol. Min. España, Madrid, 459–467.

Kerrich, R. 1977. An historical review and synthesis of research on pressure solution. *Zentbl. Miner. Geol. Paläont.* **1**, 512–550.

Kerrich, R., Allison, I., Barnett, R. L., Moss, S. & Starkey, J. 1980. Microstructural and chemical transformations accompanying deformation of granite in a shear zone at Mieville, Switzerland; with implications for stress corrosion cracking and superplastic flow. *Contr. Miner. Petrol.* **73**, 221–242.

Matte, P. 1968. La structure de la virgation hercynienne de Galice (Espagne). *Géol. Alpine* **44**, 1–128.

Nicolas, A. & Jackson, M. 1982. High temperature dikes in peridotites: origin by hydraulic fracturing. *J. Petrology* **23**, 568–582.

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

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

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

Sorby, H. C. 1863. Über kalkstein-geschiebe mit eindrücken. *Neues Jb. Miner. Geol. Paläont.* **34**, 801–807.

den Tex, E. 1977. Le socle poly-cyclique et son rôle dans l'évolution de la chaine varisque en Galice (Espagne du NW). In: *La Chaine Varisque d'Europe Moyenne et Occidentale. CNRS*, Rennes **243**, 441–451.

Thomson, J. 1849. Theoretical considerations on the effect of pressure in lowering the freezing point of water. *Trans. R. Soc. Edinb.* **16**, 575–580.

Thomson, J. 1861. On crystallization and liquefaction, as influenced by stresses tending to change of form in the crystals. *Proc. R. Soc.* **11**, 473–481.

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

White, S. H. & Wilson, C. J. L. 1978. Microstructure of some quartz pressure fringes. *Neues Jb. Miner. Abh.* **134**, 33–51.

Winkler, H. G. F. 1974. *Petrogenesis of Metamorphic Rocks*. Springer, Berlin.