Journal of Structural Geology, Vol. 13, No. 5, pp. 539 to 555, 1991
Printed in Great Britain

0191-8141/91 $03.00+0.00

Progressive folding and foliation development in a sheared, coticule-bearing phyllite

C. K. Mawer* and P. F. Williams

Department of Geology, University of New Brunswick, Fredericton, NB, Canada E3B 5A3

(*Received* 14 *September* 1989; *accepted in revised form* 22 *October* 1990)

Abstract—An evolutionary sequence of fold and foliation development interpreted to be the result of a single progressive deformation is seen in exposures of strongly sheared phyllite and coticule, 10 km south of the E–W-trending Chedabucto Fault in the Meguma Terrane of Nova Scotia (Canada). Transposition *sensu* Sander at mesoscopic and microscopic scales is fundamental to this development, as are fold hinge rotation and migration. The transposition process is cyclic, with folds of layering and foliation developing, rotating, and being refolded. Fold hinges and axial surfaces repeatedly tend towards stable end-orientations, but do not reach them due to the cyclic transposition process. The foliation is a steady-state feature, continually reconstituted by microscale transposition and dynamic recrystallization. Several important consequences of these contemporaneous processes are: (1) geometrical relationships between folds and foliation are variable in the same outcrop, even though both developed synchronously, and many folds have the geometrical relationships of cleavage-transected folds; (2) the foliation is transient in orientation and therefore cannot track the XY principal plane of total finite strain; and (3) fold hinge migration may be a common process, especially in deformations involving a large component of simple shearing. The observed relationships may be applicable to study of fold and foliation development in orogenic belts in general.

INTRODUCTION

The development of foliation, its relationship to associated folds, and the relationship of both to principal finite strain directions is a complex structural topic. Many folded rocks possess a single tectonic foliation, which, although commonly fanning around folds, generally bears a close geometric relationship to fold axial surfaces in hinge areas and is consequently known as an axial surface foliation. This relationship generally tells little about the deformation history or metamorphic history of either folds or foliation. How and when did this foliation initiate, and by which processes did it evolve? Is its 'axial surface' orientation genetic or fortuitous? What are the geometrical and temporal relationships of microscopic foliation-defining elements to the outcrop-scale foliation? Did the folds nucleate and develop in more-or-less stable orientations, or were they subject to phenomena such as hinge rotation and migration, and variations in orientation of their present 'axial surface' foliation with respect to the geometrical axial surface during fold evolution? These, and similar, questions are commonly difficult to answer in a definitive way. Many rocks have had a complex history, with processes such as transposition, deformation partioning, dynamic recrystallization, prograde metamorphism and growth of new, syntectonic mineral phases combining to obliterate evidence of earlier parts of that history.

Exposures of strongly-sheared phyllite in eastern Nova Scotia, Canada, show foliation and folds which we interpret to have developed during a single, progressive deformation. The folds are defined by thin coticule layers (original bedding) embedded in the phyllite. [Coticules are spessartine garnet-quartzites, which occur as thin laterally-persistent layers and are generally considered to be meta-tuffs; see Kramm (1976), Lamens *et al.* (1986). They are particularly useful marker horizons and younging and facing indicators, since they generally occur in shales, slates or phyllites which tend to lack such features, are of geologically instantaneous origin and thus parallel to the original accumulation surface, and are commonly graded.] Folds of the coticule layers show a direct relationship between tightness and hinge orientation, with more open folds showing steeper plunges. The folds are refolded in places. The foliation discussed occurs in the phyllite. Outcrop-scale foliation is commonly oblique to axial surfaces of tight folds, and microscopic observation of the sheared phyllite shows that foliation-defining elements are commonly oblique to both fold axial surfaces and outcrop-scale foliation. Such folds have the geometrical relationships of cleavage-transected folds (Powell 1974, Borradaile 1978). Some folds in the same exposures, however, have a foliation statistically parallel to their axial surface, while others fold this foliation. Thus, a system of considerable complexity exists in these apparently simply deformed rocks, and the relationship of the various structural elements illustrates well some of the complications that can accumulate during a progressive deformation, especially one with a large component of simple shearing. We believe the results of this study may have general applicability in orogenic belts, as tectonic straining commonly involves an important component of simple shearing.

*Present address: Proterozoic Research Group, Carpentaria Exploration Company, G.P.O. Box 1042, Brisbane, Queensland 4001, Australia.

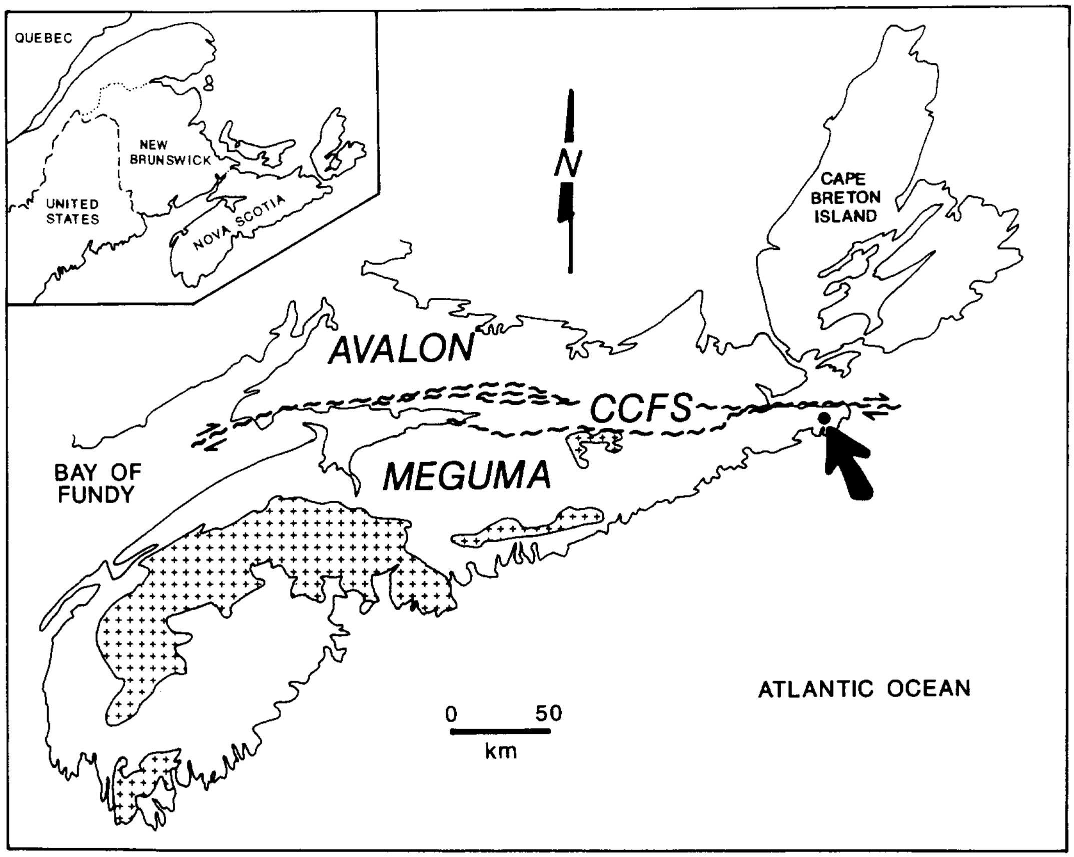

Fig. 1. Map showing the Meguma Terrane of Nova Scotia. CCFS is the Cobequid–Chedabucto fault system, which forms the northern boundary of the Meguma Terrane. Study locality shown by small black dot (arrowed) at the eastern end of the Meguma Terrane.

REGIONAL SETTING

The Meguma Terrane is a distinct lithotectonic entity in the North American Appalachian Orogen. It is interpreted (Schenk 1971, 1978) as African in origin, accreted to the more northerly Avalon Terrane prior to the formation of the present Atlantic Ocean. The Meguma Terrane is the furthest outboard Appalachian 'suspect terrane' (Williams & Hatcher 1982), and is separated from the Avalon Terrane by the Cobequid–Chedabucto fault system (Webb 1969) (Figs. 1 and 2). This fault zone is a major, steeply dipping to vertical, dextral transcurrent zone of displacement, lying parallel to and overprinting an earlier mylonitic shear zone with the same displacement sense (Mawer & Williams 1986, Mawer & White 1987).

The Meguma Terrane is largely composed of rocks of the Meguma Group, a thick Cambro-Ordovician sequence consisting of two formations, the underlying sandstone-rich Goldenville Formation, and the overlying slate/phyllite-rich Halifax Formation (Schenk 1976, 1978). The present study considers phyllites and associated minor coticule layers of the Halifax Formation (Fig. 2).

Most deformation in the Meguma Terrane occurred during the Devonian Acadian Orogeny, as did regional low- to medium-grade metamorphism and emplacement of numerous granitoid intrusions (Fyson 1966, Eisbacher 1970, Reynolds *et al.* 1973, Clarke & Halliday 1980, 1985). During the Carboniferous Hercynian Orogeny, significant deformation occurred along the Avalon–Meguma boundary zone (e.g. Williams & Hatcher 1982) reactivating the terrane-bounding mylonitic shear zone and then forming the Cobequid–Chedabucto fault system (Mawer & White 1987).

The phyllite and coticules described below are exposed approximately 10 km south of the Chedabucto Fault (Figs. 1 and 2), and lie within the sheared boundary zone. The area is approximately 0.2 km^2 and is centered at 45°16′N, 61°11′W (Fig. 2). The coticules are strongly folded. The dominant foliation visible to the naked eye in these outcrops is, at least in part, a fine-scale compositional layering. It has a statistically-defined dip of 88° towards 348°, so that it strikes at a small acute angle to the boundary of the major fault zone. This value was determined from 53 measurements of foliation distant from either folds of the coticules or intrafolial folds of the phyllite, and represents a very tight concentration of over 50% per 1% area on a contoured pole figure. A mineral lineation lying in the foliation varies from shallowly E-plunging (approximately 20°) to horizontal. This lineation is an extension lineation, defined by elongate quartz grains and trains of opaque grains and small garnet porphyroblasts. The lineation is parallel to the extension lineation developed in sheared granitoids in the area (Fig. 2) (Mawer & Williams 1986).

FOLDS IN THE COTICULE LAYERS

The coticule layers in the studied exposures are generally 0.5–1.5 cm thick though locally they have been tectonically thinned to zero. They are invariably strongly folded (Fig. 3a) and commonly have been boudinaged prior to or at some stage during the folding. The coticule layers are more resistant to erosion than the enclosing phyllite, so that minor folds protrude several centimetres above the exposure surfaces. This permits accurate measurement of interlimb angles and orientations of axial surfaces and hinge lines.

The folds of the coticule layers are non-cylindrical but, in the exposures described here, are not sheath folds (e.g. Williams & Zwart 1977, Quinquis *et al.* 1978); elsewhere in the shear zone, however, there are sheath folds within phyllite with axes that are shallowly to horizontally plunging. Throughout the shear zone the coticule folds have Z-asymmetry when viewed on horizontal exposure surfaces, but can have S- or Z-asymmetry on vertical planes even though viewed in a constant direction. Along the fold hinges, plunges vary and profiles change, in some cases abruptly. The folds are complex, nested and en échelon.

Figures 5–10 summarize orientational and geometrical data for 67 minor folds in the coticules, and are discussed below. All orientation diagrams are plotted as lower-hemisphere, equal-area projections.

Hinge lines

Hinge line orientations (Fig. 5) define a partial girdle close to the dominant foliation orientation measured in exposure, but about 6° anticlockwise of its trend. The angular spread of their distribution narrows as the hinge line orientations vary from vertical to horizontal, and falls almost totally in the eastern half of the plot. A diagram of hinge-line plunge vs interlimb angle (Fig. 6) shows a continuous progressive trend from open, steeply-plunging folds to tight-to-isoclinal, non-plunging folds. We consider this trend to be due to hinge rotation and fold tightening during progressive deformation, as discussed below. Figure 7 shows fold interlimb angle vs trend of fold hinge lines. Note the funnelling effect, whereby more-open folds have widely divergent hinge trends, and as folds become tighter their variation in trend decreases rapidly, becoming concentrated close to the trend of the dominant outcrop-scale foliation.

Axial surfaces

Poles to axial surfaces (Fig. 8) are distributed in a broad girdle that dips shallowly towards the west. There is a concentration of poles close to the pole of the obvious outcrop-scale foliation, but this concentration is displaced about 10° anticlockwise with respect to this

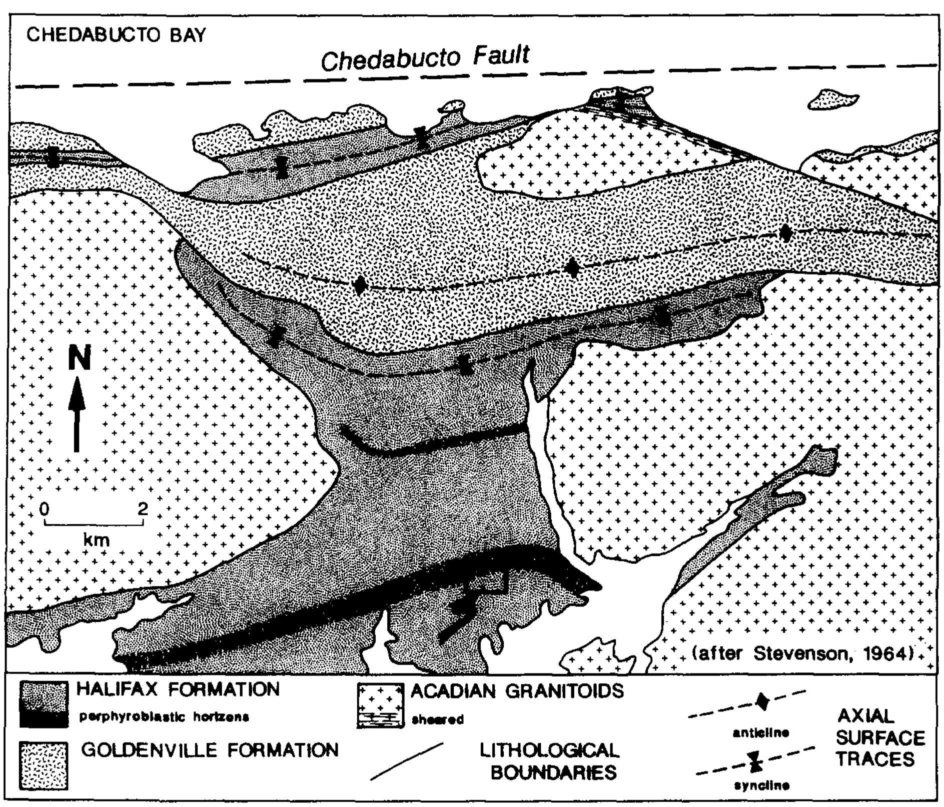

Fig. 2. Geological map of area around studied exposures, modified after Stevenson (1964). Exposures are in boxed area, arrowed. Note shear-band-like geometry of metasedimentary–granite package just south of Chedabucto Fault, indicative of dextral shearing.

pole. A diagram of axial surface dip vs fold interlimb angle (Fig. 9) can be interpreted as showing a progressive trend from open folds with moderately to steeply dipping axial surfaces, to tight to isoclinal folds with steeply to vertically dipping axial surfaces. Figure 10, axial surface dip direction vs fold interlimb angle, shows a funnelling similar to that of Fig. 7, indicating that folds become tighter as their axial surface dip direction tends towards parallelism with that of the outcrop-scale foliation.

Packets of isoclinal, horizontally plunging folds are locally refolded (Fig. 3b). The overprinting folds are open, steeply plunging, and have axial surfaces which trend at a high angle to layering and foliation.

DISCUSSION OF FOLD DATA

The trends represented by the above figures show a clear, evolutionary sequence in the development of the coticule folds, interpreted as follows. Initially folds nucleated and grew with essentially vertical hinges and axial surfaces trending approximately perpendicular to the shear zone movement vector; these folds are true drag folds (see Lister & Williams 1983). During progressive dextral simple shearing, the fold hinge lines and axial surfaces rotated so that the hinge lines tend towards parallelism with the bulk shear direction (approximately horizontal and E–W), and the axial surfaces tend towards parallelism with the bulk shear plane (approximately vertical and E–W). Synchronously, interlimb angles decreased as folds tightened from open to isoclinal, and the folds became markedly asymmetric (cf. Sanderson 1979). Further, this sequence is cyclic, as shown by the packets of isoclinal and horizontally plunging folds that have undergone another folding 'episode', and are overprinted by open, steeply plunging folds (Fig. 3b). We interpret this as a progressive sequence, rather than a series of discrete, temporally separated events.

In summary, folds and foliations are progressively transposed towards the more stable orientation of the shear zone and the resulting transposition folds and foliation can be perturbed at any stage during deformation to begin a new cycle of folding and transposition (see Williams & Zwart 1977). During shearing, the transposition foliation and associated folds thus constitute a dynamic but steady-state system, and it is for this reason that folds can be observed in various stages of development and transposition. One transposition sequence is represented diagrammatically in Fig. 11.

Which way a fold hinge will rotate within the shear plane will depend on at least two factors. If we assume that the deformation is a perfect and homogeneous simple shearing, then the fold hinge will not rotate so long as it lies in the shear plane. However, any deviation in the fold hinge orientation from the orientation of the shear plane will result in hinge rotation in response to the simple shearing (see Hobbs *et al.* 1976, p. 287). Thus segments of the hinge lines of drag folds that are doubly plunging relative to the bulk shear plane will rotate in opposite directions, even during homogeneous simple shearing, to produce sheath folds (Williams & Zwart 1977, Ramsay 1980). Different senses of rotation will also occur if the deformation is heterogeneous such that the shear vector varies in length within any given shear plane (Skjernaa 1985). If the shear is dextral throughout the zone all folds are expected to develop with initial Z-asymmetry, as viewed down-plunge. However, the asymmetry shown on a given vertical plane will depend on which way they have rotated (Fig. 12). The fact that we observe both asymmetries in vertical sections in the study area indicates that the folds did rotate in both directions, either because of plunge variations or because of heterogeneity of shearing. However, the lack of sheath folds in these outcrops indicates that variation in rotation sense was on a scale larger than that of the observed folds.

As the fold hinge lines rotated towards horizontal, it seems probable that, except in the unlikely event that the coticules behave as purely passive layers, hinges must have migrated through the layering, in a similar fashion to wave forms migrating through water. There is excellent evidence for this process (see 'Detailed Analysis of Horizontal Folds', p. 549 of this paper).

THE PHYLLITE

In outcrop, the phyllite shows a well-developed foliation, generally defined in part by a moderately to well-developed compositional layering (Fig. 4a). The presence of isolated spindle-shaped or ellipsoidal pods of compositionally identical phyllite, intrafolial to the dominant foliation and possessing intricately folded internal foliation and layering (Fig. 4a), shows that the obvious and apparently simple outcrop-scale foliation is the product of at least one cycle of near-complete transposition. Some of the transposed layering (e.g. the coticules) is of primary origin, but most compositional layering within the phyllite proper is of metamorphic origin.

A study has been made of very thin (5–7 μm) petrographic sections cut normal to the outcrop foliation and normal and parallel to the horizontal lineation.

Microstructures

In thin-section, the metamorphic compositional layering is defined by phyllosilicate-rich vs quartz-rich layers. These layers commonly show tight to isoclinal folds at this scale. Within the phyllosilicate-rich layers, and in some places within the quartz-rich layers, bands of chlorite–mica stacks (Voll 1960, Williams 1972) occur, concordant with the boundaries of the compositional layering. The metamorphic history recorded by the various assemblages is complex and is the subject of continuing study.

The phyllite is composed of more-or-less equal proportions of phyllosilicates (white mica and chlorite) and

Fig. 3. (a) Horizontal exposure of folded coticule layer in transposed phyllite. Fold enveloping surfaces trend roughly ENE–WSW, view looks NE. Scale bar is 10 cm long. (b) Refolding of isoclinal folds with horizontal hinges, plane of exposure horizontal. Scale bar is 10 cm long.

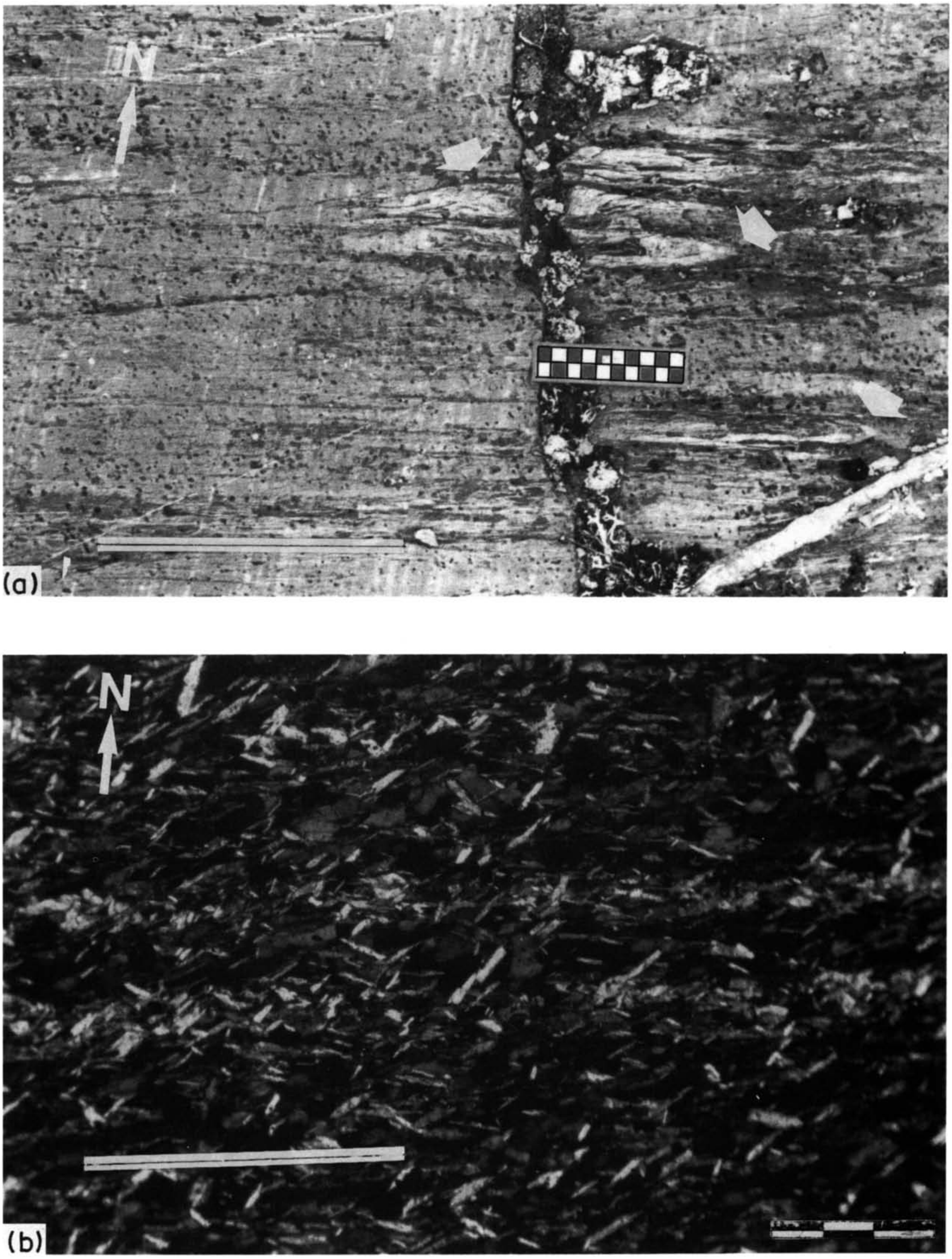

Fig. 4. (a) Intrafolial pods of intricately-folded phyllite in apparently simply-foliated phyllite matrix (arrowed), indicating intense transposition. Compositional layering parallel to white bar; white streaks trending NNE–SSW are glacial striations. Scale bar is 10 cm long. (b) Photomicrograph of typical phyllite, ultrathin section cut perpendicular to foliation and parallel to mineral extension lineation (i.e. essentially horizontal in geographical co-ordinates). Thin white line is parallel to the outcrop-scale foliation. Top of photograph towards north, crossed-polarized light, scale bar is 200 μm long.

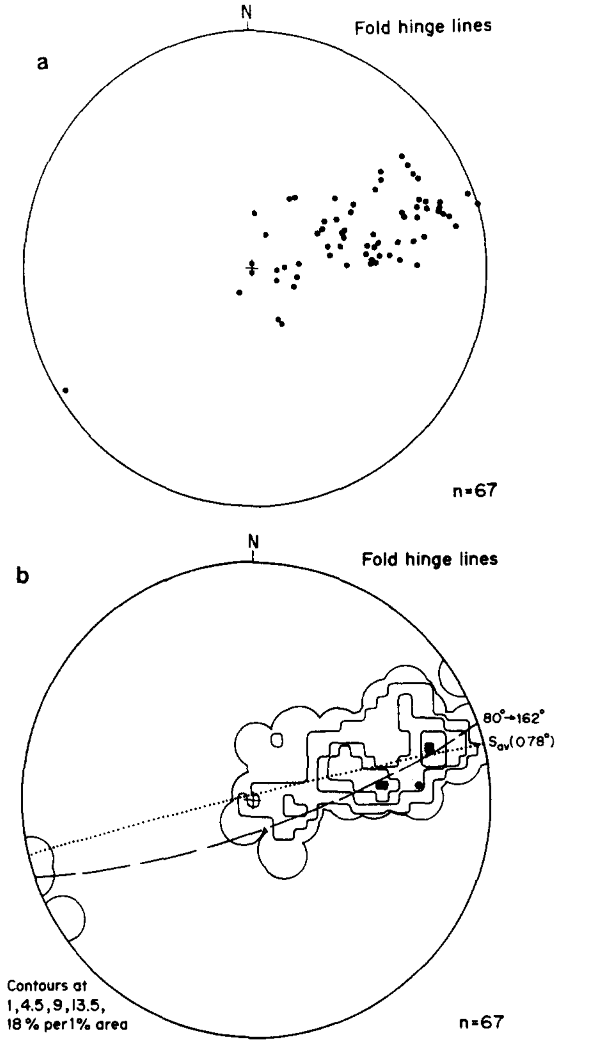

Fig. 5. Hinge-line orientation data for set of 67 minor folds of coticule layers, collected from large exposures within the boxed area of Fig. 2. (a) Lower-hemisphere equal-area projection with fold hinges plotted; (b) same data, contoured with respect to density. S_{av}, here and in following figures, is the average value for the outcrop-scale foliation (trends 078°). Dashed great circle is best-fit girdle to the data points.

quartz with generally about 5% of opaque minerals. Some layers show development of garnet (generally less than 10%), others contain cordierite porphyroblasts (less than 10%), and there are minor amounts of albite, tourmaline and zircon. Grain sizes are variable. Single phyllosilicate grains vary in length from the limit of optical detection (about 5 μm in this case) to over 100 μm but cluster between 20 and 60 μm (Fig. 13). The

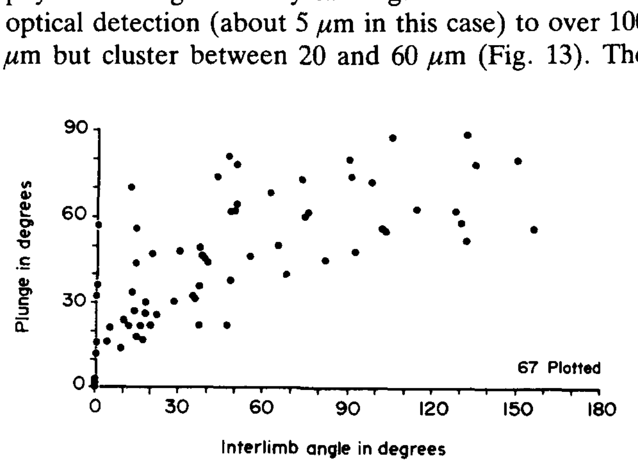

Fig. 6. Graph of hinge-line plunge vs fold interlimb angle.

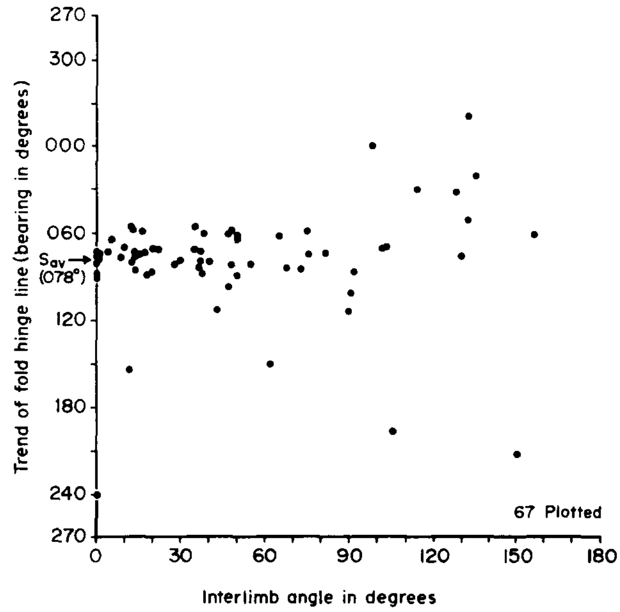

Fig. 7. Graph of hinge-line trend vs fold interlimb angle.

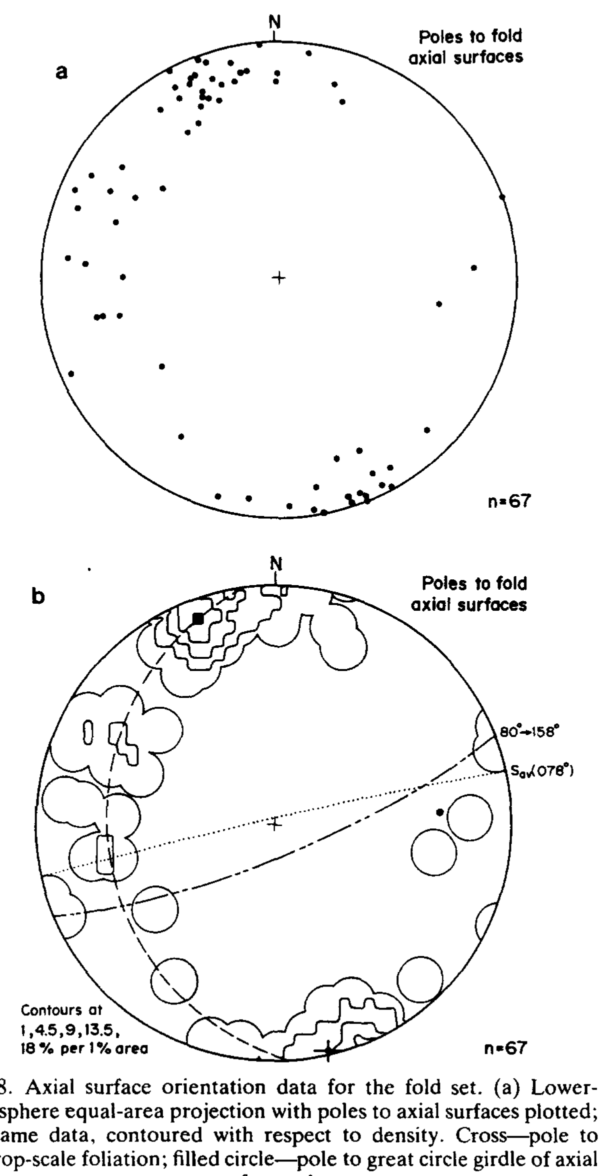

Fig. 8. Axial surface orientation data for the fold set. (a) Lower-hemisphere equal-area projection with poles to axial surfaces plotted; (b) same data, contoured with respect to density. Cross—pole to outcrop-scale foliation; filled circle—pole to great circle girdle of axial surface poles.

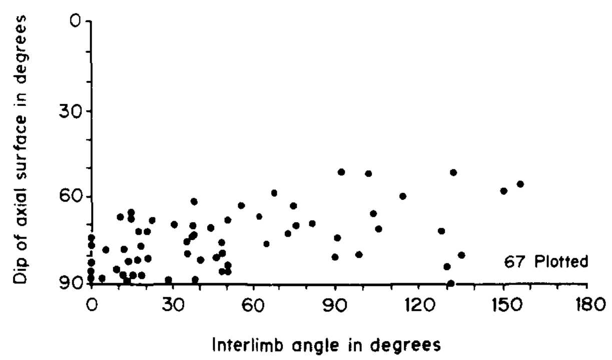

Fig. 9. Graph of axial surface dip vs fold interlimb angle.

chlorite–mica stacks show a fairly wide range in size but are generally 100–300 μm in length and 30–100 μm in width. They are elongated parallel to the foliation and commonly show kinking with serrated kink band boundaries parallel to the foliation. Locally, growth of new white mica has taken place along the kink band boundaries. Quartz grains vary in size but diameters cluster between 30 and 50 μm. The grains are commonly slightly elongate parallel to the foliation and in places show weak undulose extinction and locally poorly-developed subgrains. Opaque grains are of two varieties, a large, rounded, embayed type of similar size to the quartz grains and a smaller, elongate type. Garnets are euhedral to rounded and are of similar diameter to the quartz grains. In places the garnets are zoned with an inclusion-rich core and a clear rim. Cordierite porphyroblasts vary from about 1 to 5 mm in diameter, are equant to somewhat elongate, and are highly poikiloblastic with an internal foliation that is continuous with the external foliation, though not parallel to it. Porphyroblast rims are commonly pinitized and cores locally show sector twinning.

Texturally the phyllosilicate foliation varies from a penetrative, slaty type in the phyllosilicate-rich layers to a gradational, domainal type in the quartz-rich layers. The principal phyllosilicate orientation is not parallel to the compositional layering but lies at a small angle to it.

The presence of microscopic tight to isoclinal intrafolial folds of the metamorphic compositional layers is further evidence that the phyllite records at least one cycle of almost complete transposition of that layering. The fact that mica grains around these folds, with the exception of the chlorite–mica stacks, are generally straight and optically essentially strain-free (as are many of the slightly elongate quartz grains) indicates that syntectonic recrystallization has occurred. These mica grains are statistically axial planar to the microfolds. The cordierite porphyroblasts overgrew this latest transposed and recrystallized foliation syntectonically as shown by the fact that their internal foliation is continuous with the external foliation and they are rotated with respect to the latter.

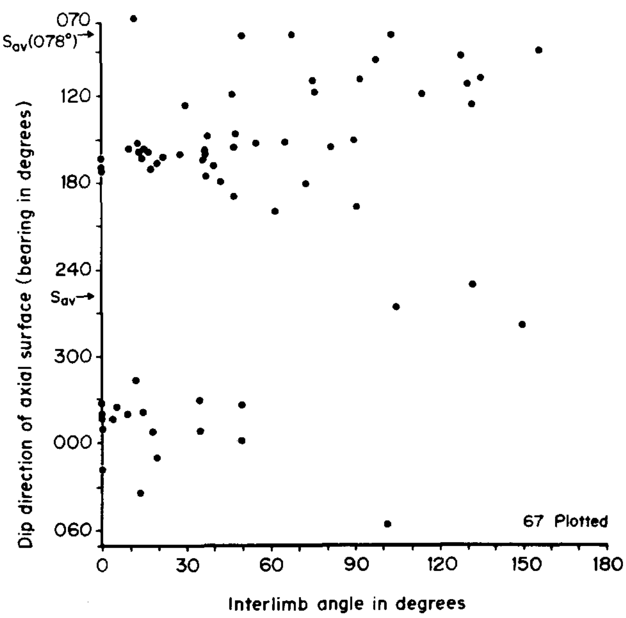

Fig. 10. Graph of axial surface dip direction vs fold interlimb angle.

Dimensional preferred orientation

Preferred orientations of long dimensions of mica and opaque grains were measured in thin sections cut parallel to lineation and perpendicular to foliation, for several specimens collected well away from coticule layers. All specimens gave essentially the same results and the results for one of them (shown in Fig. 4b) are presented and summarized below.

Mica. The azimuths of all white mica grains within a 200-grain area (not including chlorite–mica stacks) and their length/width (*L*/*W*) ratios were determined on a flat microscope stage. There is a strong orientation maximum (Figs. 14a and 16a) which is skewed, and two lesser maxima. All are anticlockwise of the outcrop-scale foliation trend, and decrease in intensity with increasing deviation from the foliation trend.

L/*W* ratios for white mica grains vary from about 1.5 to about 10 (Fig. 13). There is a definite tendency for

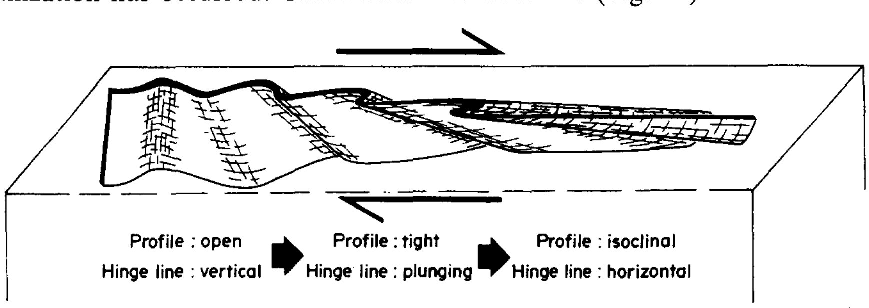

Fig. 11. Diagrammatic representation of one transposition sequence in a folded coticule layer; see text for discussion.

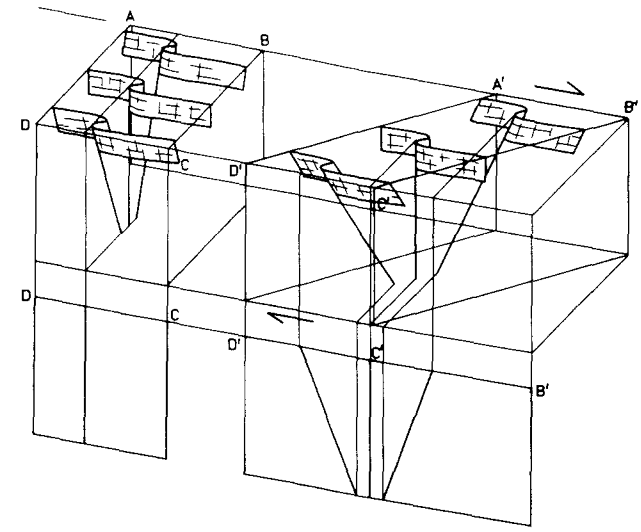

Fig. 12. Three-dimensional model showing how asymmetry of folds in vertical section depends on their pre-shearing plunge direction and hence sense of rotation during horizontal shearing; see text.

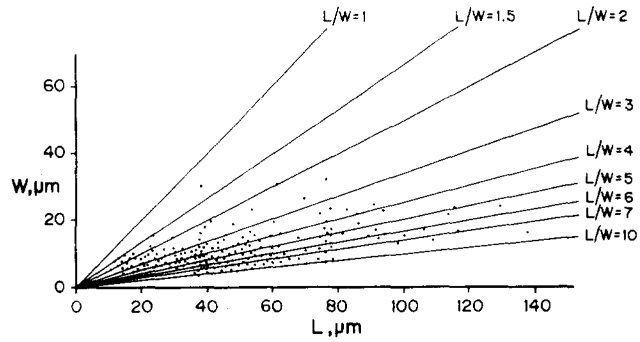

Fig. 13. Graph of length to width measurements of 200 individual matrix white mica grains, with superimposed contours of certain *L/W* ratios.

grains with low ratios to be smaller in size and therefore to cluster around the origin of the diagram (Fig. 13) more than the grains with high ratios which tend to be larger. Figure 15 shows the relationship between *L/W* ratios and azimuth and has six concentrations of more than 3% (see figure caption). Five of the concentrations correlate directly with the three major azimuth maxima described above.

The greatest azimuth concentration which falls about 10° anticlockwise of the outcrop-scale foliation trend (Figs. 14a and 16a) is seen to be defined by grains belonging to four *L/W* ratio groups (Fig. 15b, concentrations 1, 2, 4 and 6). The lesser azimuth concentrations, about 30° and 40° anticlockwise of the outcrop-scale foliation trend, are each defined by a single *L/W* ratio group (Fig. 15b, concentrations 3 and 5, respectively).

Elongate opaque grains. As with white mica grains the opaque minerals define a skewed maximum with its peak anticlockwise of the trend of outcrop-scale foliation and parallel to the mica peak (Figs. 14b and 16b). In contrast to the white mica distribution the opaque mineral concentration lacks subsidiary maxima. The two distribution patterns are superimposed in Fig. 16(c) to facilitate their comparison.

Rotated porphyroblasts

As noted above, the cordierite porphyroblasts are syntectonic. They overgrow the latest transposition foliation and are rotated relative to it, thus indicating the sense of shear parallel to the foliation.

In large sections through tightly folded coticule layers, cut parallel to the fold hinge lines and perpendicular to the axial surfaces (i.e. essentially horizontal), the porphyroblasts are rotated clockwise irrespective of position relative to the axial surfaces of the folds; that is, porphyroblast rotation (dextral) is the same for opposite limbs as well as for the fold closure. This holds for all cases studied to date. Further, it is noted that the foliation trace is not parallel to the traces of the geometrical axial surfaces, but is anticlockwise of them. This foliation is identical in morphology to that described from the phyllite specimens and it bears the same relationship to the geometrical axial surfaces as that foliation does to the outcrop-scale transposition. Consequently, tight folds are cut by the foliation with the same geometrical relationship as found in cleavage-transected folds.

The rotation of the cordierite porphyroblasts (and, as will be argued below, the dimensional preferred orientation pattern and textures of the phyllosilicates) indicates that the deformation had a component of shear-induced vorticity (Lister & Williams 1983), consistent with the shear zone environment. The shear couple operated horizontally, since the cordierite porphyroblasts show no rotation in sections perpendicular to the mineral lineation, nor in vertical sections parallel to the lineation. The sense of shear is clockwise (= dextral) and is not only constant on a scale greater than that of the folds, as demonstrated by the consistent sense of rotation of the porphyroblasts, but is apparently constant over a much greater area since the sense of rotation is the same for all specimens that we have examined (from the Chedabucto Fault to at least 15 km south and perpendicular to its strike).

DISCUSSION OF THE PHYLLITE

We interpret the textures and dimensional preferred orientation patterns reported above as being due to the concurrent operation of two processes, transposition (*sensu* Sander 1911) on a microscopic scale and dynamic recrystallization, which leads to the formation of new foliation elements. The process is summarized in Figs. 17 and 18 and is similar to the model proposed by Williams *et al.* (1977) for the development of axial-plane foliation. It differs in the cause of microfolding. The process here is one of mechanical rotation of phyllosilicates by drag folding, the bending and subsequent kinking of single phyllosilicate grains and aggregates of grains, tightening of kinks, and development of new

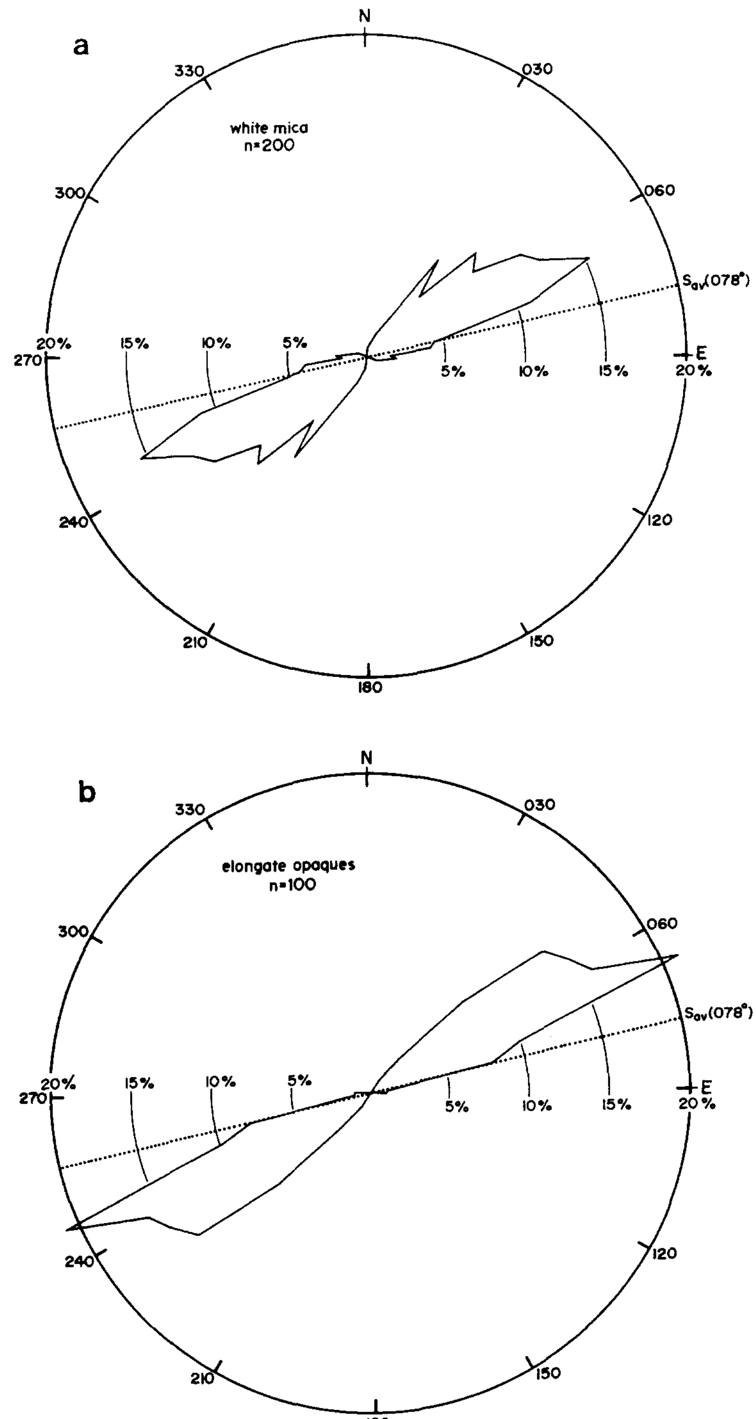

Fig. 14. Rose diagrams of (a) 200 matrix white mica and (b) 100 elongate opaque long axis orientations, measured from sample shown in Fig. 4(b). Outcrop-scale foliation orientation marked.

phyllosilicate grains parallel to segments of old grains and along kink band boundaries. This represents one cycle of the transposition and recrystallization process. The number of cycles experienced by a particular volume of the rock is determined by the magnitude of shear. This process reflects what is happening to the compositional layering at outcrop-scale, differing only in that it involves recrystallization and growth of new mica grains. Like the outcrop-scale foliation, this schistosity is also a steady-state foliation (Means 1981, Burg 1986), but is defined by rotated and new grains rather than by flattened old grains.

The processes outlined above give rise to microstructures and dimensional preferred orientations very similar to those of experimental studies (e.g. Williams *et al.* 1977, Means *et al.* 1984) and of studies of natural rocks (e.g. Knipe & White 1977, Williams *et al.* 1977, Williams 1985). From Figs. 17 and 18 it can be seen how the trimodal fabric (Figs. 14–16) has developed. The principal and smallest maxima are due to the long and short limbs, respectively, of kinked phyllosilicate grains and aggregates. Both maxima are augmented by new grains that grew mimetically on the rotated grains. The intermediate maximum is due to the new grains that grew

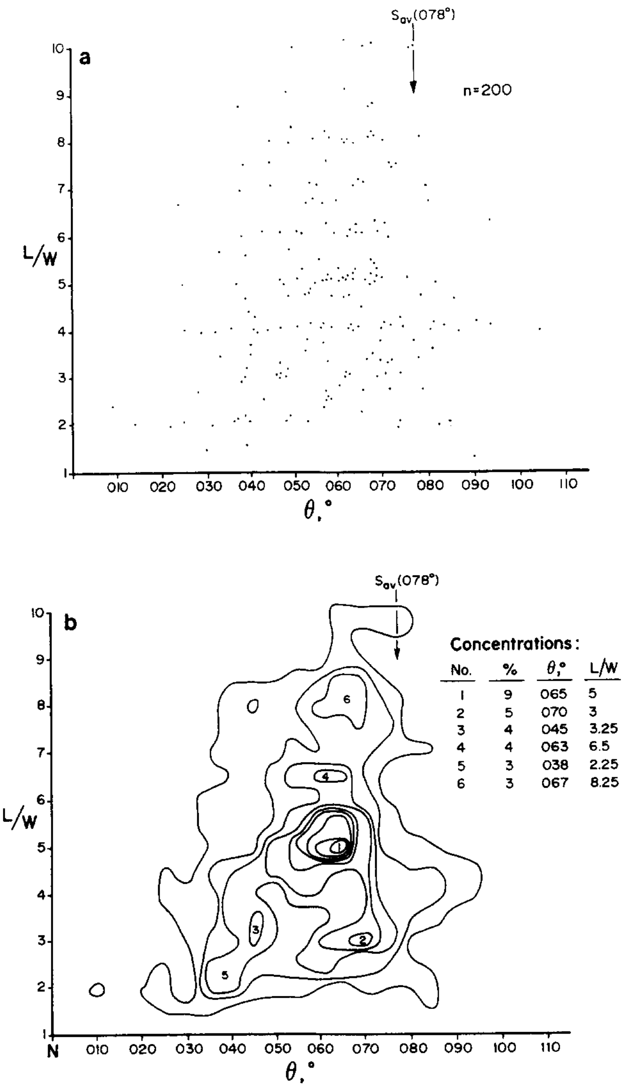

Fig. 15. Graphs of trend θ of 200 matrix white mica long axes vs individual L/W ratios, from sample shown in Fig. 4(b). (a) Scatter diagram; (b) data from (a) contoured with respect to density (the counting cell used was 10° by one L/W unit). See text for discussion.

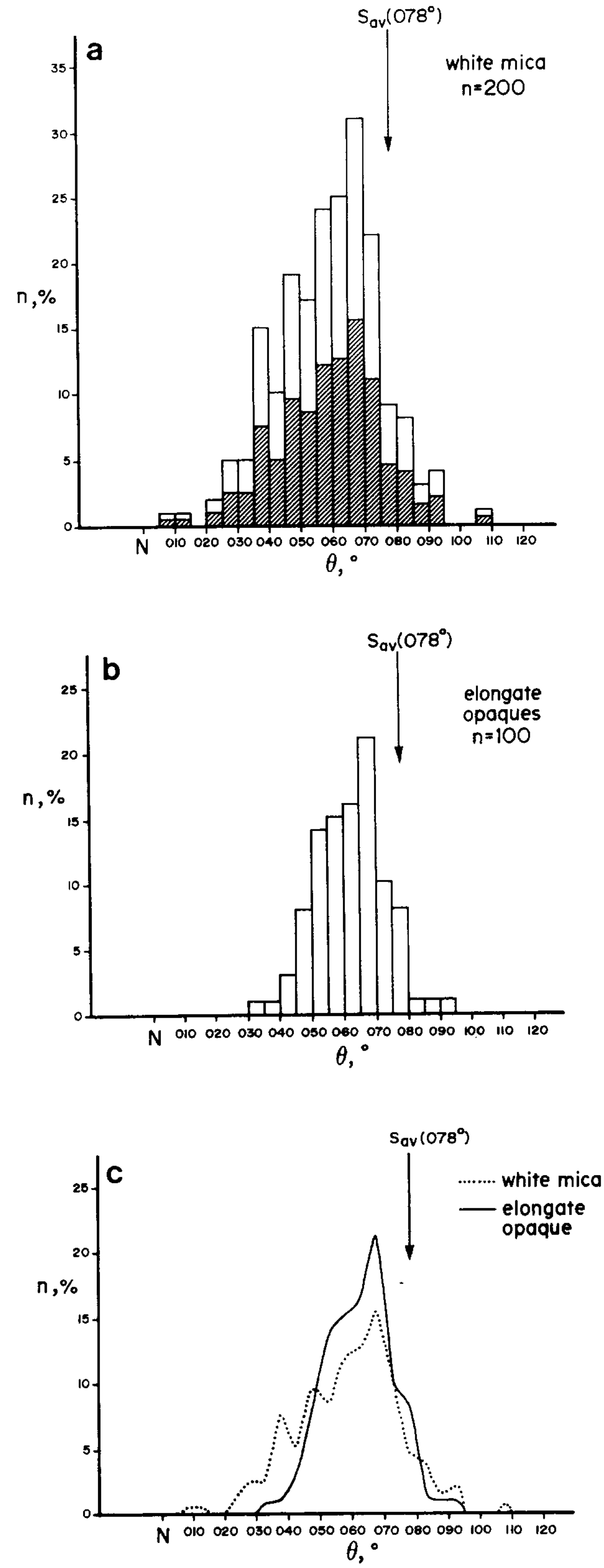

Fig. 16. Histograms of trend data for (a) 200 matrix white mica (shaded bars are percentages) and (b) 100 elongate opaque grains, (c) shows percentage curves for both data sets superimposed for comparison. Outcrop-scale foliation orientation marked.

along kink band boundaries; the kinks are asymmetric because they grew as microscopic drag folds in response to dextral shearing.

As described above, the preferred orientation of the elongate opaque minerals is similar to that of the phyllosilicate grains but lacks well-defined subsidiary maxima. This reflects a major difference in behaviour of the two minerals during deformation. The opaque minerals neither kinked nor recrystallized but simply rotated as rigid bodies. Their orientation pattern therefore reflects the general shape of the kinks. Thus, for both opaque grains and mica grains, the obliquity of the maximum preferred orientation to layering reflects the dominance of kink long limbs.

DETAILED ANALYSIS OF HORIZONTAL FOLDS

We have serially sliced several large specimens of phyllite which contain folded coticule layers. The chosen minor folds had hinge line trends essentially parallel to the shear zone and horizontal to shallowly plunging, and are tight to isoclinal in profile. They were sawn approximately perpendicular to their hinge lines into 1 cm thick

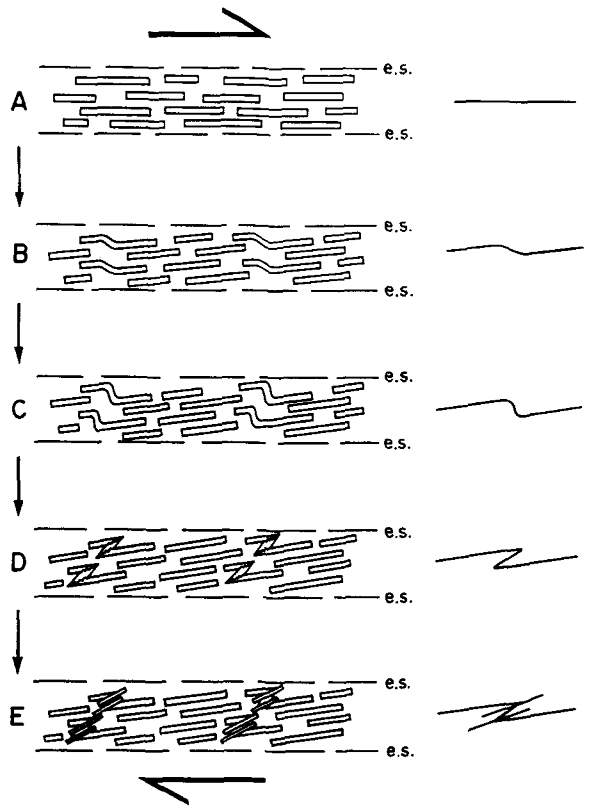

Fig. 17. Schematic diagram showing one sequence of microscale transposition (A–D) and dynamic recrystallization (E) during shearing in the phyllite matrix. Initial perturbations in the foliation are not shown in stage A. See text for discussion.

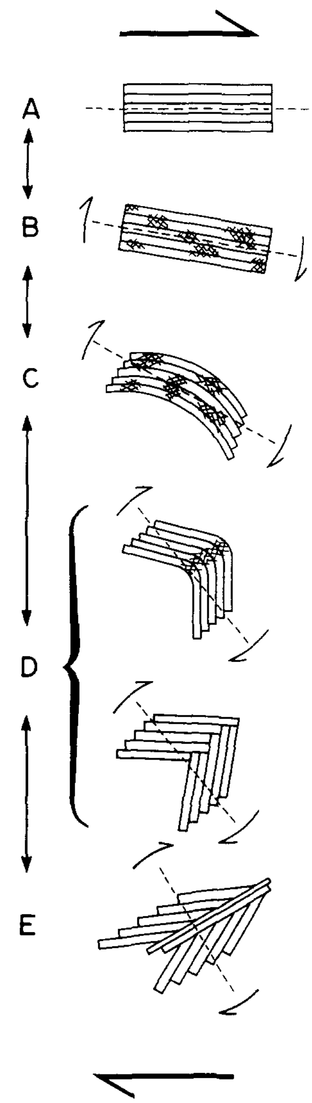

Fig. 18. Schematic diagram of one sequence of single-grain rotation, kinking and transposition (A–D) and dynamic recrystallization (E) in the phyllites. Cross-hatching indicates intracrystalline deformation. See text for discussion.

slices and thin sections were made perpendicular to the the axial surfaces of the folds both parallel and perpendicular to the hinge lines. These slices and sections demonstrate considerable variation in fold profiles and coticule layer thicknesses.

Figure 19 shows folded coticule layers traced directly from serial slices of three specimens. In each case the first slice (numbered 1) is the easternmost and numbers increase sequentially to the west. The tracings show a number of interesting features. First, fold profiles are markedly variable along the hinge line. Folds change size, shape, tightness, roundness and relative hinge-to-limb area rapidly. Their amplitudes and wavelengths are similarly variable, and individual folds can amplify, remain more-or-less constant in size or die out. Second, the thickest part of the layer migrates in relation to the fold hinge, along the length of the fold. In general the migration is constant in sense, in any one specimen, but it can show reversals (e.g. Fig. 19c). As a corollary to this migration, the layer thickness at the fold hinge must vary along the hinge line. Figure 20 shows the positions of thickest parts of the layers from profile to profile along the hinges. This variation is shown for the three specimens in Fig. 21.

We interpret the thickest parts of the folded layers to represent earlier fold hinge positions (perhaps at the metamorphic peak?). If this is so, then there has been fold hinge migration during progressive deformation of the coticule layers. Further persuasive evidence for hinge migration is seen by examining the positions of: (i) ubiquitous fine-grained, triangular, quartz-rich 'shark's fins' on the outer arcs of the folds; and (ii) concentrations of small garnets on the inner arcs of the folds (Fig. 22). These correlate very well with the thickest parts of the folded coticule layers in the sequence 'shark's fin'—thick part of layer—garnet concentration, going from outer to inner arc of individual folds. We interpret these features to be due to metamorphic differentiation. The quartz-rich 'shark's fins' formed as domains in strain-shadowed areas at fold hinge outer arcs due to quartz precipitation from solution (cf. Williams 1979), and the garnet-rich areas represent insoluble residue concentrated within fold hinge inner arcs as folds were progressively tightened and quartz was dissolved and removed. Neither feature corresponds to the present fold hinge position, and their location supports our deduction that the hinges have migrated.

Furthermore, the *thinnest* parts of the coticule layers i.e. pinches and boudin necks, show migration sympathetic with the thickest parts along the folded layers (see Fig. 19). The extension represented by these structures

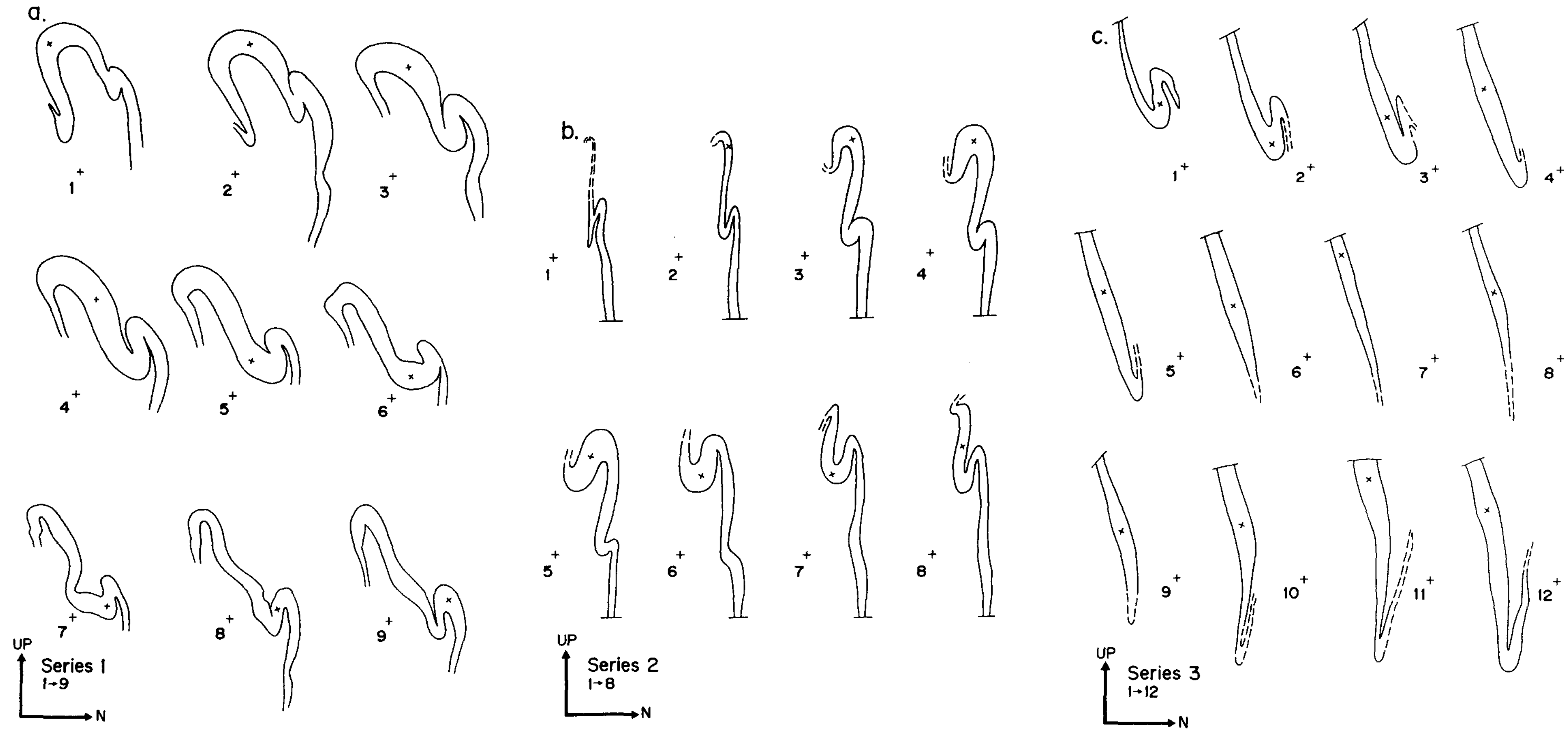

Fig. 19. (a)–(c) Serial sections of three samples of folded coticule layers, traced directly from serial slices (1 cm thick). Small crosses within layer mark thickest layer position. Large cross immediately above section number indicates constant spatial reference point, with respect to N–S and up-down. See text for discussion.

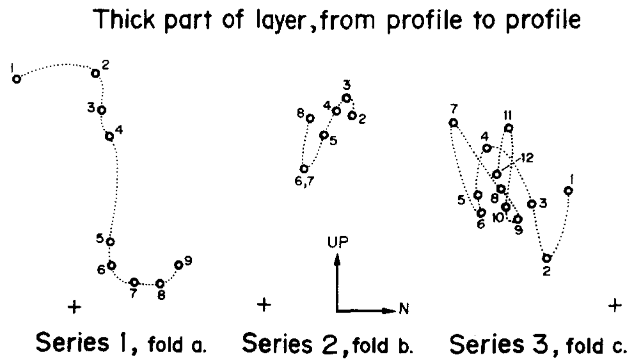

Fig. 20. Graphs of position of thickest parts of folded layers from section to section, with respect to constant spatial reference point (same large crosses as in Figs. 19a–c). Numbers correspond to sections of Figs. 19(a)–(c).

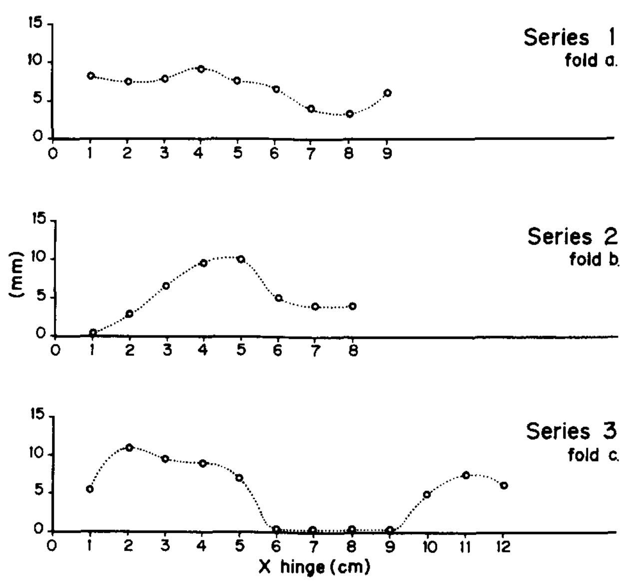

Fig. 21. Graphs of layer thickness at fold hinge vs position along fold hinge for folds shown in Figs. 19(a)–(c). Numbers along the x axes correspond to sections in those figures.

is almost certainly coeval with hinge migration, as predicted by our model (see below); boudin necks are generally filled by quartz-rich material identical to that of the 'shark's fins', and also represent precipitation sites.

The evidence presented above indicates that there has been migration of the hinges of the rotated folds (cf. Gray 1981, Treagus & Treagus 1981). Since there is no evidence of this migration in the profiles of steeply plunging folds, we conclude that the hinges migrated during fold rotation.

A possible model for this migration follows, based on the assumption that competent layers will maximize spin in a deformation that involves vorticity (e.g. simple shearing). The case for this argument has been made by Lister & Williams (1983). Figure 23 shows a horizontal fold of a perfectly passive layer with an inscribed rectangular grid in a block undergoing homogeneous simple shearing. If we consider two-dimensional strain in the plane of layering during this deformation, it can be seen that the horizontal grid lines will not rotate whereas the lines initially normal to them rotate progressively towards the movement vector. During non-coaxial deformation in a real material containing competent layers, however, deformation will be heterogeneous and competent layers will tend to maximize spin. In the extreme case (perfectly stiff competent layers) this means that original squares on the layer surface would remain squares, and in a real geological situation means that these areas shear less for a given bulk strain than if the bulk strain were homogeneous (as in Fig. 23). The extreme case is shown in Fig. 24 and can be simulated by folding a piece of paper and moving the limbs in opposite directions as appropriate to simple shearing. The paper deforms by rapid migration of the fold hinge since, given the mechanical constraints that operate, it cannot undergo any significant strain in the plane of the paper. A rock layer maximizing spin would undergo the same rotation but with a smaller magnitude of the ratio of hinge rotation to bulk strain compared with the paper

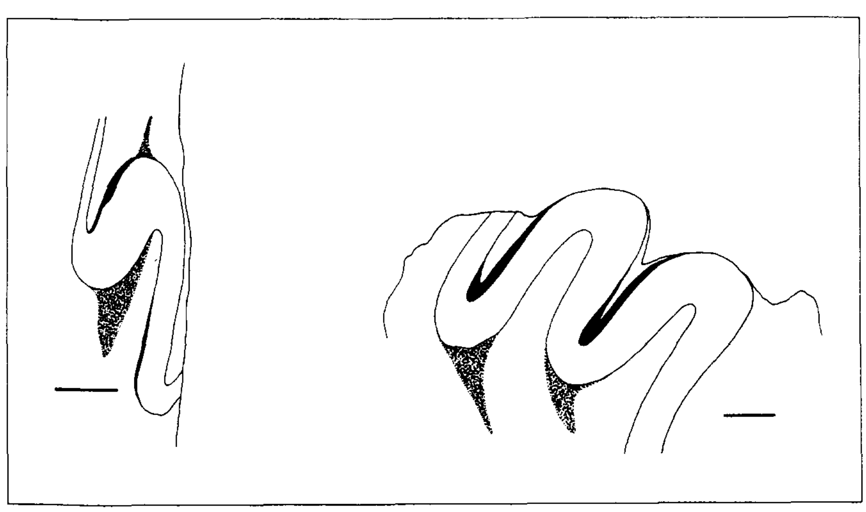

Fig. 22. Tracings from folded coticule layers in same outcrops as those of Figs. 19(a)–(c), showing that the present fold hinges do not correspond with the thickest parts of layers, or with quartz-rich 'shark's fin' segregations (stippled), or with insoluble garnet-rich residue zones (black). See text for discussion. Scale bars are 1 cm long.

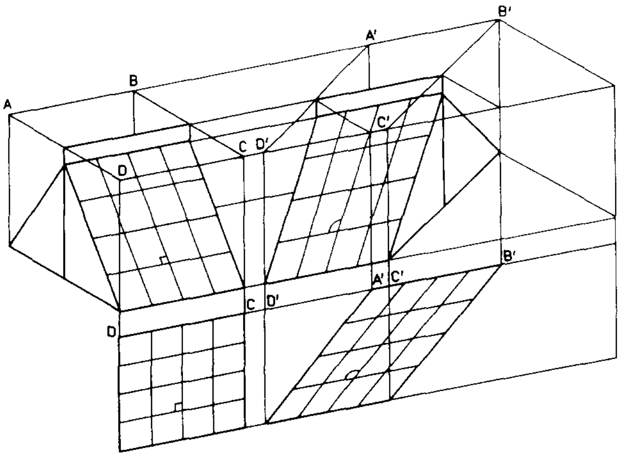

Fig. 23. Horizontal fold in perfectly passive layer undergoing homogeneous dextral simple shearing. Note that grid lines parallel to shearing vector (and hence fold hinge) will not rotate, those perpendicular to it will progressively rotate towards the vector's orientation. This is one extreme of behaviour of a folded layer during shearing. See text.

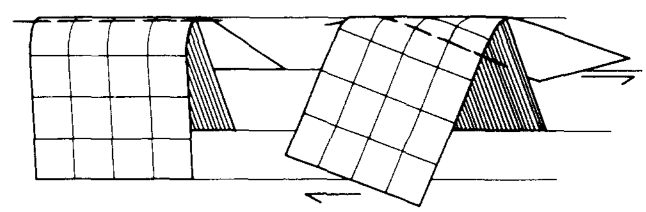

Fig. 24. A grid inscribed on an inextensible horizontally folded layer, subject to dextral simple shearing, will rotate with respect to the shearing vector but grid lines will not change their orientation on the folded layer with respect to one another. The original fold hinge (dashed line) must rotate in this situation; this is the second extreme of behaviour of a folded layer during shearing. It is probable that real deformation behaviour of competent folded layers will fall somewhere between those shown in Figs. 23 and 24. See text.

model. The sense of hinge rotation predicted by this model (Fig. 24) for dextral shearing is the same as that observed in the rocks described here. If an original axial plane is not obliterated by this process it follows that the foliation will no longer be parallel to the axial plane of the fold, but will have the geometry of a transecting cleavage.

GENERAL DISCUSSION AND CONCLUSIONS

One general process, operating at various scales, is primarily responsible for all of the structures described above, including folds cross-cut by the foliation; this is the process of transposition *sensu* Sander, operating during a progressive shearing deformation. It produces asymmetrical folds in layers such as the coticules and thereby rotates layering towards the large-scale shear zone orientation. It also produces microfolds that transpose the earlier foliation, which then also rotates towards the shear zone orientation as shearing continues. This last process is also influenced by recrystallization and new mineral growth but the foliation is primarily a product of transposition. Since the process must be cyclic and since the cycles are not necessarily in step at different scales, nor in different positions in the deforming rock mass, foliation and layering are not necessarily parallel and the foliation can cut across the fold profile (as in Fig. 25d) and at the same time be inclined to the fold axis as described above.

Thus the folds and foliation are forming synchronously, although at any one place they may not be of precisely the same age, but rather of overlapping ages. However, in view of this relationship it seems reasonable to consider the folds to be 'transected folds' *sensu stricto* (Borradaile 1978).

A number of writers (e.g. Borradaile 1978, Stringer & Treagus 1980, Treagus & Treagus 1981, Treagus 1988) have proposed models in which the foliation is assumed to develop parallel to the *XY* principal plane of the finite strain ellipsoid (used in the sense of the total strain ellipsoid for the tectonic deformation). These models offer no mechanism for foliation development but are valid so long as the foliation develops by a simple passive Marchian model (e.g. Tullis & Wood 1975). Similar models have been applied to shear zones by other writers including Sanderson *et al.* (1980), Murphy (1985) and Soper (1986).

Our model differs in several ways since it offers an actual mechanism for foliation development, and does not assume that the foliation tracks a principal plane of strain.

We envisage a continuous process of fold formation and transposition (Fig. 25). The size of the folds is determined by the size of the anisotropy in which they form. Thus coarse layering such as that of the coticules gives rise to much larger structures that the microfolds that form in the foliation. As a fold in a coticule layer rotates towards parallelism with the shear plane and shear vector, there is shearing parallel to the fold limbs and new microfolds are able to develop in the schistose layers. These microfolds perturb the existing foliation and progressively transpose it into a new orientation. As the new foliation rotates it converges on the shear plane but trails behind the axial plane of the larger fold. During development it is possible for such a foliation to track the principal plane of the incremental strain ellipsoid, but in view of the non-coaxial nature of the strain path and the fact that folding rather than simple passive Marchian behavior (March 1932) is involved, this is not very likely (see Williams 1976). Irrespective of whether or not the foliation tracks the incremental principal plane, since the foliation is transient it *cannot* track the principal plane of the finite strain ellipsoid, even if the strain history considered is restricted to that accumulating during movement on the shear zone.

A consequence of our interpretation of foliation development is that the relationship between foliation and folds is, and in most cases must be, variable (Fig. 25). Where folds and microfolds have initiated at exactly the same time, so long as outcrop-scale compositional layering and the early microscale foliation were exactly parallel, the new foliation will be parallel to the axial plane of the fold (Fig. 25b). Where a new cycle of transposition has occurred at the microscale but has not influenced the coarser layering, the foliation will 'tran-

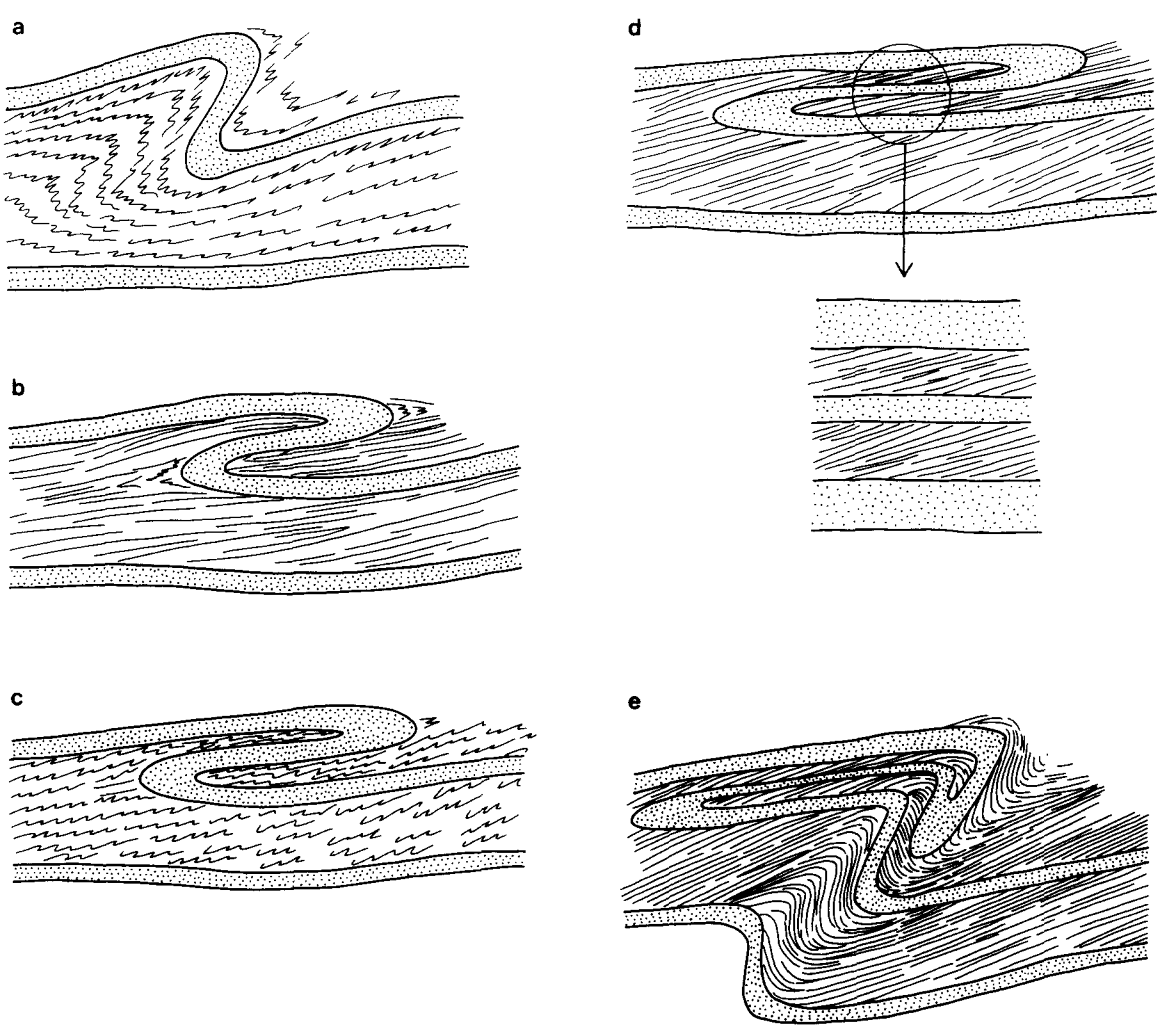

Fig. 25. Sketch diagrams of the various relationships possible between foliation and folds undergoing progressive transposition in the context of simple shearing. See text for discussion.

sect' the folds in layering (Fig. 25d). This is the situation most commonly observed in the rocks described above, and one which could be easily mistaken for two distinct, overprinting deformation events. Finally, where layering-scale folds have developed without concurrent or subsequent microscale refolding of the foliation, the latter will be refolded by the fold in layering but there will be neither a transecting foliation nor an axial plane foliation (Fig. 25e). All of these relationships are observed not only in the shear zone described here but in other ductile shear zones throughout the region (Karlstrom *et al.* 1982, Leger & Williams 1986).

Towards the end of a transposition cycle, at any given position in the shear zone, the foliation may be inclined to the principal plane of the finite strain ellipsoid only by a very small angle. However, as stated above, it will not generally be truly parallel to the XY principal plane and locally the deviation may be large. It follows that the orientation of the foliation in such cases cannot be used to determine shear strain magnitude (cf. Ramsay & Huber 1983, p. 184 *et seq.*) This technique is no doubt valid in some situations, for example where foliation is defined by clasts in a deformed conglomerate, but in any shear zone where the foliation is steady-state the method is inappropriate.

Finally we draw attention to the fact that rotation of the fold hinges involved hinge migration and that the magnitude of the migration is of the same order as the fold amplitude. This process may be more common than has been realized previously (see also Treagus & Treagus 1981, and Gray 1981), especially in deformations involving a large component of simple shearing, and may contribute to severe difficulties in unravelling the structural histories of such areas.

Acknowledgements—We thank Colleen Elliott, Bill Fyson, Bruno LaFrance, Win Means, Cees Van Staal and Joe White for discussions. Susan Aitken and Paul Chenard (UNB) drafted several figures. We thank P. W. G. Tanner and an anonymous reviewer for the excellent reviews, and Sue Treagus for her detailed comments and editorial patience beyond duty. The work was supported by an NSERC Operating Grant (to P. F. Williams), a contract from the Geological Survey of Canada (to C. K. Mawer and P. F. Williams), and Sandia National Laboratories (SURP Grant 86-5629, Task 7, to C. K. Mawer).

REFERENCES

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

Burg, J. P. 1986. Quartz shape fabric variations and c-axis fabrics in a ribbon-mylonite: arguments for an oscillating foliation. *J. Struct. Geol.* **8**, 123–131.

Clarke, D. B. & Halliday, A. N. 1980. Strontium isotope geology of the South Mountain batholith, Nova Scotia. *Geochim. cosmochim. Acta* **44**, 1045–1058.

Clarke, D. B. & Halliday, A. N. 1985. Sm/Nd isotope investigation of the age and origin of the Meguma Zone metasedimentary rocks. *Can. J. Earth Sci.* **22**, 102–107.

Eisbacher, G. H. 1970. Deformation mechanisms of mylonitic rocks and fractured granites in Cobequid Mountains, Nova Scotia, Canada. *Bull. geol. Soc. Am.* **81**, 2009–2020.

Fyson, W. K. 1966. Structures in the Lower Palaeozoic Meguma Group, Nova Scotia. *Bull. geol. Soc Am.* **77**, 931–944.

Gray, D. R. 1981. Cleavage-fold relationships and their implications for transected folds: an example from southwest Virginia, U.S.A. *J. Struct. Geol.* **3**, 265–277.

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

Karlstrom, K., van der Pluijm, B. A. & Williams, P. F. 1982. Structural interpretation of the eastern Notre Dame Bay area, Newfoundland: regional post-Middle Silurian thrusting and asymmetrical folding. *Can. J. Earth Sci.* **19**, 2325–2341.

Knipe, R. J. & White, S. H. 1977. Microstructural variation of an axial plane cleavage around a fold—a H.V.E.M. study. *Tectonophysics* **39**, 355–380.

Kramm, U. 1976. The coticule rocks (spessartine quartzites) of the Venn-Stavelot Massif, Ardennes, a volcaniclastic metasediment? *Contr. Miner. Petrol.* **56**, 135–155.

Lamens, J., Guekens, F. & Viaene, W. 1986. Geological setting and genesis of coticules (spassartine metapelites) in the Lower Ordovician of the Stavelot Massif, Belgium. *J. geol. Soc. Lond.* **143**, 253–258.

Leger, A. & Williams, P. F. 1986. Transcurrent faulting history of southern New Brunswick. *Geol. Surv. Pap. Can.* **86-1B**, 111–120.

Lister, G. S. & Williams, P. F. 1983. The partitioning of deformation in flowing rock masses. *Tectonophysics* **92**, 1–33.

March, A. 1932. Mathematisch Theorie der Regelung nach der Korngestalt bei Affiner Deformation. *Z. Kristallogr.* **81**, 285–294.

Mawer, C. K. & White, J. C. 1987. Sense of displacement on the Cobequid-Chedabucto fault system, Nova Scotia, Canada. *Can. J. Earth Sci.* **24**, 217–223.

Mawer, C. K. & Williams, P. F. 1986. Structural study of highly deformed phyllite and granite, vicinity of White Head Village, S.E. Nova Scotia. *Marit. Sedim. & Atlantic Geol.* **22**, 51–64.

Means, W. D. 1981. The concept of steady state foliation. *Tectonophysics* **78**, 179–199.

Means, W. D., Williams, P. F. & Hobbs, B. E. 1984. Incremental deformation and fabric development in a KCl/mica mixture. *J. Struct. Geol.* **6**, 391–398.

Murphy, F. C. 1985. Non-axial planar cleavage and Caledonian sinistral transpression in eastern Ireland. *Geol. J.* **20**, 257–279.

Powell, C. McA. 1974. Timing of slaty cleavage during folding of Precambrian rocks, northwest Tasmania. *Bull. geol. Soc. Am.* **85**, 1043–1060.

Quinquis, H., Audren, C. Brun, J.-P. & Cobbold, P. R., 1978. Intense progressive shear in Ile de Groix blueschists and compatibility with subduction or obduction. *Nature* **273**, 43–45.

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

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

Reynolds, P. H., Kublick, E. E. & Muecke, G. K. 1973. Potassium–argon dating of slates from the Meguma Group, Nova Scotia. *Can. J. Earth Sci.* **10**, 1059–1067.

Sander, B. 1911. Uber Zussamenhange zwischen Teilbewegung und Gefuge in Gesteinen. *Tschermaks miner. petrogr. Mitt.* **30**, 281–314.

Sanderson, D. J. 1979. The transition from upright to recumbent folding in the Variscan fold belt of southwest England: a model based on the kinematics of simple shear. *J. Struct. Geol.* **1**, 171–180.

Sanderson, D. J., Andrews, J. R., Phillips, W. E. A. & Hutton, D. H. W. 1980. Deformation studies in the Irish Caledonides. *J. geol. Soc. Lond.* **137**, 289–302.

Schenk, P. E. 1971. Southeastern Atlantic Canada, northwestern Africa, and continental drift. *Can. J. Earth Sci.* **8**, 1218–1251.

Schenk, P. E. 1976. A regional synthesis (of Nova Scotia geology). *Marit. sedim. Atlantic Geol.* **11**, 17–24.

Schenk, P. E. 1978. Synthesis of the Canadian Appalachians. In: IGCP Project 27, *Caledonian–Appalachian Orogen of the North Atlantic Region. Geol. Surv. Pap. Can.* **78-13**, 111–136.

Skjernaa, L. 1985. Some possible ways to develop sheath-folds. *Int. Conf. Tectonic and Structural Processes.* (Abs. Volume). Instituut voor Aardwetenschappen, Utrecht, 123.

Soper, N. J. 1986. Geometry of transecting, anastomosing solution cleavage in transpression zones. *J. Struct. Geol.* **8**, 937–940.

Stevenson, I. M. 1964. Geology, Chedabucto Bay, Nova Scotia. Map 1156A, 1:63,360. Geological Survay of Canada, Ottawa.

Stringer, P. & Treagus, J. E. 1980. Non-axial planar S_1 cleavage in the Hawick Rocks of the Galloway area, Southern Uplands, Scotland. *J. Struct, Geol.* **2**, 317–331.

Treagus, J. E. & Treagus, S. H. 1981. Folds and the strain ellipsoid: a general model. *J. Struct. Geol.* **3**, 1–17.

Treagus, S. H. 1988. Strain refraction in layered systems. *J. Struct. Geol.* **10**, 517–527.

Tullis, T. E. & Wood, D. S. 1975. Correlation of finite strain from both reduction bodies and preferred orientation of mica in slate from Wales. *Bull. geol. Soc. Am.* **86**, 634–638.

Voll, G. 1960. New work on petrofabrics. *Geol. J.* **2**, 503–567.

Webb, G. W. 1969. Paleozoic wrench faults in Canadian Appalachians. In: *North Atlantic Geology and Continental Drift* (edited by Kay, M.). *Mem. Am. Ass. Petrol. Geol.* **12**, 754–786.

Williams, H. & Hatcher, Jr. R. D. 1982. Suspect terranes and accretionary history of the Appalachian orogen. *Geology* **10**, 530–536.

Williams, P. F. 1972. Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia. *Am. J. Sci.* **272**, 1–47.

Williams, P. F. 1976. Relationships between axial plane foliations and strain. *Tectonophysics* **30**, 181–196.

Williams, P. F. 1979. The development of asymmetrical folds in a cross-laminated siltstone. *J. Struct. Geol.* **1**, 19–30.

Williams, P. F. 1985. Multiply deformed terrains—problems of correlation. *J. Struct. Geol.* **7**, 269–280.

Williams, P. F., Means, W. D. & Hobbs, B. E. 1977. Development of axial-plane slaty cleavage and schistosity in experimental and natural materials. *Tectonophysics* **42**, 139–158.

Williams, P. F. & Zwart, H. J. 1977. A model for the development of the Seve-Koli Caledonian nappe complex. In: *Energetics of Geological Processes* (edited by Saxena, S. K. & Bhattacharji, S.). Springer-Verlag, New York, 169–187.