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Deformation path partitioning within a transpressive shear zone, Marble Cove, Newfoundland

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Abstract—Detailed study of the structures in Marble Cove, Newfoundland, indicates that they formed during a single transpressional event, during which deformation was partitioned between dominantly strike-slip and dominantly dip-slip domains. Marble Cove exhibits steeply dipping foliations, steeply to shallowly plunging stretching lineations, greenschist facies mineral assemblages, and abundant quartz veins. Within the broad domain where stretching lineations plunge steeply, sense-of-shear indicators and asymmetry of quartz-*c*-axis distributions in mylonitized quartz veins indicate dominantly reverse motion with a component of dextral shear. Adjacent greenschists exhibit orthorhombic symmetry and apparently accommodated pure shear. The abrupt transition to a narrow, high strain domain of shallowly plunging stretching lineations is accommodated by a rheologically distinct serpentinite body. Shear bands in this domain indicate dextral strike-slip movement with a reverse component. Smaller, conjugate shear bands are evident in thin section, indicating a component of pure shear. Sense-of-shear indicators and asymmetry of *c*-axis distributions in mylonitized quartz veins also record dextral strike-slip and reverse shear with a component of pure shear. Crenulations and asymmetric folds in foliation overprint all other fabrics in Marble Cove and record a final phase of reverse shear with a dextral component.

With partitioning of this kind, structures recording strike-slip movement may be limited in areal extent, and therefore less likely to be exposed, than structures recording dip-slip motion. Such partitioning within transpressional shear zones may therefore impede recognition of strike-parallel motions within orogenic belts.

INTRODUCTION

Areas of transpression, which accommodate both transcurrent and convergent motion, often resulting from oblique collision at plate boundaries (Harland 1971), have been described from the deeper crustal levels of many ancient orogenic belts (e.g. Brun & Burg 1982, Ratliff *et al.* 1988, Borradaile *et al.* 1988, Caron & Williams 1988a,b, Hudleston *et al.* 1988, Holdsworth 1989, 1991, Hansen 1989, Holdsworth & Strachan 1991, Vauchez & Nicolas 1991, D'Lemos *et al.* 1992, Robin & Cruden 1994). Many of these areas exhibit structures, such as mineral lineations, folds, foliations, and shear zones, with orientations that vary within the study area. In general, these have been interpreted as recording partitioning of strain in order to accommodate contraction, extension, and transcurrent motion either sequentially (e.g. Ratliff *et al.* 1987, Hansen 1989) or simultaneously (e.g. Brun & Burg 1982, Borradaile *et al.* 1988, Caron & Williams 1988a,b, Hudleston *et al.* 1988, Holdsworth 1989, 1991, Holdsworth & Strachan 1991, Robin & Cruden 1994).

Several mathematical models have been developed in order to understand the factors controlling both the three-dimensional distribution of structures of different orientations and partitioning of deformation within transpression zones. Sanderson & Marchini (1984) considered transpression within a laterally confined zone experiencing a combination of horizontal shortening, vertical lengthening, and strike-slip translation. Homogeneous deformation of the zone was modeled by the application of a deformation matrix that could be factorized into pure shear and simple shear components. Fossen & Tikoff (1993) modified this approach with the use of a deformation matrix for simultaneous simple shearing, pure shearing, and volume change. In both of the above models, deformation is homogeneous within the transpression zone, so structures can vary in orientation only with variations in strain or in the imposed components of shortening relative to translation. Tikoff & Teyssier (1994) and Teyssier *et al.* (1995) subsequently modified the approach of Fossen & Tikoff (1993) to model partitioning of strike-slip and contractional deformation between discrete faults and diffuse deformation zones, such as that found in shallow crustal regions of orogenic belts (e.g. Walcott 1978, Mount & Suppe 1987, Oldow *et al.* 1990, Pinet & Cobbold 1992).

Modeling of partitioning in deeper crustal, ductile regimes requires a different approach. Sanderson & Marchini (1984) discussed difficulties with the physical model that forms the basis for their work and studies by Fossen & Tikoff (1993), Tikoff & Teyssier (1994) and

Teyssier *et al.* (1995), particularly in modeling deformation in the ductile field. Principal among these concerns is the mechanical character of the boundary between the transpression zone and the rigid blocks within which the zone is confined. The boundary allows for frictionless extrusion of material vertically, but maintains a coherent bond with material horizontally. In other words, material cannot slip in the horizontal direction, but must slip freely in the vertical direction. Robin & Cruden (1994) have modified this approach in order to more realistically model ductile transpression zones. In their physical model, material is not allowed to slip freely in any direction along the boundary between the transpression zone and surrounding rigid blocks. This constraint necessarily results in a heterogeneous distribution of strain across the transpression zone. In addition, they have considered both strike-slip and oblique translation across the transpression zone, as well as variable amounts of shortening ('press' component) and translation ('trans' component) in their continuum mechanical models. Robin & Cruden (1994) in this way successfully modeled the structures present in the Archean Larder Lake-Cadillac deformation zone of Canada and the Proterozoic Mylonite Zone of southwest Sweden. With oblique transpression, they determined that the orientations of 'foliation' (plane perpendicular to the direction of minimum principal strain rate) and 'lineation' (direction of maximum principal strain rate) will vary asymmetrically across the zone of transpression, effectively partitioning the zone into regions of dominantly dip-slip and dominantly strike-slip deformation. By varying 'press' and 'trans' components in transcurrent transpression, they produced a regularly oriented 'foliation' and a 'lineation' that varies symmetrically in orientation within the zone. More important, they demonstrated that, with a large 'press' component, the axis of rotation, or vorticity vector, will vary with respect to the lineation. Rather than being constrained to an orientation perpendicular to the 'lineation' (the two-fold axes of symmetry of a monoclinic system, cf. Passchier & Simpson 1986) it will be oblique to parallel to the 'lineation', and the tectonite will have triclinic symmetry.

In this paper, we report on the development of complex structures in a transpressional shear zone, illustrating one way in which both strike-slip and dip-slip motions may be accommodated at depth within an orogenic belt. Within Marble Cove, Newfoundland, we find a stretching lineation and associated foliation that vary in orientation asymmetrically across the transpression zone. These variations differ from those that would be expected from the model of Robin & Cruden (1994); however, we suggest that these differences may relate to the variation in lithology within Marble Cove as well as the degree of deformation exhibited by the 'blocks' that confine the transpression zone. In addition, the results of this and the study of Robin & Cruden (1994) suggest that the structures most easily linked to strike-parallel motion may be limited in areal extent, and are therefore less likely to outcrop, than structures recording dip-slip motions. The type and complexity of structures developed in ductile transpression may therefore impede recognition of strike-parallel motions in orogenic belts.

REGIONAL GEOLOGICAL SETTING: THE HUMBER/DUNNAGE BOUNDARY

The Baie Verte-Brompton Line (Williams & St-Julien 1978, 1982) is a regionally extensive boundary separating rocks of continental margin affinity, which form the eastern edge of the Humber Zone, from oceanic rocks of the Dunnage Zone (inset, Fig. 1). This boundary was defined as a "narrow structural zone marked by discontinuous ophiolite complexes" (Williams & St-Julien 1978), a definition which, in Newfoundland, best fits the section of the boundary exposed on the Baie Verte Peninsula. Deformation associated with the boundary, however, extends over a broader zone on the peninsula. Within this broad zone, evidence for early thrusting has been well documented, and is recorded in part by small mafic and ultramafic bodies that have been structurally interleaved with the metasedimentary rocks of the Fleur de Lys Supergroup west of the Baie Verte-Brompton Line (Bursnall 1975, Williams 1977, Williams *et al.* 1977, van Berkel *et al.* 1986, fig. 1). Williams & St-Julien (1978, 1982) drew the Baie Verte-Brompton Line through greenschist and mélange of the Birchy Complex where it extends to the northeast coast of the Baie Verte peninsula (Fig. 1). Hibbard (1982, 1983) suggested that the Baie Verte-Brompton Line is represented by the Marble Cove Slide (Fig. 2). In either case, the broad zone of deformation along the boundary between the Humber and the Dunnage zones includes Marble Cove, the subject of this paper.

Ductile structures along the Baie Verte-Brompton Line were attributed by most earlier workers to dip-slip motions interpreted to record episodes of orthogonal collision in the northern Appalachians (Kidd 1974, Bursnall 1975, Bursnall & DeWitt 1975, Williams 1977, Williams *et al.* 1977, Hibbard 1982, 1983, Williams & St-Julien 1982, Dewey *et al.* 1983). Most of these authors suggested two major deformation events: westward-directed thrusting of the Dunnage above the Humber Zone (interpreted to be associated with the Ordovician Taconic orogeny), followed by eastward-directed thrusting (interpreted as part of the Devonian Acadian orogeny). Structures west of the Baie Verte-Brompton Line are generally believed to be largely related to the earlier episode of deformation; overprinting structures correlated with the younger deformation become increasingly dominant with increasing proximity to the line (Bursnall & DeWitt 1975, Hibbard 1983). Shear zones both along and adjacent to this boundary, however, dip steeply to vertically — orientations which are incompatible with an origin by thrusting and suggest remobilization of the boundary zone. Kennedy (1975) first noted that the major faults that form the boundaries between many of the tectonostratigraphic zones in Newfoundland recorded components of strike-slip movement, citing two main lines of

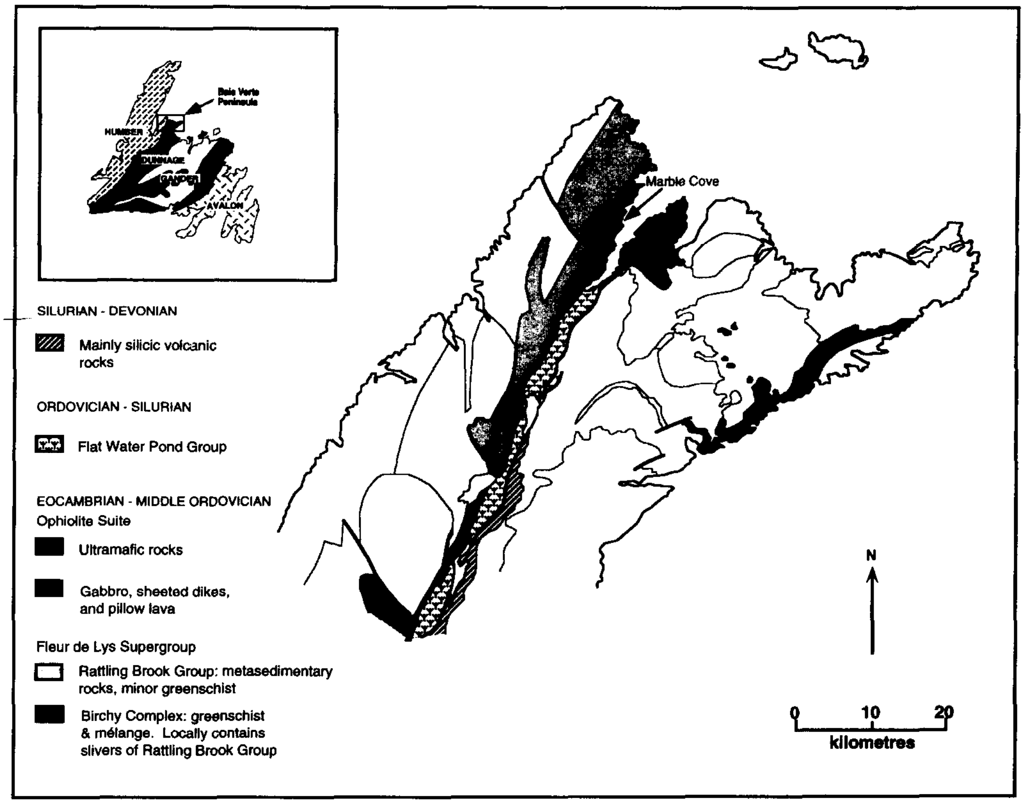


Fig. 1. Inset map shows the location of the Baie Verte Peninsula relative to tectonostratigraphic zones delineated by Williams *et al.* (1988). Enlargement of the Baie Verte Peninsula highlights the lithologic units extended, at least locally, along the Baie Verte-Brompton Line (modified from Hibbard 1982). Contacts between units which are not considered here are shown in gray. Tectonic boundaries such as faults are shown with bold lines; barbs on thrust faults point towards the hanging-wall. The location of Marble Cove is indicated.

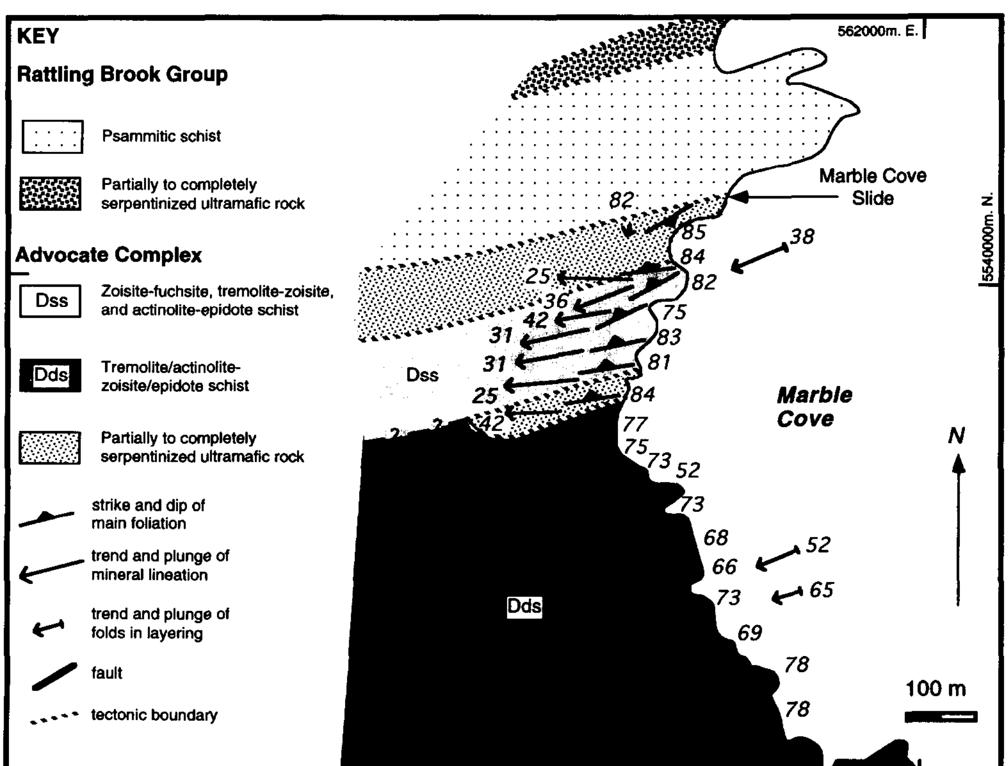


Fig. 2. Detailed map of Marble Cove, modified from Bursnall & Hibbard (1980), showing variation in orientation of main foliation, mineral lineation, and hinges to folds in compositional layering. Dominantly strike-slip domain (Dss) and dominantly dip-slip domain (Dds) are labeled. Linear data are illustrated by symbols with length scaled according to plunge: the shallower the plunge, the longer the symbol, and vice versa. Some symbols represent the mean value of several measurements taken in a small area.

evidence: (1) the faults have straight traces and separate zones of different stratigraphy over large distances, and (2) the faults bound regions of slightly different tectonic strike. Subsequent studies have documented evidence for strike-slip motion along the Baie Verte-Brompton Line in Newfoundland (Piasecki 1988, Goodwin & Williams 1989, 1990, B. Dubé pers. comm. 1994) and Québec (Kirkwood 1989, Malo & Béland 1989, Malo *et al.* 1992). The Newfoundland workers report structures that record both dextral and sinistral strike-slip motion, although their interpretations of the relative timing and significance of these events differs. These differences will not be addressed here. The majority of these workers (Kirkwood 1989, Malo & Béland 1989, Malo *et al.* 1992, L. B. Goodwin unpub. data, B. Dubé pers. comm. 1994) have found evidence of two episodes of dextral strike-slip motion. The deformation in Marble Cove correlates with the younger of these episodes of dextral motion.

The Humber Zone is generally believed to contain rocks that range from Precambrian to Ordovician in age (see review by Hibbard 1983). Peak metamorphism (amphibolite facies) is associated with porphyroblast growth overprinting the regional foliation (Jamieson & Vernon 1987) in the Early Silurian (Cawood *et al.* 1994). This regional foliation was created during the early phase of dextral strike-slip deformation, which post-dates thrusting (Goodwin & Williams 1989, 1990). The younger episode of dextral strike-slip deformation, associated with greenschist facies metamorphism, is therefore younger than Early Silurian on the Baie Verte Peninsula. In Québec, this phase of deformation is constrained to be Devonian in age (Kirkwood 1989, Malo & Béland 1989, Malo *et al.* 1992).

THE MARBLE COVE SLIDE

The Marble Cove Slide was first recognized by Watson (1947) and named and described by Bursnall (1975) as a structure that developed at a relatively late stage in the deformational history of the Baie Verte Peninsula. The Marble Cove Slide juxtaposes the rocks of the easternmost Birchy Complex and a sliver of the Rattling Brook Group against rocks of the Advocate Complex (Figs. 1 and 2). The latter were entirely derived from mafic igneous intrusive and extrusive protoliths. Original rock textures are visible in locally preserved lenses that experienced less extensive alteration and deformation than the surrounding rocks.

Psammitic schists of the Rattling Brook Group in Marble Cove exhibit multiple foliations and complex overprinting relationships, consistent with a long and complex deformational history. The main foliation in these psammitic schists strikes northeast (Bursnall 1979, Bursnall & Hibbard 1980). The Marble Cove Slide separates these psammitic schists from rocks of the Advocate Complex in Marble Cove, which exhibit structures that are quite regular in orientation and character. The main foliation in the Advocate Complex strikes east–northeast (Fig. 2). This foliation and associated structures are the focus of the study presented here.

STRUCTURES IN MARBLE COVE

We have divided the rocks of the Advocate Complex into two structural domains on the basis of the character and orientation of macroscopic structures. In particular, we consider the variation in pitch of a mineral lineation present throughout the study area. The dominantly dip-slip domain (Dds) is distinguished by a steeply to moderately plunging mineral lineation, generally pitching $>45°$ in the foliation plane. The dominantly strike-slip domain (Dss) is characterized by a shallowly to moderately plunging mineral lineation, typically pitching $<45°$ in the foliation plane. In addition to the mineral lineation, several foliations are present in Marble Cove. The main foliation is parallel to compositional layering, except where it crosses the hinges of folds in compositional layering. A second foliation, defined by the preferred dimensional alignment of dynamically recrystallized grains (cf. Simpson & Schmid 1983), is evident in thin sections of quartz vein mylonites in both domains. We follow Law *et al.* (1984) in calling the main foliation S_A and the shape foliation defined by recrystallized grains S_B. Shear band foliations (cf. Berthé *et al.* 1979, White *et al.* 1980, Gapais & White 1982; extensional crenulation cleavages of Platt 1979 and Platt & Vissers 1980; C′ foliation of Ponce de Leon & Choukroune 1980) are also evident in Dss.

In the next sections we describe these structures in detail, focusing on comparison of the two structural domains in the Advocate Complex. In the Discussion, we consider the significance of differences between structures developed in Dss and Dds and propose a model for formation of all of the structures in Marble Cove during a single deformational event.

Mineral parageneses and fabric-forming elements

A petrographic study of samples from the two structural domains indicates that they both contain rocks with the same mineral assemblages, though the dominant assemblage is different in each domain. Parageneses indicative of greenschist facies metamorphism, common to both domains, are: (1) tremolite + zoisite + quartz + albite + sphene ± chlorite ± carbonate (more common in Dss), and (2) actinolite + epidote + quartz + albite + chlorite + sphene ± carbonate (dominant in Dds). In addition, zoisite–fuchsite schist is found in Dss. The main foliation, S_A, is defined in both structural domains by the preferred dimensional alignment of the amphiboles, epidote group minerals, chlorite, and, locally, fuchsite and is subparallel to compositional layers and laminae (Figs. 3a–c). A laminar fabric is evident in sections cut parallel to the mineral lineation and at right angles to S_A (Figs. 3a & b). In sections cut at right angles to both S_A and the

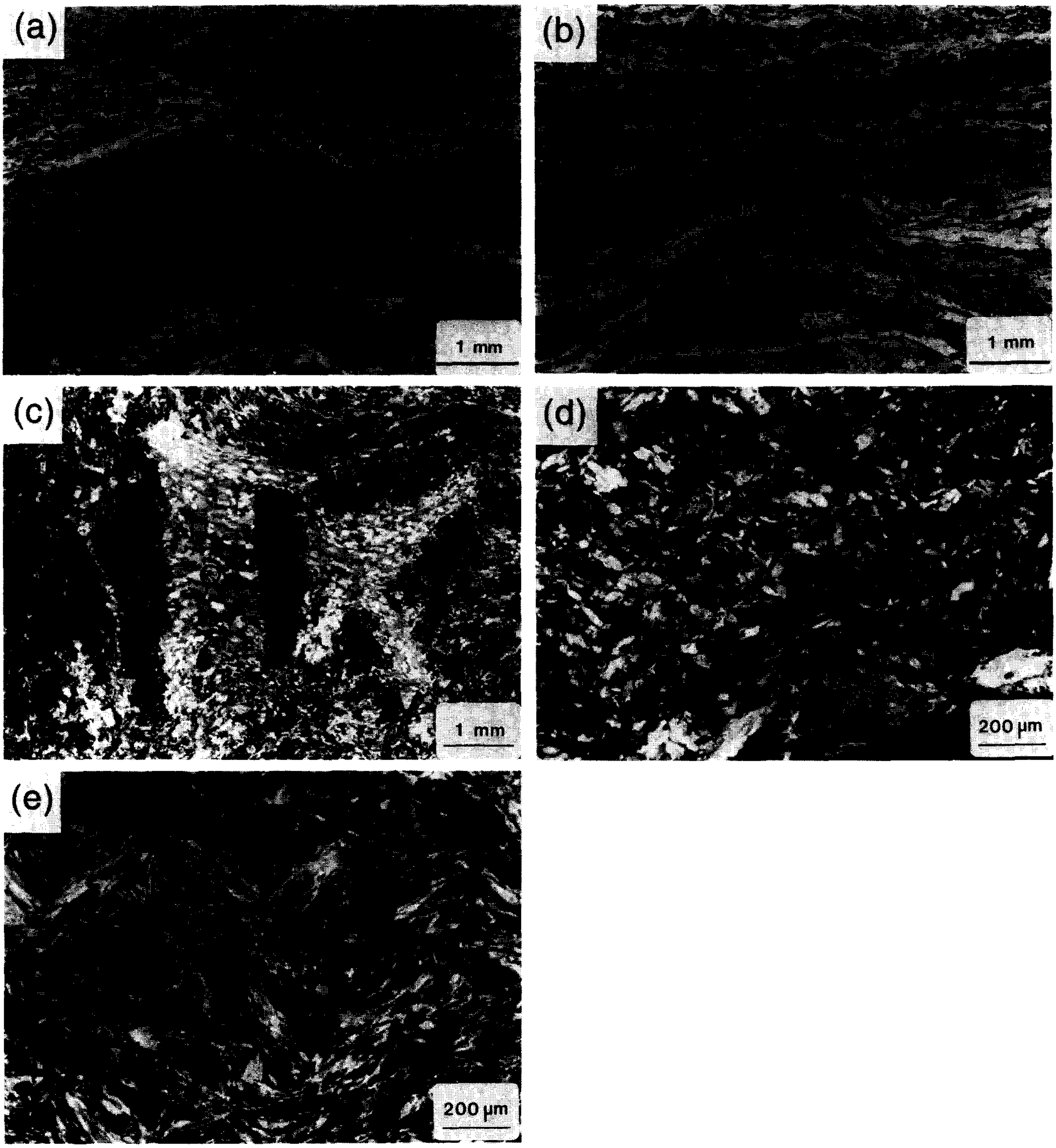


Fig. 3. Photomicrographs of greenschists: (a) and (b) taken in plane light; (c, d, e) taken with crossed polars. Sections cut at right angles to main foliation, S_A; trace of S_A is subhorizontal in all photos. (a, b, c, d, and e) are viewed parallel to stretching lineation, direction of plunge to the left; (d) is perpendicular to stretching lineation. (a) Dss: Dominant shear bands indicate dextral strike-slip sense of shear (shown by arrows). (b) Dss: Same sample as that shown in (a). Conjugate shear bands (arrows) are locally evident, and record a component of shortening at right angles to S_A. (c) Dss: Clusters of epidote-group minerals (E) are commonly extended parallel to lineation, with quartz (Q) filling in gaps between boudins. Gypsum plate inserted to highlight quartz grains, which are elongate parallel to lineation. (d) Dds: Amphibole is the main fabric-forming mineral. At right angles to lineation, basal sections through lineation-forming amphibole grains are evident (e.g. A); foliation is anastamosing and difficult to distinguish. (e) Foliation and lineation forming grains (in this case, amphibole) are bent and broken along hinges of crenulations (C).

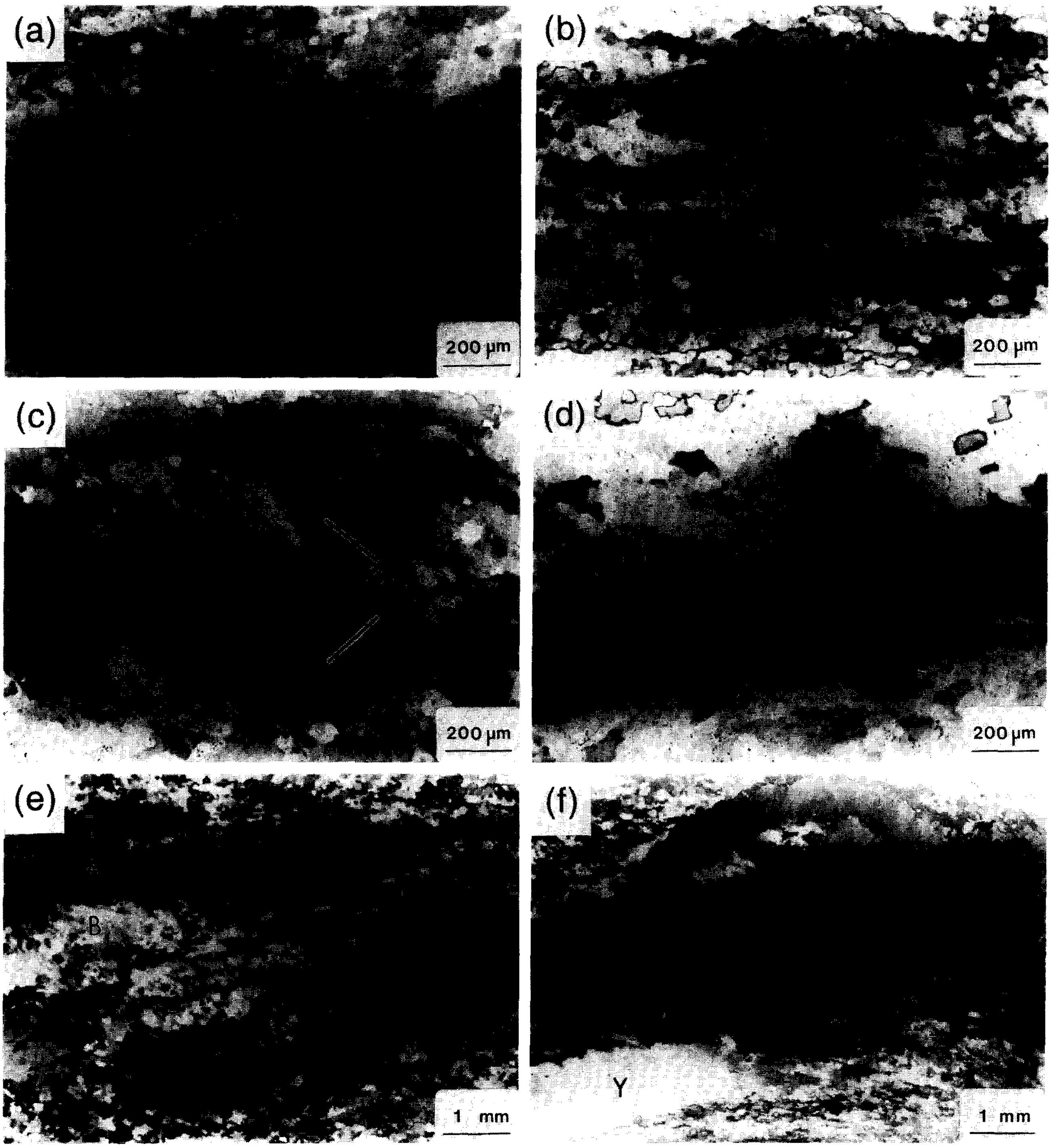


Fig. 4. Photomicrographs of quartz vein mylonites, taken with crossed polars and gypsum plate inserted. All sections cut at right angles to S_A, which is subhorizontal in all photos. (a, c, e, and f) viewed parallel to stretching lineation, with direction of plunge to left, (b) and (d) are perpendicular to stretching lineation. (a) Dds: Strong asymmetry is evident parallel to lineation. Oblique angle between S_A (horizontal), and grain shape preferred orientation S_B (marked with bar) indicates top-to-the-right shear (top-to-the-southeast, reverse + dextral motion in geographic coordinates). (b) Dds: No asymmetry is evident perpendicular to lineation. Note sutured grain boundaries and subgrains and new grains along larger grain boundaries. (c) Dss: S_A is indicated by subhorizontal dark, micaceous fuchsite grains. a-, b- and c-domains are labeled; direction of elongation of quartz grains in adjacent a- and b-domains marked by bars. See text for further details. (d) Dss: At right angles to the stretching lineation, boundaries between domains are irregular and asymmetry within domains is less pronounced than parallel to lineation. (e) Dss: Quartz domains are laterally discontinuous, commonly exhibiting lens-like morphology elongate parallel S_A. Representative a-, b- and c-domains are labeled. (f) Tails containing subgrains and recrystallized grains extend from relict quartz grains (B and Y) that exhibit differently oriented c-axes, forming domains with c-axis orientations similar to the original grains. An a-domain is forming from the blue grain B and a b-domain is forming from the yellow grain Y.

lineation, this laminar fabric is less evident; mineral grains are more equant in shape, and S_A is poorly defined (Fig. 3d). S_A in mylonitized quartz veins, present in both structural domains, is defined by trails of epidote group minerals and fuchsite and by boundaries between domains of grains of differing crystallographic orientation (Fig. 4).

The mineral lineation is defined predominantly by the preferred dimensional alignment of amphiboles and epidote group minerals in greenschists (Fig. 3). Quartz is typically 'flattened' within S_A in greenschists and within S_B in quartz vein mylonites, and is elongate parallel to the mineral lineation. Local boudinage of clots of epidote group minerals is evident along the mineral lineation, where quartz fills in the gaps between boudins, indicating that the lineation is a true stretching lineation (Fig. 3c). Quartz vein mylonites exhibit the maximum fabric asymmetry in sections cut parallel to the mineral lineation and at right angles to the foliation in both Dds, and Dss (compare Figs. 4a & b, as well as 4c & d), indicating that the stretching lineation is approximately normal to the bulk vorticity vector.

Main foliation and mineral lineation

The variation in orientation of S_A and the mineral lineation across the two structural domains in Marble Cove is shown in Fig. 2. There clearly is not a smooth variation from steeply to shallowly plunging lineations from south to north across Marble Cove. The lineations exhibiting the shallowest plunges are clustered in a narrow area between serpentinite bodies within the dominantly zoisite–fuchsite and tremolite–zoisite schists, defining Dss. Steeply plunging lineations are present across a wider area in the southern part of the cove, designated Dds. The boundary between the two domains is abrupt, and occurs within a lens of serpentinized ultramafic rock. Both Dss and Dds include narrow zones in which the lineations differ by at least 10° in plunge from lineations in surrounding areas. In general, however, lineations within Dss pitch less than 45° within the foliation plane, indicating dominantly strike-slip motion, and lineations within Dds pitch more than 45° within the foliation plane, recording dominantly dip-slip motion.

Foliations and lineations are only locally developed within the altered ultramafic rocks of Marble Cove. The orientations of the foliation in these rocks is highly variable, and the rocks are typically disrupted by numerous cross-cutting structures that are both brittle and ductile in character. The mean orientations of the lineation and foliation in serpentinized ultramafic rocks are shown in Fig. 2.

The lineation and poles to S_A in each structural domain are plotted in Fig. 5; equal area plots do not include measurements from the altered ultramafic rocks.

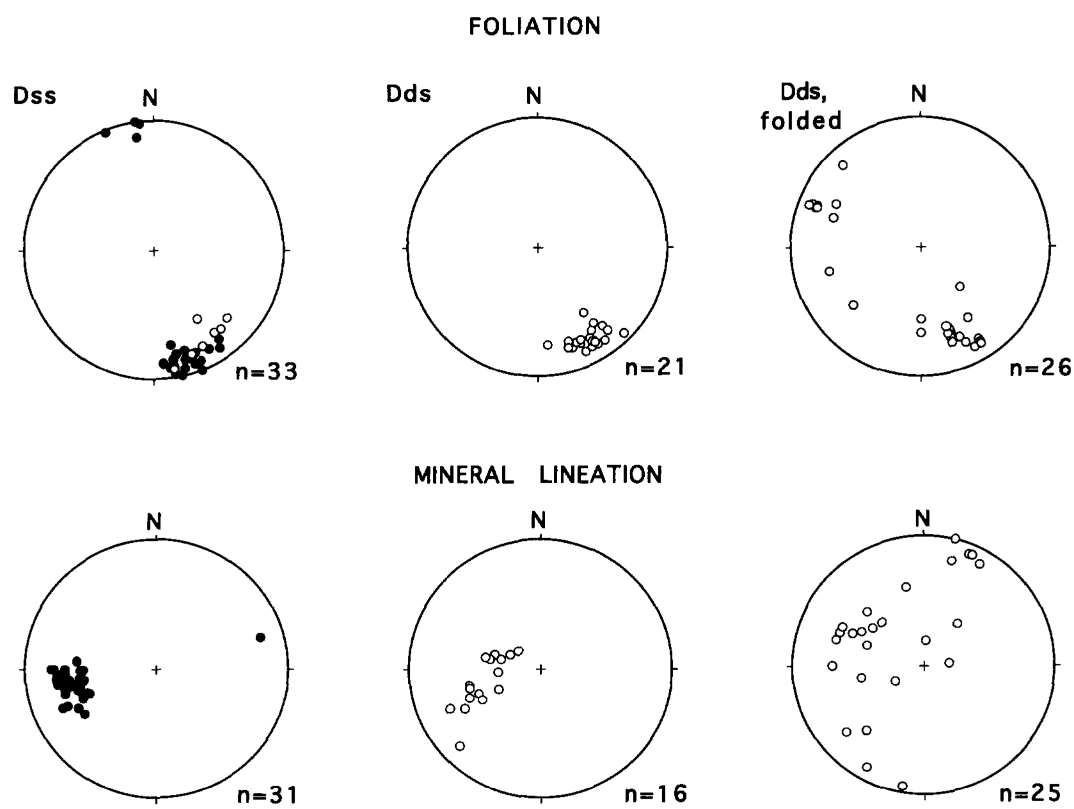


Fig. 5. Lower hemisphere, equal-area projections of mineral lineation and poles to S_A in Dss (left), Dds (center) and areas in Dds where S_A and lineation are folded (right). The open circles in the Dss foliation plot represent poles to S_A within which the lineation suggests a dominance of dip-slip movement. Note that these surfaces are similar in orientation to those of Dds. The number of data points (n) is indicated for each plot.

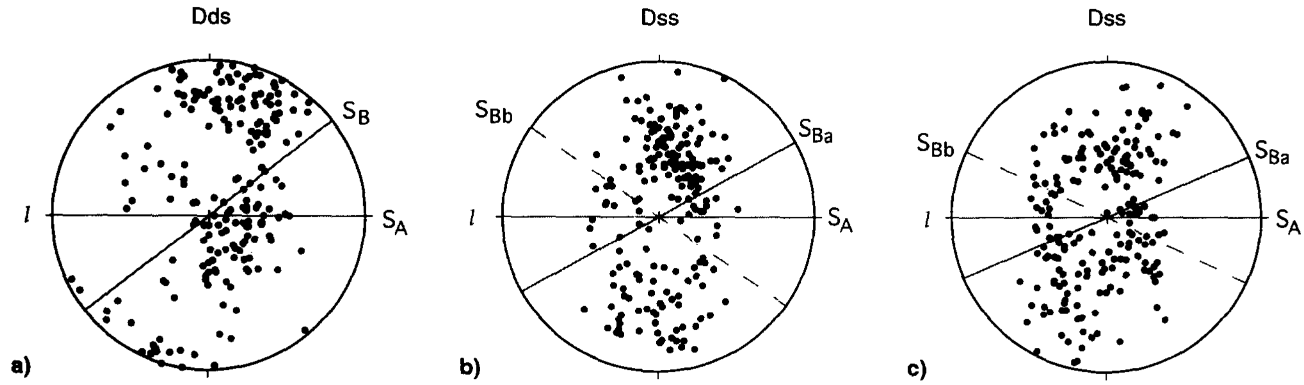


Fig. 6. Lower hemisphere, equal-area plots of *c*-axes in samples from Dds (a) and Dss (b) and (c). Each plot includes 200 points. A sample of 200 *c*-axes is sufficient to identify the maximum point concentration, to determine the area of empty space within 2%, and to determine the areas of higher point concentrations to better than 2% (J. Starkey written comm. 1990). In all plots, the trace of the main foliation, S_A, is horizontal, as marked. The down plunge direction of the stretching lineation, *l*, is to the left. The grain shape preferred orientation S_B is shown in each plot; in plots of *c*-axes for Dss, the orientation for the volumetrically dominant domains, S_{Ba}, is shown with a solid line and the orientation of the subordinate domains, S_{Bb}, is shown with a dashed line. Sense of asymmetry of *c*-axis distribution relative to main foliation S_A and down-plunge direction of lineation is dextral in each case, consistent with the dominant grain shape preferred orientation. (a) *c*-axes in Dds form a single, strongly asymmetric girdle. (b) Dss: Single girdle exhibiting less pronounced asymmetry than is evident in Dds. (c) Dss: Crossed circle girdle with one arm much less pronounced than the other.

Although there is overlap in the orientation of S_A between domains, the mean orientation is slightly different in each domain. S_A typically dips about 10° more steeply in Dss than in Dds, and has a slightly more easterly strike. Within Dss, the foliation surfaces most similar in orientation to those in Dds include lineations that pitch more than 45° in the plane of the foliation. In the field, no cross-cutting relationship between the differently oriented foliation surfaces is evident. Note also that there is significantly less variation in the orientation of the stretching lineation in Dss than in Dds. The orientations of the lineation within Dss form a tight cluster on the equal-area plot, while those within Dds are somewhat scattered. This may reflect the presence of unrecognized folds in foliation (discussed in later sections), the magnitude of strain experienced by each domain, and/or strain partitioning within Dds. Bursnall (1975) first recognized evidence of high strain within Dss. Observations that suggest that Dss accommodated more strain than Dds include: (1) with the exception of quartz vein mylonites, the grain size of rocks in Dss is generally smaller than the grain size of rocks in Dds (which may also be related to strain rate), (2) both foliation and lineation are better developed in Dss and the variation in orientation of structures in Dss is less than that in Dds (cf. White & Flagler 1992, Goodwin & Wenk 1995), and (3) shear bands are well developed in Dss but absent in Dds.

Sense-of-shear indicators

Sense-of-shear indicators have been evaluated within both the greenschists and quartz vein mylonites found in both Dss and Dds. To evaluate quartz vein mylonites, we consider (1) the observed sense of obliquity between the main mylonitic foliation, S_A, and the preferred dimensional alignment of dynamically recrystallized quartz grains, S_B, and (2) the asymmetry of quartz *c*-axis distributions (see reviews by Simpson & Schmid 1983 and Price 1985). All lineation-parallel photomicrographs and plots of *c*-axis distributions (Figs. 3, 4 and 6) are oriented with the trace of S_A horizontal and the down-plunge direction of the lineation to the left in order to facilitate comparison with the map in Fig. 2. That is all of these figures are viewed to the north or northeast.

Microstructures within all quartz vein mylonites in Marble Cove are similar to those created through experimental deformation of quartzite in dislocation creep regime 3, the highest temperature deformation regime documented by Hirth & Tullis (1992). Within regime 3, strain is accommodated by dislocation glide and recovery is accommodated by dislocation climb and both rotation and migration recrystallization. Dynamic recrystallization is evidenced by extensive subgrain formation (rotation recrystallization) and strongly sutured grain boundaries (migration recrystallization) (Fig. 4). As previously discussed, mineral assemblages in adjacent rocks record greenschist facies metamorphism, which indicates relatively low temperatures during deformation. However, the extreme alteration to dominantly hydrous phases experienced by the mafic igneous protoliths to these metamorphic tectonites, and the ubiquitous presence of veins, suggests a significant influx of water before or during deformation. The high water activity may have facilitated grain boundary migration. In addition, regime deformation mechanisms active during high temperature, high strain rate experiments could be active at lower temperature conditions under the lower strain rates typical of naturally deformed rocks.

Rocks within Dds are typically schistose, exhibiting orthorhombic fabric symmetry. Both macroscopic and microscopic sense-of-shear indicators are absent. Quartz vein mylonites within Dds, however, exhibit distinctly monoclinic symmetry. S_B is well developed, and exhibits a consistent obliquity with respect to S_A in all of the samples examined for this study (Fig. 4a). This obliquity indicates top-to-the-southeast reverse motion with a dextral component. This is borne out by the strongly asymmetric distribution of quartz c-axes relative to S_A (Fig. 6a).

Dss differs from Dds in that it is a recognizable high strain zone; the main foliation and stretching lineation are clearly visible from the sea as the cove is approached by boat. Macroscopically and microscopically visible shear bands locally developed in Dss, particularly in zoisite–fuchsite mylonites, record dextral strike-slip motion with a component of reverse (NW over SE) shear (Fig. 3a). Locally developed conjugate shear bands are also evident in thin section (Fig. 3b), suggesting that the pure shear component of the deformation was relatively strong with respect to the simple shear component during development of shear bands (Williams & Price 1990). The microstructures of quartz vein mylonites are also complex (Figs. 4c–f) and are distinctly different from those in Dds. A domainal microstructure is defined by lenses and laminae of quartz grains and subgrains with similar crystallographic orientations, as indicated by insertion of a gypsum plate (Figs. 4c & e). These domains are bounded by S_A, and the shape preferred orientation of quartz grains and subgrains, S_B, varies between domains. The shape preferred orientations, and corresponding crystallographic preferred orientations, are similar to those described by Celma (1982) in the Cap de Creus quartz mylonites of Spain. Celma defined three domains in thin sections cut at right angles to S_A and parallel to the stretching lineation (Figs. 4c, e & f): (1) **A-domains** are characterized by a blue color with insertion of the gypsum plate, and exhibit a positive angle of inclination between S_B and S_A. These domains are volumetrically dominant in Marble Cove. (2) **B-domains** are characterized by a yellow color with insertion of the gypsum plate, and exhibit a negative angle of inclination of S_B with respect to S_A. (3) **C-domains** exhibit a red to violet color with insertion of a gypsum plate and show no shape preferred orientation; C-domains are the least volumetrically important in Marble Cove.

The sense of obliquity of S_B with respect to S_A in the dominant A-domains (Figs. 4c & e) indicates dextral shearing with a component of reverse (top-to-the-southeast) shearing in Dss. This is supported by the asymmetry of quartz c-axes with respect to S_A (Figs. 6b & c). Measurements of c-axes in quartz mylonites from Dss were made in traverses across S_A in order to obtain statistically representative samples of all three domains. The resulting pole figures exhibit a less pronounced asymmetry than is present in Dds (compare Fig. 6a with Figs. 6b & c); in fact, Fig. 6(c) is nearly a crossed-circle girdle with one well developed girdle and a second weak girdle. Gapais & Cobbold (1987) have demonstrated that single c-axis girdles can develop from crossed-circle girdles with increasing strain in simple shear. That is, the difference between c-axis orientations in Dss and Dds might, at least in part, be due to differences in strain. We suggest, however, that higher strain in quartz vein mylonites in Dds relative to those in Dss is not enough to account for the difference in character of c-axis distributions in the two domains. In particular, we note one c-axis distribution (Fig. 6b), that is a single girdle exhibiting a slight dextral asymmetry, which could record components of both dextral simple shear and pure shear. This interpretation is consistent with our observation that adjacent greenschists record components of both pure and simple shear.

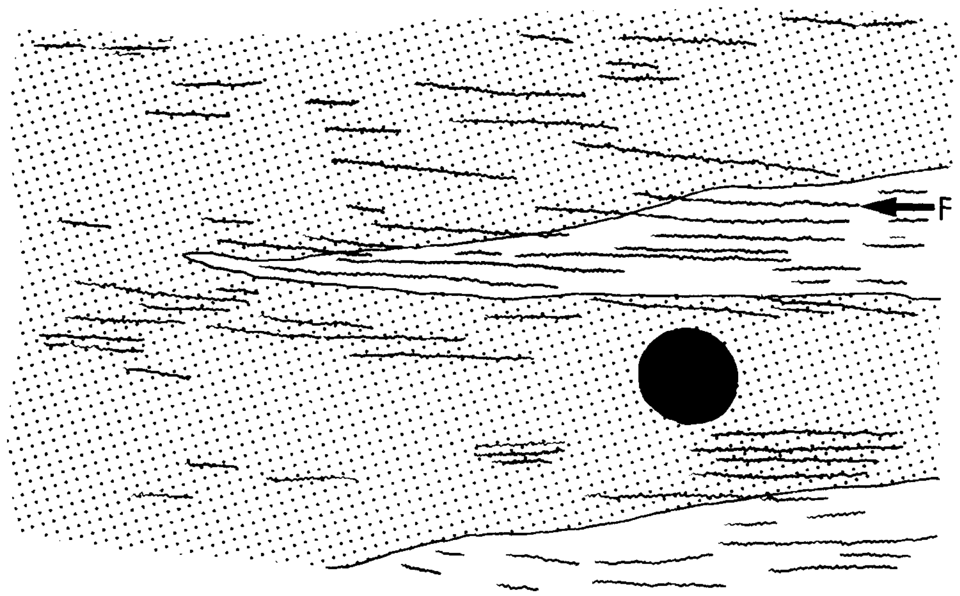
Eisbacher (1970) and Celma (1982) recorded similar domainal microstructures in quartzite mylonites. They suggested that this microstructure partly reflects the orientations of coarse host grains from which the fine-grained mylonite was derived. Relict quartz grains locally preserved within the deformed veins in Marble Cove are relatively coarse-grained. The variation in orientation of the original grains does appear to be reflected in domains of deformed grains, new grains, and subgrains (Fig. 4f). However, we also suggest that this domainal microstructure reflects deformation through a combination of pure shear and simple shear, which is also consistent with the findings of Celma (1982). We therefore conclude that the quartz vein mylonites and the rocks that host them record both (1) dextral strike-slip shear with a reverse component of movement, and (2) a pure shear component of deformation.

Folds in compositional layering

Two types of folds are found in Marble Cove: folds in compositional layering and folds in foliation. The first is discussed below; the second is discussed in the following section.

The hinges of rare folds in compositional layering are roughly parallel to the stretching lineation in the rocks in which they occur (Fig. 2). This is also evident in thin section, where small-scale fold hinges may be seen in sections cut at right angles to both the foliation and the lineation. Within Dss, the foliation is axial planar with respect to the folds. Within Dds however, the foliation is oblique to the folds (Fig. 7). In both Dss and Dds, there is no clear evidence, either macroscopic or microscopic, for any fabric older than this foliation. Thus it is reasonable to interpret the folds and foliations in each domain as having formed during a progressive deformation.

Folds such as those found in Dss, with fold hinges parallel to the stretching lineation and axial planar cleavage, are common in high strain zones (e.g., Sanderson 1973, Escher & Watterson 1974, Ratliff *et al.* 1988, White & Flagler 1992). Oblique-cleavage folds, such as those in Dds, have been described from shear zones in which the strain path is non-coaxial (Lafrance & Williams 1992). We interpret the relationships described here to indicate that the folds and cleavage formed succesively during a progressive deformation, with the foliation overprinting the folds. We suspect that this overprinting relationship is no longer distinguishable within Dss because it has experienced a higher degree of strain. With increasing strain, we would expect transposition of layering into the plane of the foliation, a



b

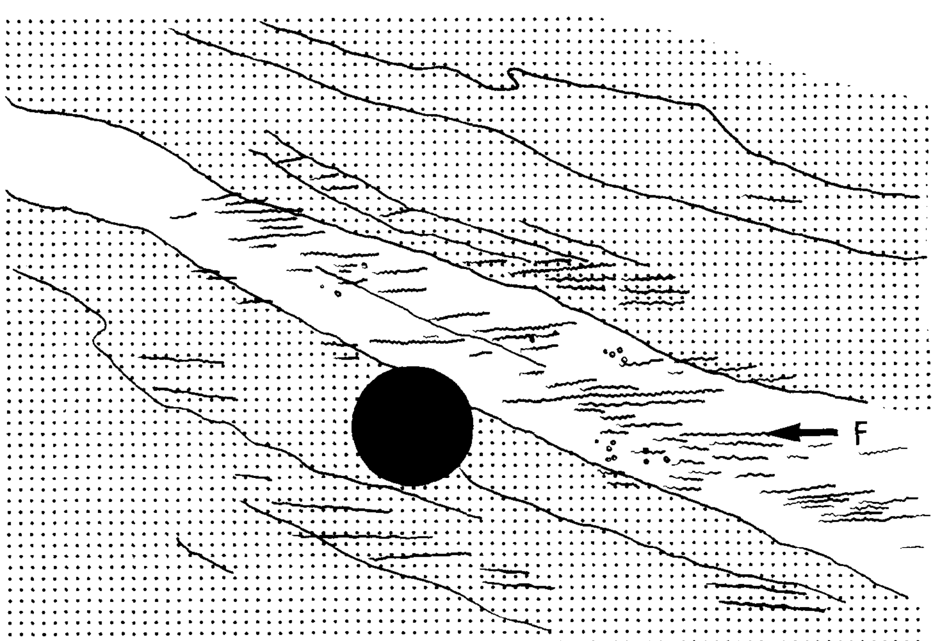


Fig. 7. Sketch from photograph of relationship between foliation in Dds (F) and fold in compositional layers, shown (a) perpendicular to fold hinge and stretching lineation and (b) parallel to fold hinge and stretching lineation. The lens cap is 5 cm in diameter.

decreasing the angle of obliquity between the foliation and the axial planes of folds in layering. At high strains, the obliquity would no longer be recognizable.

Crenulation lineation and folds in foliation

A crenulation lineation and meter-scale folds in foliation affect the rocks of both structural domains in Marble Cove (Figs. 8–10). Unlike all of the other structures described above, these features are similarly oriented in both domains, with crenulations roughly parallel to folds in foliation. The slight difference in attitude between crenulations and folds in Dss versus those in Dds is due to the differences in orientation of the foliation being folded (Fig. 5). That is, the foliation exhibits a more easterly strike and a steeper dip in Pss than in Dds. Both the small (crenulation) and large-scale folds appear to post-date peak metamorphism. The crenulation lineation deforms the minerals that define the main foliation and stretching lineation, and no mineral growth accompanies the folding (Fig. 3e).

The crenulation lineation is pervasive within Dds, but is only locally developed in Dss. Similarly, folds in foliation are common in Dds (particularly in one area; Fig. 9), but only one fold in foliation has been found in Dss. In Dds, the crenulation lineation is oriented at right

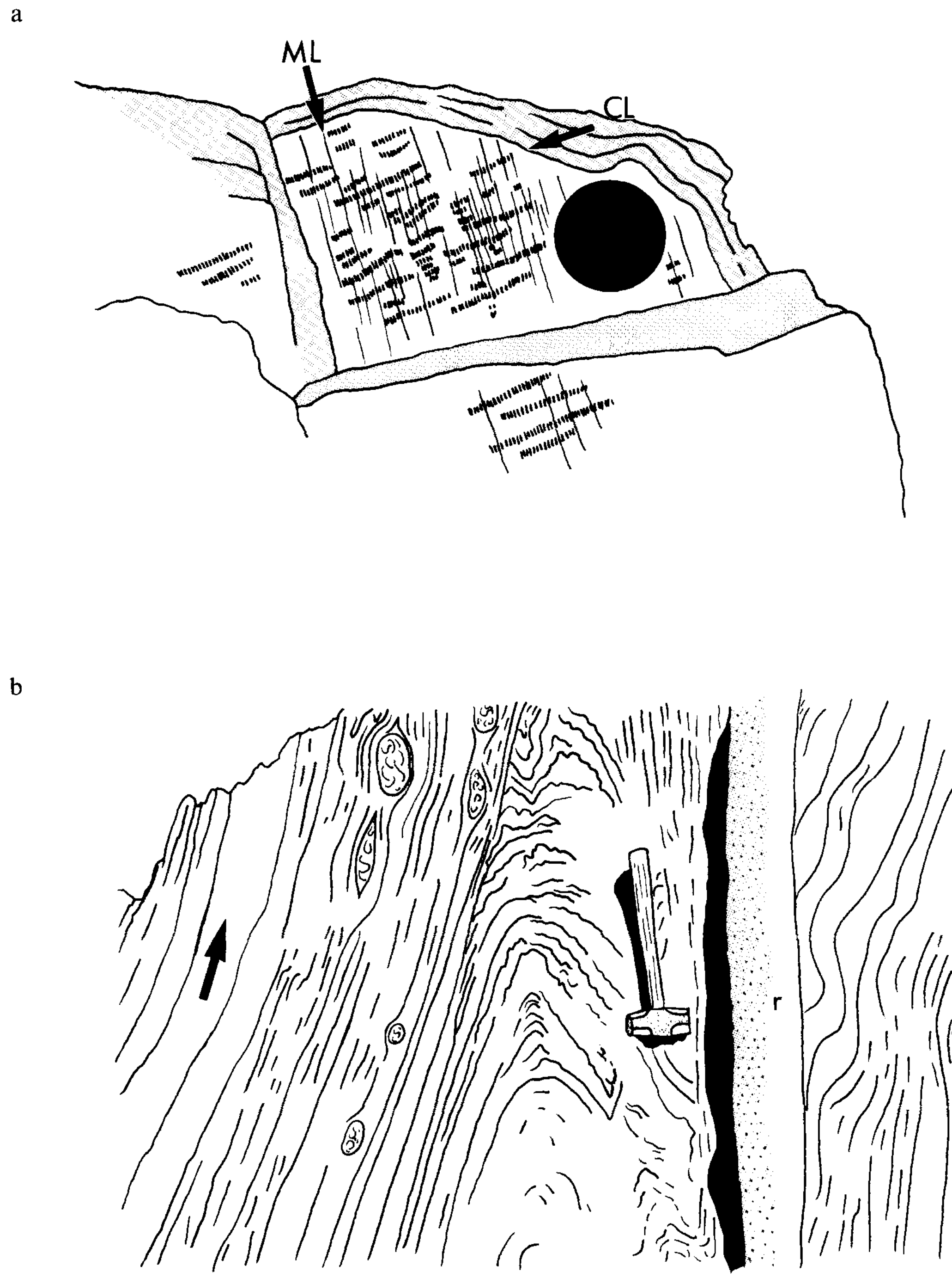


Fig. 8. Sketches (from photographs) of (a) foliation surface, showing relative orientation of mineral lineation (ML) and crenulation lineation (CL) in Dds. The lens cap is 5 cm in diameter. (b) Fold in foliation, showing typical extensive shearing of southeast limb.

angles to the mineral lineation, creating a distinctive tweed pattern on the foliation planes (Fig. 8a). The crenulation lineation is oblique to the mineral lineation in Dss. Crenulations are parasitic with respect to the larger folds in foliation, and are parallel to the hinges of adjacent folds. Crenulations developed in the absence of macroscopic folds locally exhibit an asymmetry consistent with top-to-the-southeast shear (that is they are z-folds when viewed down plunge). Folds in foliation are typically symmetrical; of the hinges measured (Fig. 10), 5 exhibit dextral (top-to-the-southeast) asymmetry, one exhibits sinistral (top-to-the-northwest) asymmetry, and the remainder are neutral. Folds may be open to tight; the more open folds generally exhibit greater variation in the orientation of hinges, which may be curvilinear. The tight folds — and some of the open folds — have strongly sheared SE limbs (Fig. 8b), suggesting top-to-the-southeast motion. Axial planes to folds and rare axial planar foliations are similar in orientation to the main foliation in Dds.

The consistency of sense-of-shear between asymmetric folds in foliation and quartz vein mylonites in Dds, the

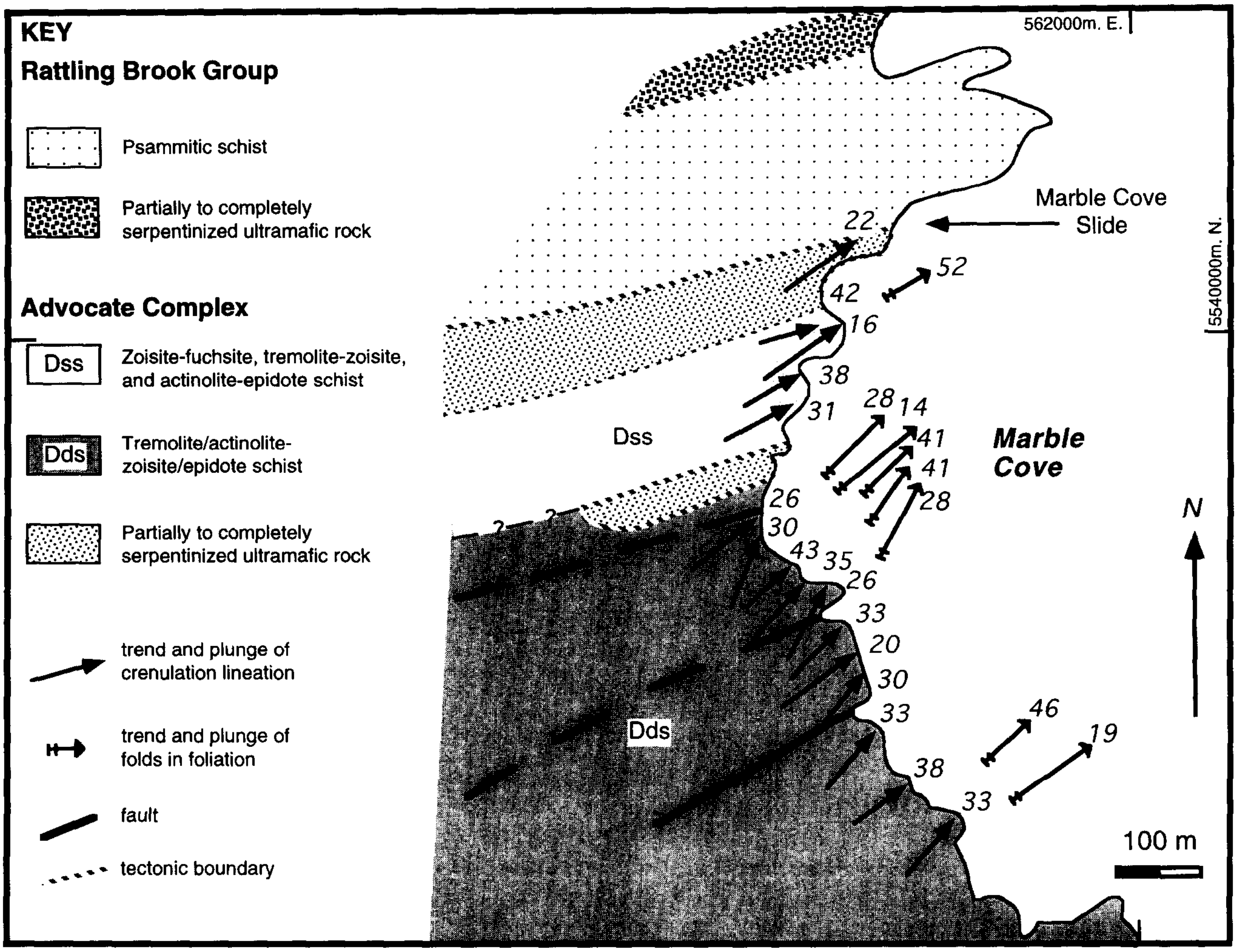


Fig. 9. Detailed map of Marble Cove showing the variation in orientation of crenulation lineation and hinges of folds in foliation within Dds and Dss. Linear features are scaled by plunge, as in Fig. 2.

regular orientation of fold hinges at right angles to the stretching lineation in Dds, and the fact that folds are preferentially developed in Dds, all suggest that the earlier and later structures belong to the same deformation episode. Similar development of multiple generations of structures during a single deformation event has been demonstrated by Hudleston *et al.* (1988) and White & Flagler (1992). Microstructural evidence suggests that this continuous deformation took place as the area cooled, as metamorphic minerals did not grow during the formation of crenulation lineations.

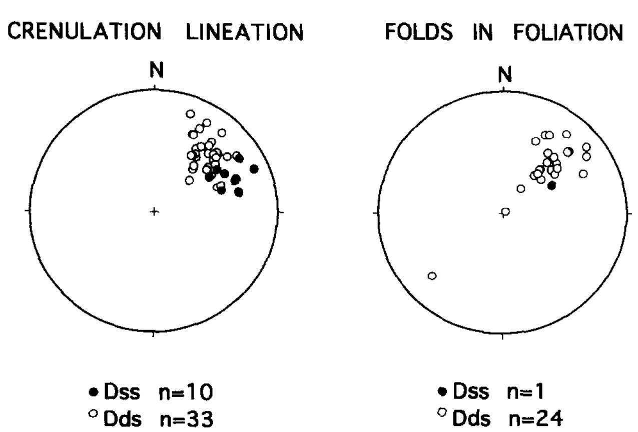


Fig. 10. Lower hemisphere, equal-area plots of crenulation lineation and hinges of folds in foliation in both Dss and Dds. Multiple measurements were made of curvilinear hinges, resulting in the spread in orientations seen in the plot.

As the youngest structures present throughout Marble Cove, the crenulation lineation and folds in foliation provide us with a time line. By the time they formed, all of the distinctive structures of both structural domains had already developed.

DISCUSSION

Partitioning of pure shear and simple shear in Dss and Dds

Throughout Dss, both within and outside quartz vein mylonites, there is evidence for components of both pure and simple shear, as detailed in earlier sections. In contrast, these components appear to be largely partitioned within Dds. Fabrics within quartz vein mylonites in Dds exhibit monoclinic symmetry, suggesting noncoaxial deformation. There is no evidence for a significant pure shear component. The tremolite–zoisite and actinolite–epidote schists in Dds, however, typically exhibit orthorhombic symmetry; the fabric is symmetrical about the lineation. This higher symmetry suggests a strong component of pure shear, with extension parallel to the stretching lineation. We suggest that this partitioning of components of pure and simple shear in Dds is a three-dimensional equivalent of the stretching faults described from analog models by Means (1989, 1990), though we note that this proposed model leads to strain compatibility problems. Stretching faults are distinguished by stretching within fault blocks that occurs parallel to the slip direction of the fault; slip is confined to the fault plane. In Marble Cove, non-coaxial shear in Dds appears to be largely confined to the narrow quartz veins (younger folds in foliation are an important exception); stretching in the rock surrounding the quartz vein mylonites occurs parallel to extension in the mylonites.

This partitioning of simple shear into quartz vein mylonites in Dds may also explain the fine grain size of the quartz grains relative to those in quartz vein mylonites in Dss (Fig. 4). In Dss, both the quartz veins and the surrounding rocks are accommodating both pure and simple shear. By accommodating the majority of simple shear in Dds, quartz vein mylonites may experience a faster strain rate.

Relative timing of formation of structures in Dss and Dds

In the following paragraphs, we discuss evidence for the relative timing of deformation in Dss and Dds. We evaluate three possibilities: (1) that the structures in Dds and Dss formed sequentially, (2) that the structures in Dds and Dss formed simultaneously and Dss represents an area that was rotated subsequent to ductile deformation, and (3) that the structures in Dds and Dss formed during a single deformation event.

The features in Dds and Dss formed at different times. Hansen (1989) documented overprinting of structures formed during dip-slip shear by dextral strike-slip shear zones during terrane accretion in a transpressional tectonic regime within the Teslin suture zone in the Yukon. The complexities that Hansen (1989) documents might not be expected in an area like Marble Cove, where the deformational event being investigated postdates accretion. Further, overprinting relationships are absent from the rocks in Marble Cove. However, the mean orientation of S_A in Dds is slightly counterclockwise of that in Dss (Fig. 7), and could be interpreted as having been rotated into Dss during dextral shear. Since Dss exhibits evidence of greater strain than Dds, we can conceive of earlier structures within the zone being erased. The latter could explain the lack of overprinting relationships between structures in each domain.

Factors that argue against this model of dip-slip followed by strike-slip deformation include: (1) deformation within both domains took place under the same metamorphic conditions. While this scenario is possible, it is unlikely. (2) All sense-of-shear indicators across Marble cove record variable components of reverse and dextral shear. This also seems fortuitous if they record sequential events.

Foliation in both domains was subsequently affected by folding and formation of a crenulation lineation. This model therefore requires three deformational episodes: (a) reverse shear with a dextral component to form the structures in Dds, (b) dextral shear with a reverse component to form the features in Dss, and (c) reverse shear with a dextral component to develop the crenulation lineation and associated folds in foliation.

Structures in Dds and Dss were once parallel, but Dss was rotated (rigid body rotation) subsequent to the main deformational event. Rotation could be accommodated by the serpentinite bodies that bracket Dss (Fig. 2).

The main factor that argues against this model is that Dds and Dss do not exhibit structures that are identical except in orientation. Rocks in Dss exhibit a better developed and more regularly oriented foliation and lineation, development of shear band foliation(s), and finer grain size than rocks in Dds. More than rigid-body rotation is required to explain these differences. Again, this model requires multiple deformation episodes: initial formation of structures, rigid-body rotation, then formation of the crenulation lineation.

Dss and Dds formed simultaneously within a single deformational event. Dss and Dds accommodated dextral strike-slip and reverse shear as well as pure shear within a transpressional regime. The crenulation lineation and associated folds in foliation represent the waning stages of deformation, accommodating the final phase of shortening. As mineral growth slowed or ceased prior to formation of these folds, it suggests that the area cooled during this progressive event. We prefer this model, as it is consistent with the lack of overprinting relationships, with mineral assemblages throughout Marble Cove that suggest deformation under greenschist facies conditions, and with sense-of-shear indicators across Marble Cove that record variable components of reverse and dextral shear. This is the simplest model, requiring a single progressive deformation event.

Several other workers have interpreted variably plunging stretching lineations and associated structures as having formed simultaneously in transpressional shear zones in orogenic belts in Canada (Caron & Williams 1988a & b, Holdsworth 1989, 1991, Robin & Cruden 1994), the United States (Hudleston *et al.* 1988), Greenland (Holdsworth & Strachan 1991), and Sweden (Robin & Cruden 1994). Garnett & Brown (1973) described a zone of variably pitching lineations in southern New Brunswick, in which there are gradations between highly strained rock with a down-dip stretching lineation, less strained rock with a shallowly plunging stretching lineation, and unstrained granitoid. They suggested that the deformed region is essentially a reverse shear zone, and that the area of shallowly plunging lineations is due to the buttressing effect of the

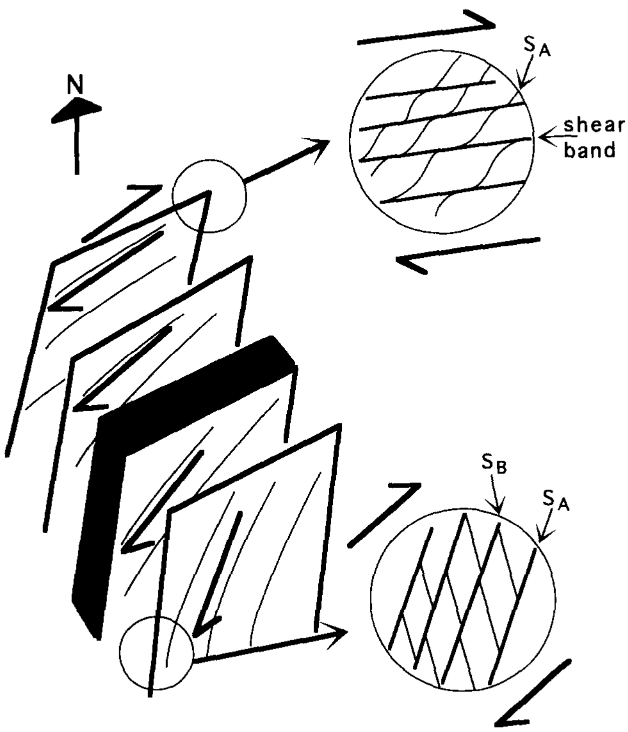


Fig. 11. Schematic illustration of relationships within sheared rocks of the Advocate Complex, Marble Cove. Dark layer represents serpentinized ultramafic rock between Dss and Dds. Within Dss, a shallowly plunging stretching lineation, shear bands (inset), and both obliquity of elongate recrystallized grains and subgrains and asymmetry of *c*-axis distributions in quartz vein mylonites record dominantly dextral strike-slip shear. Within Dds, steeply plunging stretching lineations and both obliquity of elongate grains (inset) and *c*-axes in quartz vein mylonites record dominantly reverse shear.

granitoid and does not reflect movement within the shear zone as a whole.

The mathematical models of Sanderson & Marchini (1984), Fossen & Tikoff (1993) and Robin & Cruden (1994) suggest a number of structures that might be found within transpression zones. Those most similar to the structures in Marble Cove are found in the Proterozoic Mylonite Zone in Sweden, described and modeled by Robin & Cruden (1994). The Mylonite Zone includes a variably oriented foliation and lineation which are asymmetrically distributed across the shear zone. As in Marble Cove, the bulk vorticity vector is oriented at approximately right angles to the mineral lineation. The Mylonite Zone accommodated sinistral and reverse motion, and has been modeled as an oblique transpression zone with a small pure shear component (Robin & Cruden 1994). An important difference between the Mylonite Zone and Marble Cove is that the dominantly strike-slip zone occurs adjacent to the hanging wall in Marble Cove, opposite to the orientation of the analogous area of the Mylonite Zone.

Robin & Cruden (1994) also demonstrated that with equal components of pure shear and simple shear ('press' component = 'trans' component) the bulk vorticity vector will locally depart significantly from a position at right angles to the stretching lineation, and that the lineations will plunge steeply throughout the shear zone. The latter observation is also indicated by Sanderson & Marchini (1984) and Fossen & Tikoff (1993). When applied to Marble Cove, these results suggest that the 'press' component of deformation is small relative to the 'trans' component, except perhaps within the greenschists of Dds. Within Dss, where we see evidence for both pure and simple shear, the pure shear component must be significantly less than the simple shear component.

Considering our own observations and the work of previous authors, we propose a model by which the structures in Marble Cove formed (Fig. 11). We first note that the psammitic schists of the Rattling Brook Group north of the Marble Cove slide (cf. Figs. 2 and 9) did not develop a pervasive fabric that could be correlated with fabrics developed in the rocks of the Advocate Complex to the south. Second, rocks of the Rattling Brook Group are separated from those of the Advocate Complex by serpentinized ultramafic rock. Third, rocks within Dss are separated from rocks within Dds by a string of serpentinized ultramafic bodies (Figs. 2 and 9; Bursnall & Hibbard 1980, 1983). Further, rocks within Dss can be distinguished lithologically from those within Dds in that tremolite–zoisite mineral assemblages are dominant over actinolite–epidote bearing rocks, and zoisite–fuchsite rocks are present. There are therefore three distinct lithologic packages — rocks of the Rattling Brook Group, of Dss, and of Dds — separated by serpentinized ultramafic rocks. We suggest that there was a lithologic control on strain partitioning in Marble Cove. Deformation was localized within Dss and Dds, while the more competent rocks of the Rattling Brook group accommodated little of the strain. Strike-parallel motions were mainly localized within Dss, while shortening was accomplished largely through deformation within Dds. Misfit between these three lithologically distinct and differently deforming rock packages was accommodated at least in part within and probably along the margins of the serpentinized ultramafic rock bodies. We suggest that the differences between the structures developed in Marble Cove and those described in the Mylonite Zone and modeled by Robin & Cruden (1994) may be in part due to the presence of these lithologically distinct packages. While the hanging-wall was relatively rigid, as their model requires, the footwall (rocks of the Advocate Complex) was not. Further, the contacts with the serpentinite bodies may have allowed some degree of slip, which is not accounted for in the Robin & Cruden model.

CONCLUSIONS

Deformation in Marble Cove was partitioned between a broad domain exhibiting a steeply plunging stretching lineation and a narrow high strain domain containing a shallowly plunging stretching lineation. Both domains include evidence for pure shear and simple shear (dominantly dextral in Dss and reverse in Dds), though pure and simple shear are largely partitioned between schists and quartz vein mylonites in Dds. Coincidence of deformation between the two domains is indicated by the lack of overprinting relationships, identical mineral assemblages within both domains, and late-stage folds in foliation that affect both Dss and Dds. We therefore conclude that they formed simultaneously in a transpressive regime. The rheologically distinct serpentinized ultramafic bodies that bound Dss to the north and south probably helped alleviate problems in strain compatibility between Dss and surrounding rocks.

The example of Marble Cove indicates that deformation can be partitioned within ductile transpression zones; brittle flower structures need not extend into homogeneous, oblique-slip shear zones at depth. In addition, this study indicates that structures that differ significantly in orientation and style may still be genetically related. A further point important to mapping in general, and particularly relevant to the Appalachian orogen, is this: because the strike-slip domain is very narrow relative to the dip-slip domain, it is less likely to outcrop and may be overlooked. Structures developed in transpression may therefore be misinterpreted as reflecting dominantly dip-slip deformation regimes.

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REFERENCES

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

Borradaile, G., Sarvas, P., Dutka, R., Stewart, R. & Stubley, M. 1988. Transpression in slates along the margin of an Archean gneiss belt, northern Ontario—magnetic fabrics and petrofabrics. *Can. J. Earth Sci.* **25**, 1069–1077.

Brun, J.-P. & Burg, J.-P. 1982. Combined thrusting and wrenching in the Ibero–Armorican arc: a corner effect during continental collision. *Earth Planet. Sci. Lett.* **61**, 319–332.

Bursnall, J. T. 1975. Stratigraphy, structure and metamorphism west of Baie Verte, Burlington Peninsula, Newfoundland. Unpublished Ph.D. thesis. Cambridge University.

Bursnall, J. T. 1979. Geology of part of the Fleur de Lys map area (12I/1E0) Newfoundland. In: *Report of Activities for 1978* (edited by R. V. Gibbons) *Newfoundland Department of Mines and Energy, Mineral Development Division Report* **79-1**, 68–74.

Bursnall, J. T. & DeWitt, M. J. 1975. Timing and development of the orthotectonic zone in the Appalachian orogen of northwest Newfoundland. *Can. J. Earth Sci.* **12**, 1712–1722.

Bursnall, J. T. & Hibbard, J. 1980. Fleur de Lys (12I/1) and Horse Islands (2L/4) map areas. Newfoundland Department of Mines and Energy, Mineral Development Division Map 80–30.

Caron, A. & Williams, P. F. 1988a. The multistage development of the Dover Fault Zone in northeastern Newfoundland: the late stages. *GAC-MAC-SEG Program with Abstracts* **13**, A17.

Caron, A. & Williams, P. F. 1988b. The kinematic indicators of the Love Cove Group in Northeastern Newfoundland. *GAC-MAC-SEG Program with Abstracts* **13**, A17.

Cawood, P. A., Dunning, G. R., Lux, D. & van Gool, J. A. M. 1994. Timing of peak metamorphism and deformation along the Appalachian margin of Laurentia in Newfoundland: Silurian, not Ordovician. *Geology* **22**, 399–402.

Celma, A. G. 1982. Domainal and fabric heterogeneities in the Cap De Creus quartz mylonites. *J. Struct. Geol.* **4**, 443–455.

D'Lemos, R. S., Brown, M. & Strachan, R. A. 1992. Granite magma generation, ascent and emplacement within a transpressional orogen. *J. Geol. Soc. Lond.* **149**, 487–490.

Dewey, J. F., Kennedy, M. J. & Kidd, W. S. F. 1983. A geotraverse through the Appalachians of northern Newfoundland. *Profiles of Orogenic Belts Geodynamics Series* **10**, 205–241.

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

Escher, A. & Watterson, J. 1974. Stretching fabrics, folds, and crustal shortening. *Tectonophysics* **22**, 223–231.

Fossen, H. & Tikoff, B. 1993. The deformation matrix for simultaneous simple shearing, pure shearing and volume change, and its application to transpression–transtension tectonics. *J. Struct. Geol.* **15**, 413–422.

Gapais, D. & Cobbold, P. R. 1987. Slip system domains. 2. Kinematic aspects of fabric development in polycrystalline aggregates. *Tectonophysics* **138**, 289–309.

Gapais, D. & White, S. H. 1982. Ductile shear bands in a naturally deformed quartzite. *Textures and Microstructures* **5**, 1–17.

Garnett, J. A. & Brown, R. L. 1973. Fabric variation in the Lubec-Belleisle Zone of southern New Brunswick. *Can. J. Earth Sci.* **10**, 1591–1599.

Goodwin, L. B. & Wenk, H.-R. 1995. Development of phyllonite from granodiorite: Mechanisms of grain-size reduction in the Santa Rosa mylonite zone, California. *J. Struct. Geol.* **17**, 689–707.

Goodwin, L. B. & Williams, P. F. 1989. Foliation development during strike-slip movement around the Baie Verte Flexure, Newfoundland. *EOS* **70**, 1311.

Goodwin, L. B. & Williams, P. F. 1990. Strike-slip motion along the Baie Verte Line, Newfoundland. *Atlantic Geology* **26**, 170.

Hansen, V. L. 1989. Structural and kinematic evolution of the Teslin suture zone, Yukon: record of an ancient transpressional margin. *J. Struct. Geol.* **11**, 717–733.

Harland, W. B. 1971. Tectonic transpression in Caledonian Spitsbergen. *Geol. Mag.* **108**, 27–42.

Hibbard, J. 1982. Significance of the Baie Verte flexure, Newfoundland. *Bull. geol. Soc. Am.* **93**, 790–797.

Hibbard, J. 1983. Geology of the Baie Verte Peninsula, Newfoundland. Mineral Development Division, Department of Mines and Energy, Government of Newfoundland and Labrador Memoir 2.

Hirth, G. & Tullis, J. 1992. Dislocation creep regimes in quartz aggregates. *J. Struct. Geol.* **14**, 145–160.

Holdsworth, R. E. 1989. The Start-Perranporth Line: a Devonian terrane boundary in the Variscan orogen of southwest England? *J. Geol. Soc. Lond.* **146**, 419–421.

Holdsworth, R. E. 1991. The geology and structure of the Gander–Avalon boundary zone in northeastern Newfoundland. *Current Research, Newfoundland Department of Mines and Energy, Geological Survey Branch Report* **91-1**, 109–126.

Holdsworth, R. E. & Strachan, R. A. 1991. Interlinked system of ductile strike-slip and thrusting formed by Caledonian sinistral transpression in northeastern Greenland. *Geology* **19**, 510–513.

Hudleston, P. J., Schultz-Ela, D. & Southwick, D. L. 1988. Transpression in an Archean greenstone belt, northern Minnesota. *Can. J. Earth Sci.* **25**, 1060–1068.

Jamieson, R. A. & Vernon, R. H. 1987. Timing of porphyroblast growth in the Fleur de Lys Supergroup, Newfoundland. *J. metamorphic Geol.* **5**, 273–288.

Kennedy, M. J. 1975. Repetitive orogeny in the northeastern Appalachians: new plate models based upon Newfoundland examples. *Tectonophysics* **28**, 39–87.

Kidd, W. S. F. 1974. The evolution of the Baie Verte Lineament, Burlington Peninsula, Newfoundland. Unpublished Ph.D. thesis. Cambridge University.

Kirkwood, D. 1989. Géologie structurale de la région de Percé. Gouvernement du Québec, Ministère de l'Énergie et des Ressources, Direction générale de l'exploration géologique et minérale **ET 87-17**, 33 pp.

Lafrance, B. & Williams, P. F. 1992. Oblique-cleavage folds: a critical discussion with examples from the Canadian Appalachians. *Tectonophysics* **207**, 315–330.

Law, R. D., Knipe, R. J. & Dayan, H. 1984. Strain path partitioning within thrust sheets: Microstructural and petrofabric evidence from the Moine thrust zone at Loch Eriboll, northwest Scotland. *J. Struct. Geol.* **6**, 477–497.

Malo, M. & Béland, J. 1989. Acadian strike-slip tectonics in the Gaspé region, Quebec Appalachians. *Can. J. Earth Sci.* **26**, 1764–1777.

Malo, M., Kirkwood, D., De Broucker, G. & St-Julien, P. 1992. A re-evaluation of the position of the Baie Verte-Brompton Line in the Quebec Appalachians: the influence of Middle Devonian strike-slip faulting in the Gaspé Peninsula. *Can. J. Earth Sci.* **29**, 1265–1273.

Means, W. D. 1989. Stretching faults. *Geology* **17**, 893–896.

Means, W. D. 1990. One dimensional kinematics of stretching faults. *J. Struct. Geol.* **12**, 267–272.

Mount, V. S. & Suppe, J. 1987. State of stress near the San Andreas fault: Implications for wrench tectonics. *Geology* **15**, 1143–1146.

Oldow, J. S., Bally, A. W. & Avé Lallement, H. G. 1990. Transpression, orogenic float, and lithospheric balance. *Geology* **10**, 991–994.

Passchier, C. W. & Simpson, C. 1986. Porphyroclast systems as kinematic indicators. *J. Struct. Geol.* **8**, 831–843.

Piasecki, M. A. J. 1988. Strain-induced mineral growth in ductile shear zones and a preliminary study of ductile shearing in western Newfoundland. *Can. J. Earth Sci.* **25**, 2118–2129.

Pinet, N. & Cobbold, P. R. 1992. Experimental insights into the partitioning of motion within zones of oblique subduction. *Tectonophysics* **206**, 371–388.

Platt, J. P. 1979. Extensional crenulation cleavage. *J. Struct. Geol.* **1**, 95.

Platt, J. P. & Vissers, R. L. M. 1980. Extensional structures in anisotropic rocks. *J. Struct. Geol.* **2**, 397–410.

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

Price, G. P. 1985. Preferred orientations in quartzites. In *Preferred Orientation in Deformed Metals and Rocks: An Introduction to Modern Texture Analysis* (edited by Wenk, H.-R.) Academic Press, Inc. 385–406.

Ratliff, R. & Morris, A., Dodt, M. 1988. Interaction between strike-slip and thrust-shear: deformation of the Bullbreen Group, central-western Spitsbergen. *J. Geol.* **96**, 339–349.

Robin, P.-Y.F. & Cruden, A. R. 1994. Strain and vorticity patterns in ideally ductile transpression zones. *J. Struct. Geol.* **16**, 447–466.

Sanderson, D. J. 1973. The development of fold axes oblique to the regional trend. *Tectonophysics* **16**, 55–70.

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

Simpson, C. & Schmid, S. M. 1983. An evaluation of criteria to deduce the sense of movement in sheared rocks. *Bull geol. Soc. Am.* **94**, 1281–1288.

Teyssier, C., Tikoff, B. & Markley, M. 1995. Oblique plate motion and continental tectonics. *Geology* **23**, 447–450.

Tikoff, B. & Teyssier, C. 1994. Strain modeling of displacement-field partitioning in transpressional orogens. *J. Struct. Geol.* **16**, 1575–1588.

van Berkel, J. T., Johnston, H. P. & Currie, K. L. 1986. A preliminary report on the geology of the southern Long-Range, southwest Newfoundland. *Geological Survey of Canada* **86**, 157–170.

Vauchez, A. & Nicolas, A. 1991. Mountain building: strike-parallel motion and mantle anisotropy. *Tectonophysics* **185**, 183–201.

Walcott, R. I. 1978. Present tectonics and Late Cenozoic evolution of New Zealand. *Geophys. J. R. astr. Soc.* **52**, 137–164.

Watson, K. de P. 1947. Geology and mineral deposits of the Baie Verte-Ming's Bight area, Newfoundland. *Geological Society of Newfoundland Bulletin* **21**, 48 pp.

White, J. C. & Flagler, P. A. 1992. Deformation within part of a major crustal shear zone, Parry Sound, Ontario: structure and kinematics. *Can. J. Earth Sci.* **29**, 129–141.

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

Williams, H. D. 1977. Ophiolitic mélange and its significance in the Fleur de Lys Supergroup, northern Appalachians. *Can. J. Earth Sci.* **14**, 987–1003.

Williams, H. & St-Julien, P. 1978. The Baie Verte-Brompton Line in Newfoundland and regional correlations in the Canadian Appalachians. *Geol. Surv. Can.* **78-1A**, 225–229.

Williams, H. & St-Julien, P. 1982. The Baie Verte-Brompton Line: Early Paleozoic continent–ocean interface in the Canadian Appalachians. *Geological Association of Canada* **27**, 179–207.

Williams, H., Colman-Sadd, S. P. & Swinden, H. S. 1988. Tectonic-stratigraphic subdivisions of central Newfoundland. In *Current Research, Part B. Geol. Surv. pap. Canada* **88-1B**, 91–98.

Williams, H., Hibbard, J. P. & Bursnall, J. T. 1977. Geological setting of asbestos-bearing rocks along the Baie Verte Lineament, Newfoundland. *Geol. Surv. Can.* **77-1A**, 351–360.

Williams, P. F. & Price, G. P. 1990. Origin of kinkbands and shear-band cleavage in shear zones; an experimental study. *J. Struct. Geol.* **12**, 145–164.