Journal of Structural Geology, Vol. 11, No. 6, pp. 705 to 716, 1989
Printed in Great Britain

0191-8141/89 $03.00 + 0.00

Structural evolution of a transpression zone in north central Newfoundland

BRUNO LAFRANCE

University of New Brunswick, Department of Geology, Fredericton, N.B., Canada E3B 5A3

(*Received* 18 *July* 1988; *accepted in revised form* 6 *April* 1989)

Abstract—A dextral transpression zone in Notre Dame Bay, north central Newfoundland, was active in Middle to late Silurian times following the closure of a marginal basin. The deformation in the transpression zone was complex and involved several generations of structures. Deformation was initially distributed over wide, low-strain zones through the formation of a penetrative regional cleavage. With steepening and overturning of the strata, the orientation of the sequence was favorable for strike-slip bedding-parallel movement and the deformation became more efficiently accommodated by shear within the incompetent lithologic horizons, forming localized shear zones in high-strain zones. The sheared incompetent horizons define a belt of mélange, the 'Boones Point Complex', centered on the contact between a volcanic island arc terrane to the north and a marginal basin terrane to the south. Two cleavages with similar orientation and morphology were generated at different times during the structural evolution of the transpression zone. One of the cleavages is regional in distribution, and is transposed by the shear zones. Folds with no axial-planar cleavage, but a cleavage cutting obliquely across their hinges, are reported. They have the geometry of transected folds but are clearly the result of overprinting by a younger generation of cleavage.

INTRODUCTION

IN DEFORMED low-grade metamorphic areas, the difficulty in differentiating between different generations of mesostructures may lead to an oversimplified interpretation of their geological history. Structural analysis of a low-grade area in the Dunnage Zone, north central Newfoundland (Fig. 1), reveals a more complex story than that previously proposed (Heyl 1936, Helwig 1967, 1970, Horne 1968, 1969). The Dunnage Zone (Williams 1979) is the remnant of an early ocean that once separated the North American and Avalonian continent in Cambrian to Ordovician times. It generally has a NE-trending tectonic fabric, but the trend swings towards a more easterly orientation to the north in Notre Dame Bay. The geology of the study area in the Bay of Exploits is characterized by two distinct terranes—an island arc volcanic terrane to the north (Strong 1977) and a marginal basin terrane (van der Pluijm 1987, van der Pluijm & van Staal 1988) to the south (Fig. 2). In both terranes, the stratigraphy may be divided into a Lower to Middle Ordovician volcanic unit overlain by a sedimentary unit consisting mainly of turbiditic siltstones and sandstones (Dean 1978, Kean *et al.* 1981, but see Nowlan & Thurlow 1984, Dunning *et al.* 1987). The volcanic unit in the marginal basin terrane consists of a collage of oceanic island volcanics, island arc volcanics and ocean floor fragment volcanics (Dean 1978, Jacobi & Wasowski 1985, Swinden 1985, Wasowski & Jacobi 1986). The age of the sedimentary unit is well constrained in the southern marginal basin terrane by numerous Upper Ordovician to Lower Silurian fossil horizons (Neuman 1976, 1984, McKerrow & Cocks 1981, Arnott *et al.* 1985, van der Pluijm *et al.* 1987, Elliott 1988). The contact between the island arc volcanic terrane and the marginal basin terrane is faulted. An E–W-trending belt of mélanges, the Boones Point Complex (Fig. 1), is centered on the contact between the two terranes (Nelson 1981).

Previous structural studies in Notre Dame Bay emphasized an early deformation history associated with normal faulting or thrust faulting followed by regional tectonic folding (Nelson 1981, Karlstrom *et al.* 1982, Arnott 1983, Reusch 1983, van der Pluijm 1984, 1986, Elliott 1985, Pickering 1987). The Boones Point Complex has been interpreted as a belt of mélanges representing subaqueous débris flows deposited and deformed by an overriding S-directed nappe in late

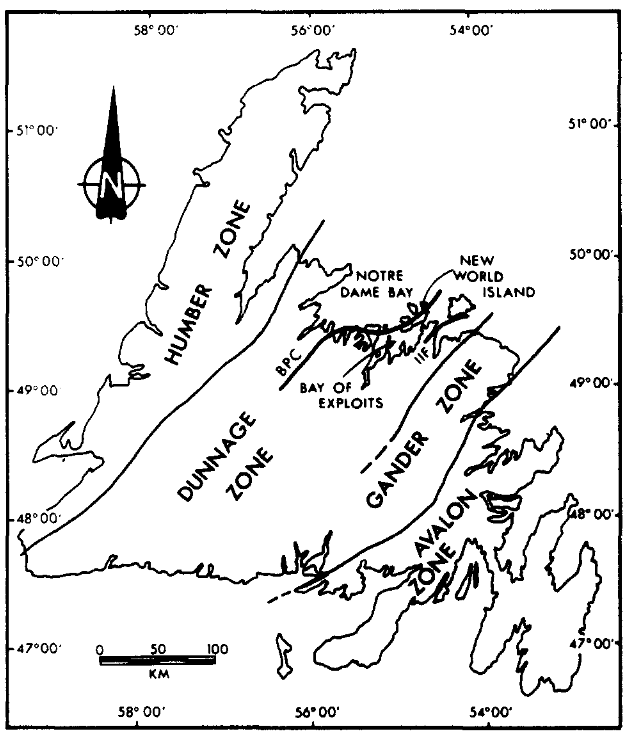

Fig. 1. The tectonostratigraphic zones of Newfoundland (Williams 1979). BPC = Boones Point Complex, IIF = Indian Island Fault.

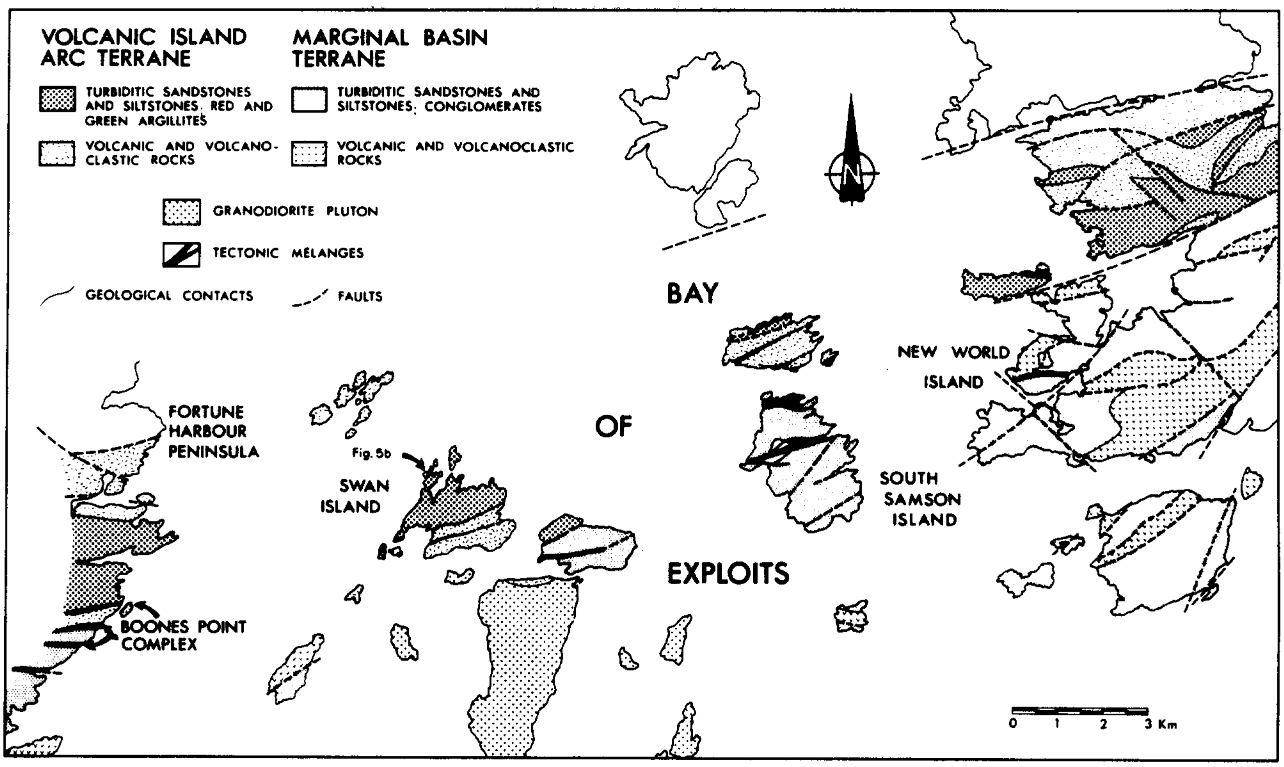

Fig. 2. Geological map of the Bay of Exploits. The Boones Point Complex consists of anastomosing horizons of mélange centered on the contact between the island arc volcanic terrane to the north and the marginal basin terrane to the south. Geology compiled from Horne (1968), Antonuk (1986), Elliott (1988) and mapping by the author.

Ordovician to Silurian times (Nelson 1981). Contrary to the previous interpretation, the mélanges are found, through detailed structural mapping in the Bay of Exploits, to be tectonic in origin, and are interpreted to be the product of deformation in a transpression zone.

Transpression describes a complex movement involving both shortening normal to the zone boundary and transcurrent shear parallel to the zone boundary (Harland 1971, Sanderson & Marchini 1984). In cases where the strata are shallowly-dipping or horizontal, the deformation will be initially distributed over wide areas through compressive structures such as folds and thrusts which will result in the local steepening of the strata (Harland 1971). Subsequent deformation will become more efficiently accommodated within the incompetent steeply-dipping horizons, and will result in transcurrent faulting parallel to bedding. Thus the incompetent horizons will localize the formation of shear zones within the transpression zone.

The Notre Dame Bay transpression zone is a very low-grade metamorphic zone with a complex deformational history involving several generations of structures. The deformation is partitioned into 'low-strain zones' and 'high-strain zones'. High strain zones are narrow zones, less than 1 km wide, characterized by intense folding and shear zones. They are localized at the contact between lithological units and terranes of contrasting competence. Low-strain zones may reach 5 km in width and are characterized by a strong penetrative regional foliation and by the relative absence of folds (Williams *et al.* 1988). The structural evolution of the transpression zone will be described in detail with special emphasis on overprinting relationships in high-strain zones. Structures with the geometry of transected folds are observed in the high-strain zones, and are found to be the result of overprinting of the folds by a younger generation of cleavage. This interpretation will be discussed further with reference to other examples of transected folds described in the literature.

DEFORMATION STRUCTURES

Structures in the volcanic arc and marginal basin terranes are especially well developed in the turbiditic siltstone and sandstone sequences. Bedding is generally overturned and N-younging, and continuity across and along strike is commonly broken by both approximately bedding-parallel and transverse high-angle faults. Continuous sequences kms wide, however, are present on eastern New World Island and on Swan Island in the Bay of Exploits. The structural evolution of the region has been mainly derived from excellent overprinting relationships displayed in a high-strain zone on Swan Island (Fig. 2), which is located immediately north of the contact between the island arc volcanic terrane and the marginal basin terrane. The Boones Point Complex outcrops on the south side of the island.

Structures in the region are divided into four generations of folds (Table 1) and two generations of shear zones. Two cleavages are related to the folds but only one (S_3) is a regional penetrative cleavage. S_3 generally dips more shallowly than bedding and strikes anticlockwise of bedding except on the short limbs of F_3 folds

Table 1. Fold generations

F_4	—mesoscopic open to tight folds —steeply-plunging dextral folds —fold S_3 and the transposed cleavage in the mélange
F_3	—mesoscopic and macroscopic folds with sharp hinges and straight limbs —upward-facing S-verging dextral folds —penetrative axial-planar regional foliation (S_3)
F_2	—mesoscopic folds with sharp hinges and straight limbs, and with high periodicity and short wavelengths of the order of a few meters —upward-facing S-verging folds —weak axial-planar cleavage (S_2)
F_1	—mesoscopic, isoclinal, intrafolial, rootless fold —E–W-trending steeply-dipping and plunging —overprinted by S_3 and the transposed cleavage in shear zones —generally associated with mélanges and shear zones

where it dips more steeply than bedding. A weak axial-planar cleavage (S_2) is locally associated with F_2 folds but, on Swan Island, S_2 was completely obliterated by an overprinting cleavage.

F_2 folds are similar in style to F_3 folds, but overprinting of F_2 folds by F_3 structures is observed both at the macroscopic and mesoscopic scale. Figure 3 is a diagrammatic form surface representation of northwest Swan Island which has been divided into structural domains based on the orientation of F_2 fold axes. On a synoptic equal-area projection, F_2 fold axes define a broad girdle which, together with the rotation of their axial surfaces, indicates that they were refolded by a large-scale fold. Since F_2 folds are very asymmetrical, the axis of folding of bedding (S_0) the long limb on F_2 folds should approximate the orientation of the large-scale fold axis. As

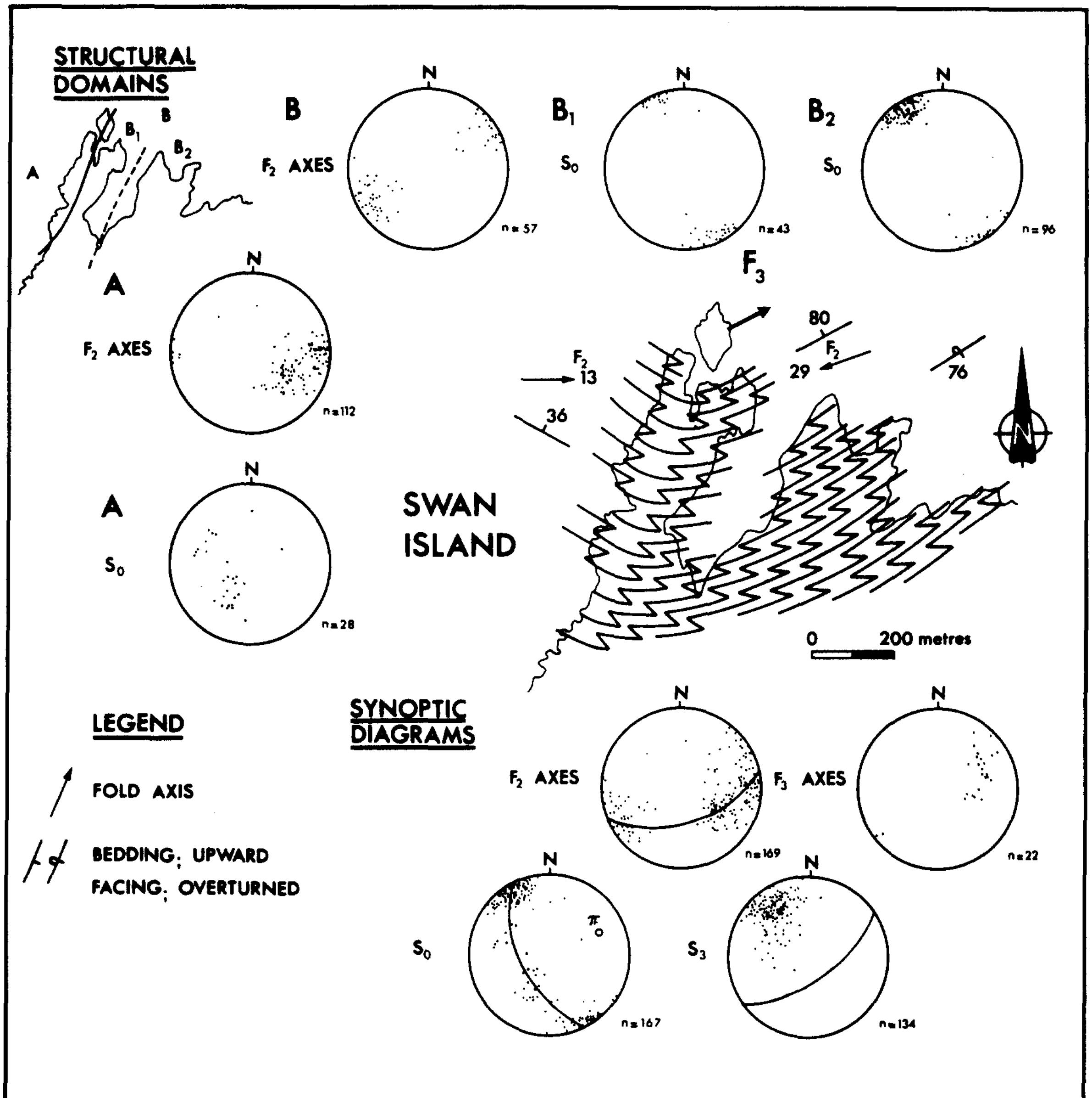

Fig. 3. Diagrammatic form surface representation of northwest Swan Island. Equal-area, lower-hemisphere projections of poles to bedding on the long limb of F_2 folds (S_0), of poles to S_3 (the regional cleavage), and of F_2 and F_3 axes. The orientation data for F_3 folds is from south Swan Island, and includes both measurements of actual fold axes and of bedding–cleavage (S_3) intersection lineation in cases where the fold axis could not be directly measured. Boundaries of cylindrical F_2 domains A and B are shown at upper left. Domain B is divided in subdomains B_1 and B_2 based on the orientation of S_0.

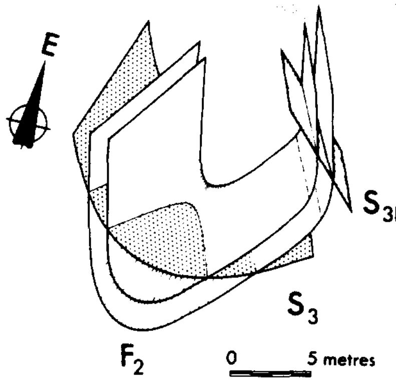

Fig. 4. Schematic representation of field structures shown in Fig. 5(b). F_2 fold overprinted by S_3. S_3 is bent, has an interlimb angle larger than the F_2 interlimb angle, and the fold in S_3 plunges more steeply to the east than F_2. (In the diagram, the fold is tilted to the west to better display the geometry.) The shallowly-dipping limb of F_2 is also partially overturned by a shear zone with formation of S_{3B} in the overturned beds.

shown in Fig. 3, the poles to bedding lie in a great circle whose pole (π) lies within the regional cleavage (S_3), which in turn has an orientation similar to the axial plane of the large-scale fold. There are relatively few examples of mesoscopic folds with the regional cleavage as their axial-planar cleavage (Fig. 5a) but the average orientation of their fold axes is similar to the orientation of the π axis. The regional cleavage and the large-scale fold are therefore interpreted as S_3 and F_3, respectively.

The geometry of the overprinting of mesoscopic F_2 folds by S_3 depends on the orientation of the enveloping surface of F_2 folds with respect to S_3. Figure 4 is a diagrammatic representation of an F_2 fold from domain A (Fig. 3) on the shallowly-dipping limb of the large-scale F_3 fold. A field example of the same structure is shown in Fig. 5(b). In the profile plane of the fold, a cleavage (S_3) intersects the same bedding surface twice (see also Fig. 5b). S_3 is also folded, with an interlimb angle larger than the interlimb angle of the folded bedding surface. The axis of the folded cleavage plunges more steeply than the F_2 fold axis. This fold–cleavage geometry was interpreted by van der Pluijm (1984, 1986) as the result of folding two mutually inclined surfaces, but it will be shown in the next section to be the result of simple overprinting of the fold (F_2) by the cleavage (S_3). Both models will be analysed further in the next section.

Figures 5(c) & (d) show the profile plane and a horizontal plane, approximately parallel to the fold axis, of an inclined shallowly-plunging F_2 fold from domain B (Fig. 3). No axial-planar cleavage is associated with the fold. The cleavage in the overturned limb of the fold appears axial planar in the profile plane, but cuts obliquely across the fold axis at an angle of approximately 10° in an anticlockwise sense as seen on the plane parallel to the fold axis. The fold–cleavage geometry is similar to the geometry of transected folds reported by various workers. Transected folds by definition form more or less synchronously with the cleavage (Borradaile 1978). In this case, however, the cleavage is S_3 and it overprints the fold. The implications of this interpretation for models of cleavage transected fold formation will be discussed below.

Domain B (Fig. 3) has been further divided into two subdomains based on the orientation of the N-younging limb of F_2 folds. In subdomain B_1 nearest the hinge region of the large-scale F_3 fold, the N-younging beds are both upward-facing (dipping north) and downward-facing (dipping south), whereas in subdomain B_2 to the east, the N-younging beds dip more northeasterly and are predominantly downward-facing. F_2 fold hinges mesoscopically folded with their N-younging limbs overturned to the north have also been observed. Overturning of the beds can therefore be explained by syn- or post-F_3 folding.

Two types of shear zones are observed: (1) shear zones of regional extent and associated with mélanges (type 1 shear zones); and (2) narrow shear zones with a vertical movement vector and without major brecciation (they will be referred to as 'VM shear zones'). Type 1 shear zones occur in argillite horizons between volcanic sequences or in argillite-rich horizons in well stratified sedimentary sequences. The shear zones are usually parallel to the local stratigraphy but they also cut across and transpose earlier F_2 and F_3 structures. Near the shear zones, continuous, steeply-dipping, sedimentary sequences with a well developed S_3 cleavage grade into broken disrupted horizons. A new cleavage similar in morphology and in orientation to S_3 is present in the argillite matrix between the boudinaged and fractured bed fragments (Fig. 6a). With increasing strain, the bed fragments are rotated into parallelism with the cleavage surfaces, which then envelop and flow around the bed fragments and blocks (Fig. 6b). The new cleavage, although similar in orientation to S_3, only formed after fragmentation of the beds in shear zones. Since the shear zones also transpose earlier F_2 and F_3 structures, they are interpreted to form following F_3 during a progressive deformation. The sheared argillite-rich horizons form tectonic mélanges which are part of the Boones Point Complex.

VM shear zones are present in the shallowly-dipping stratified sequences (Domain A, Fig. 3). They are less than 1 m wide, and dip 20–30° to the southeast (Fig. 6c). The sense of movement is south side up. Beds are continuous across the shear zones, but are steeply-dipping and overturned to the north within the shear zones. There is a cleavage (S_{3B}) in the incompetent overturned mudstone beds (Fig. 6c) with similar morphology and orientation to S_3, striking anticlockwise to bedding and dipping more shallowly than bedding. S_{3B} and bedding intersect to define a shallowly E-plunging lineation. North-younging limbs of F_2 folds are commonly overprinted by shear zones which resulted in the partial overturning of the limbs and in the formation of a shear zone cleavage (S_{3B}) (Fig. 4). The shear zones generally have a south side up sense of movement but they also occur in conjugate pairs (Fig. 6d) which may also have involved brittle behavior. VM shear zones are probably relatively late in the deformation history and are interpreted to be younger than F_3 structures on the

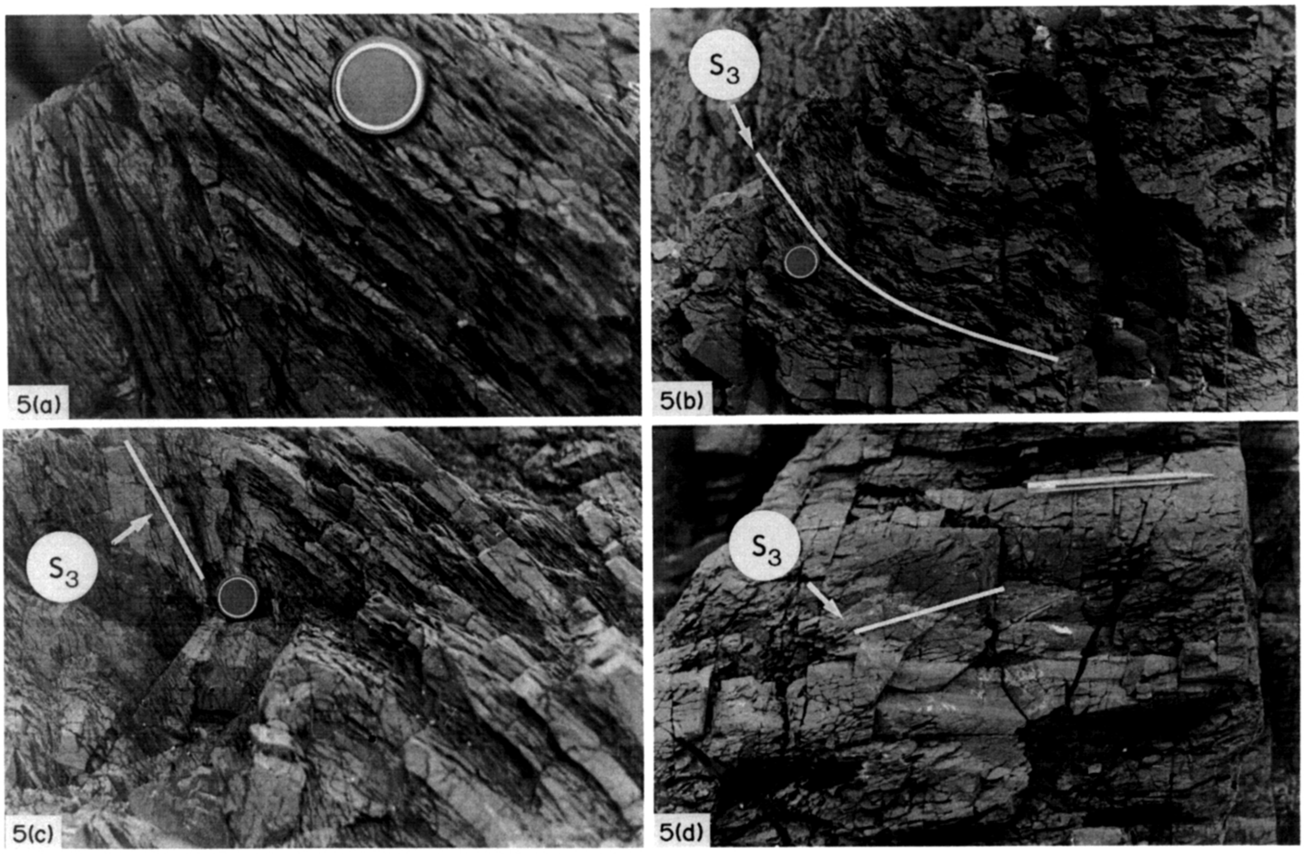

Fig. 5. (a) F_3 fold with its well developed axial-planar cleavage (S_3). Vertical section: viewed looking east. (b) F_2 fold overprinted by a cleavage (S_3). The field location of the fold is indicated on Fig. 2. The cleavage is bent with an interlimb angle larger than the interlimb angle of the folded beds. S_3 is parallel to the white line. Vertical section: viewed looking east. (c & d) Shallowly-plunging inclined F_2 fold from domain B in Fig. 3. (c) Vertical face: view to east. The fold is overprinted by a cleavage (S_3), parallel to the white line, which appears axial planar on the overturned limb. (d) Plan view of fold in (c). Photograph is taken looking SE. The trace of the cleavage (parallel to the white line) cuts obliquely across the fold axis with a 10° anticlockwise angle. The fold axis is parallel to the pen.

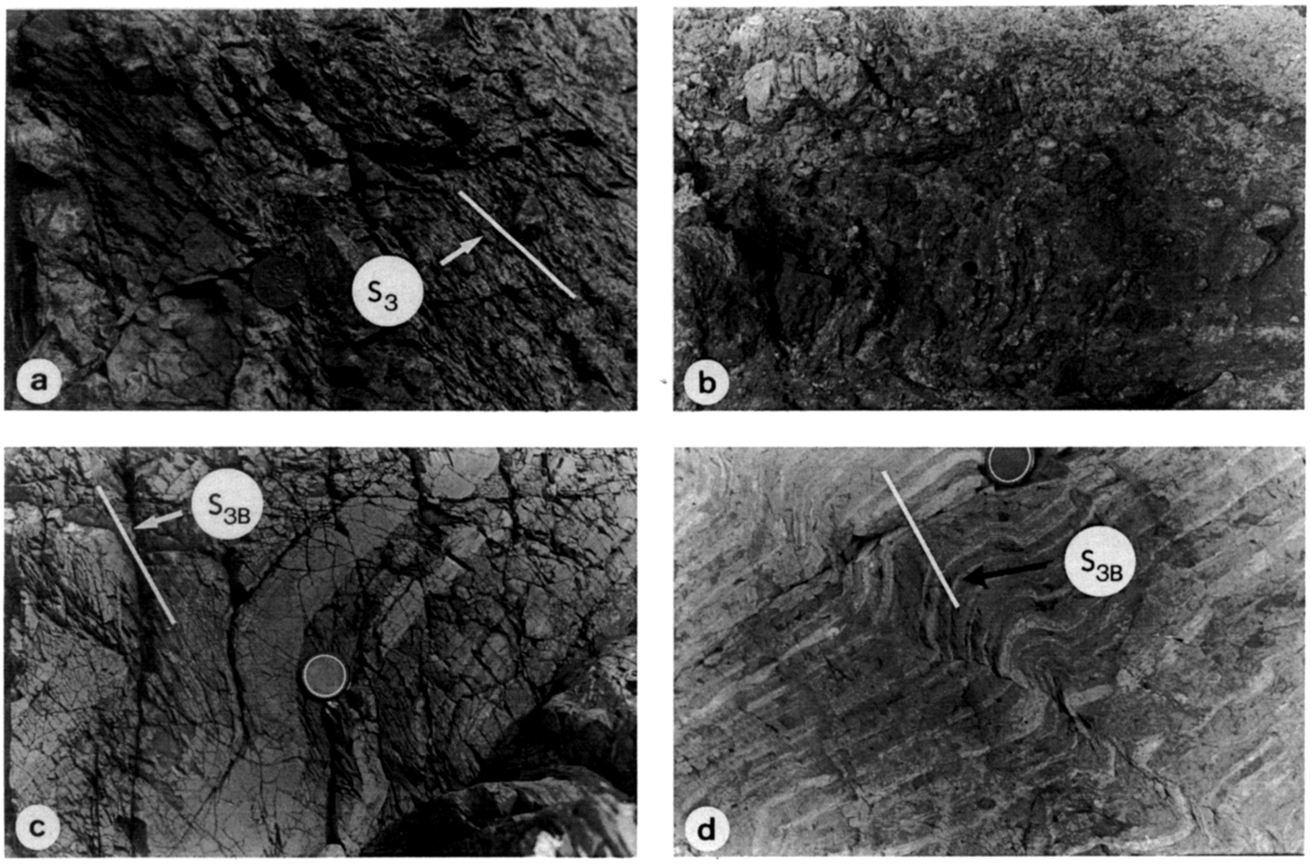

Fig. 6. Shear zone associated with mélanges (a) & (b), and shear zones with a vertical movement vector and without major brecciation (c) & (d). (a) The deformation was localized in argillite-rich incompetent horizons. The sandstone beds are boudinaged and fragmented. A cleavage similar in orientation and morphology to S_3 (parallel to the white bar), is present in the argillite matrix between the bed fragments. Vertical section: viewed looking east. (b) With increasing deformation, the bed fragments and the foliation are rotated into parallelism. The sheared argillite-rich horizons form tectonic mélanges. A steeply plunging dextral fold (F_4) overprints the transposed foliation. Horizontal section: west and east are on the left and right, respectively, of the photograph. (c) Shear zone in low-strain zones with south-side up movement. A cleavage (S_{3B}), parallel to the white line, is only developed in the overturned mudstone beds in the shear zone. Vertical section: viewed looking east. (d) Conjugate shear zones. S_{3B}, parallel to the white line, is present only in the overturned mudstone beds of the south-side up member of the conjugate pair. Brittle failure was associated with the shear zones formation. Vertical section: viewed looking east.

basis of: (1) their limited occurrence on the shallowly-dipping limb only of the large-scale F_3 fold; and (2) their style of deformation which reflects the localization of the deformation in narrow zones as opposed to F_3 structures which are regional in extent. No overprint between VM shear zones and F_4 folds is observed. The symbol S_{3B} is used to represent the cleavage formed in VM shear zones to stress its similarity with the regional S_3 cleavage.

DISCUSSION

The mélanges and F_1 folds in eastern Notre Dame Bay are interpreted to be due to either: (1) early thrusting (Nelson 1981, van der Pluijm 1984, 1986, Pickering 1987, Elliott 1988); or (2) soft- or wet-sediment deformation associated with superficial slumping and gravity sliding (Helwig 1967, 1970, Horne 1968, Eastler 1971, Arnott 1983, Pickering 1987). In an environment of turbiditic sandstone deposition, disruption of lithological horizons may be due to slumping or gravity sliding. The deformation in the Boones Point Complex, however, occurred in rocks that were already consolidated. West of the studied area, for example, mafic volcanic blocks in the Boones Point Complex are strongly foliated and elongated parallel to the shear-zone foliation (Nelson 1981). This observation argues strongly against soft-sediment deformation of the sedimentary rocks surrounding the volcanic blocks. The mélanges on South Samson Island are interpreted as olistostromes by McKerrow & Cocks (1981). A penetrative cleavage in the mélanges, however, flows around a 30 m thick block of pebble–cobble conglomerate indicating that the block was consolidated prior to the deformation in the mélanges. Folded quartz veins are also observed in the mélanges. The mélanges can therefore be interpreted as sheared dismembered volcanic and turbiditic sandstone–siltstone sequences which were consolidated prior to deformation. Other structures in the area interpreted as slump folds (Horne 1969) have recently been demonstrated to be tectonic folds (Elliott & Williams 1988).

The deformation in the Boones Point Complex was interpreted by Nelson (1981) to be due to thrusting. The mélanges, however, are in shear zones which transpose the regional S_3 foliation and are, therefore, relatively late in the tectonic history of the region. Folds overprinted by the foliation in the mélanges were originally thought to be F_1 folds, but are now interpreted as tectonic F_2 or F_3 folds. There may have been, in the Boones Point Complex, disrupted horizons related to an early thrusting event, but the late overprinting shearing would make them indistinguishable from shear-zone mélanges.

To the south of the Bay of Exploits, the Dunnage Zone is deformed by large open folds with wavelengths of 1–5 km. F_2 folds in eastern Notre Dame Bay have been interpreted as parasitic folds on the N-younging limbs of these regional folds (Karlstrom *et al.* 1982, Reusch 1983, van der Pluijm 1984, 1986, Pickering 1987). Since bedding in the Notre Dame Bay area is generally N-younging and overturned, the area has been interpreted as lying on the N-younging overturned limbs of large-scale folds that were repeated by faulting. Alternatively, F_2 folds may have formed by shortening across an initially N-dipping sequence since they are S-verging and upward-facing. The absence of thick sequences of S-younging beds in the area representing the S-facing limbs of regional F_2 folds favors the second model. Furthermore, the style of F_2 folds is more consistent with shortening of an initially N-dipping sequence than with regional open folding.

The fold–cleavage geometry shown in Fig. 4 is unusual with a folded cleavage intersecting the same bedding surface twice. van der Pluijm (1984, 1986) explained this geometry as the response to a space problem arising from folding of two mutually inclined surfaces (Figs. 7a1 & a2). The space problem is accommodated by slip on the cleavage plane accompanied by rotation of the intersection lineation toward the bedding surface fold axis (Fig. 7a3). The profile plane of the fold would then intersect the intersection lineation twice, and the fold–cleavage geometry would be similar to Fig. 4. The model requires the intersection lineation, parallel to bedding and initially perpendicular to the fold axis, to undergo an angular shear of 55–65°. The planarity of the limbs as well as the lack of disruption of the folds, however, argues against intense ductile shear in the limbs parallel to the fold axis. Further, mesoscopic structures indicative of the deformation of lithified or partially lithified sediments in low-grade shear zones, such as foliations and lineations defined by elongated grains or en échelon veins and fractures (Knipe & White 1979), are not observed. On the other hand, if the beds were unlithified, an angular shear of 55–65° would cause thinning or even boudinage of the limbs. None of these structures are observed on the limbs or in the hinges of the folds. Since no evidence of significant ductile shear in the limbs of the folds parallel to the fold axis is observed, the model does not adequately explain the fold geometry. Furthermore, since the folds are shallowly plunging, any space problems in the hinges of the folds would be accommodated by vertical extension rather than by horizontal extension parallel to the fold axes. An alternative model is proposed where the cleavage, instead of being folded by the fold, overprints it. The deformation is accommodated by layer-parallel shear in the incompetent mudstone layers. The planarity of the limbs and the absence of a cleavage in the competent sandstone layers add credence to this interpretation. Bending of the overprinting cleavage may be explained by reverse refraction of the cleavage due to layer-parallel shear during the reopening of the fold (Fig. 7b) (Williams 1985).

The validity of the model, however, depends on the overall rotation or bending of the cleavage during reopening of the fold. The rotation of the cleavage on the right limb of the fold with respect to an external frame of reference is partitioned into an anticlockwise spin component (SPIN) resulting from the rotation of

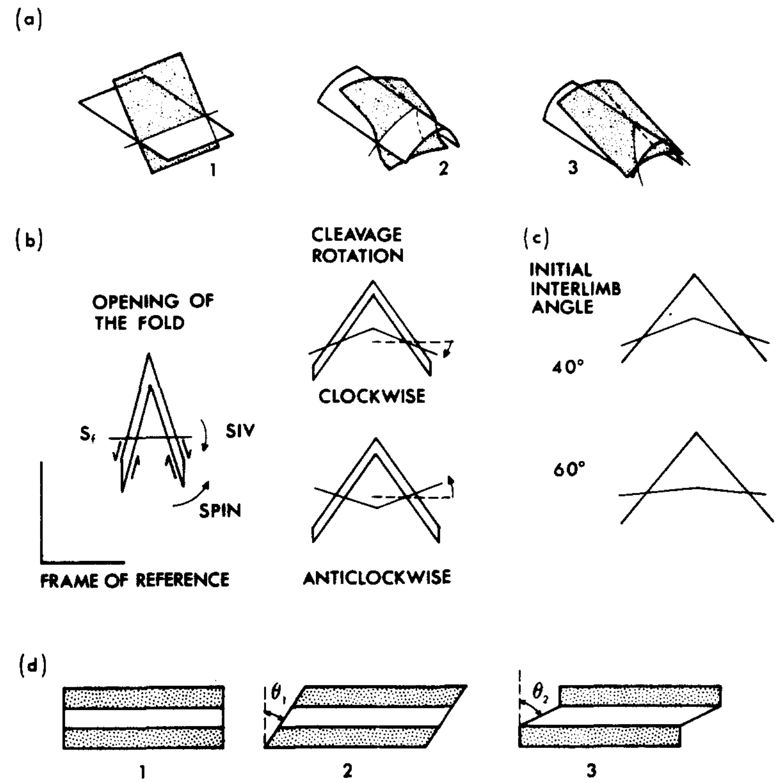

Fig. 7. Models for the development of the fold–cleavage geometry shown in Figs. 5 and 6. (a) van der Pluijm's (1984, 1986) model to explain the F_2–S_3 geometry shown in Fig. 5(b). (1) Folding of two inclined surfaces creates (2) a space problem which is accommodated (3) by the rotation of the intersection lineation towards the fold axis. (b) An overall clockwise rotation of the cleavage on the right limb of the fold with respect to an external frame of reference produces the F_2–S_3 geometry observed in the field (SIV > SPIN). S_f = cleavage, SIV = shear induced vorticity component, SPIN = component of angular rotation of the limbs. (c) Rotation of the cleavage resulting from the opening of the fold for initial interlimb angles of 40° and 60°. F_2 interlimb angles observed in outcrop are around 80°. (d) (1) Deformation of interlayered sandstone (stippled) and mudstone (blank) beds assuming (2) no competence contrast between the sandstone and mudstone beds. (3) Deformation with strain partitioning between the sandstone and mudstone beds. The clockwise rotation of the cleavage in the mudstone beds due to shear induced vorticity is larger in case (3) than in case (2) (ie. $\theta_2 > \theta_1$).

the limb during the opening of the fold, and a clockwise shear-induced component (SIV) resulting from layer-parallel shear in the limb (Fig. 7b) (Means *et al.* 1980, Lister & Williams 1983). An overall clockwise and anticlockwise rotation, respectively, of the cleavage would result in the fold geometries shown in Fig. 7(b). The rotation of the cleavage was calculated for equal increments of angular layer-parallel shear and angular opening of the fold for initial interlimb angles of 40° and 60° (Fig. 7c) assuming similar mechanical behavior for the sandstone and mudstone beds (Figs. 7d1 & d2). Sandstone beds are more competent than mudstone beds, however, causing deformation to be preferentially accommodated by bedding-parallel shear in the mudstone layers which accentuates the shear induced vorticity (SIV) component in these and therefore the clockwise rotation of the cleavage in the incompetent mudstone layers (Fig. 7d3). Figure 7(c), therefore, represents cases for which the clockwise shear induced rotation of the cleavage is minimized. In all cases, the overall rotation of the cleavage with respect to the right limb is clockwise. Overprinting of the fold by the cleavage is therefore an acceptable model.

The model assumes that the cleavage behaves as a material plane, and consequently does not track the *XY* plane of strain during non-coaxial deformation. The cleavage is defined by the dimensional preferred orientation of phyllosilicate grains and by discrete planar concentrations of insoluble materials. A foliation defined only by the preferred dimensional orientation of phyllosilicate grains could possibly track the *XY* finite plane of strain by grain-boundary sliding of the phyllosilicates grains (Williams 1976). A foliation defined also by planes of insoluble materials could only track the *XY* finite plane of strain if migration of material across the planes takes place at exactly the right rate (Williams 1976), which is highly improbable. It is therefore reasonable to assume that the cleavage behaves as a material plane.

The shear zones associated with mélanges are localized zones of intense deformation which overprint F_2 and F_3 structures. Beds are brittlely disrupted and possibly reflect, assuming otherwise similar conditions of deformation, an increase in strain rate. Since a cleavage similar in morphology and orientation to S_3 is developed in the mudstone matrix between boudinaged sandstone bed fragments, the shear zones are interpreted to have formed as deformation, which was initially distributed over a wider area, became concentrated within incompetent layers following F_3 folding during a progressive deformation. VM shear zones (Fig.

6c) resulted in the local overturning of shallowly-dipping sequences (domain A in Fig. 3) and in the formation of a cleavage (S_{3B}) similar in morphology and orientation to the regional cleavage (S_3). Since brittle deformation is also associated with VM shear zones, they could be causally related with type 1 shear zones and could have formed to accommodate vertical extension, which is generally associated with transpression zones (Sanderson & Marchini 1984). Another interpretation is also possible. Beds in Notre Dame Bay are generally overturned to the north which, as discussed earlier, is interpreted to be either syn- or post-F_3. Since VM shear zones also resulted in the overturning of bedding on shallowly-dipping sequences (Domain A), VM shear zones and the regional overturning of bedding could be causally related and interpreted to be younger than F_3 structures and type 1 shear zones.

F_2, F_3 and F_4 folds are all spatially associated with the high-strain zone. Since the shear zones associated with the mélanges overprint dextral F_3 folds and are overprinted by dextral F_4 folds, the sense of movement in the high-strain zone is interpreted to be dextral. Since F_3 fold axes as well as the regional cleavage generally trend anticlockwise of F_2 fold axes, this geometry also indicates clockwise rotation of F_2 folds prior to F_3 folding in a dextral zone. Overturning of the beds may have occurred either following or late during the formation of F_3 folds and mélanges.

The discussion up to this point has been on structures formed in a high-strain zone on Swan Island but the same structures are present in high-strain zones on New World Island (Williams *et al.* 1988). Low-strain zones are characterized by a relative absence of folds, but also by the presence of the regional cleavage (S_3). Since structures (F_2, F_3, S_3) have formed by compression in zones that were undergoing dextral transcurrent shear, Notre Dame Bay can therefore be interpreted as a transpression zone.

Transected folds

Various models have been proposed to explain the formation of transected folds—that is, the more or less synchronous development of folds and a cleavage obliquely cutting across them (Powell 1974, Stringer 1975, Borradaile 1978, Sanderson *et al.* 1980, Stringer & Treagus 1981, Treagus & Treagus 1981, Murphy 1985, Soper 1986, Blewett & Pickering 1988). These models are derived from deformed zones involving only one generation of cleavage or fold. Underlying these models is the assumption that since no truly axial-planar cleavage is associated with the first generation of folds formed within the deformed zones, then the cleavage cutting obliquely across the fold hinges developed more or less synchronously with the folds. Movement zones may however have a complex structural history involving several generations of structures. Five generations of structures are associated with the Notre Dame Bay transpression zone. F_2 is the first generation of folds formed within the movement zone. On Swan Island, only one cleavage is associated with F_2 folds and it cuts obliquely across the hinges and the limbs of F_2 folds. A strong bedding-parallel foliation is also observed on the microscopic scale. Both the bedding-parallel foliation and the cleavage are defined by the preferred orientation of phyllosilicates and detrital grains, and by discrete concentrations of insoluble material. No axial-planar cleavage is observed; it may have been completely obliterated by the cleavage which cuts obliquely across the fold hinges, but the presence of a strong bedding-parallel foliation indicates that it may only have been locally developed. As previously discussed, the cleavage associated with F_2 folds is S_3, and it is overprinting the folds. In movement zones producing several generations of structures, transected folds may therefore simply be the result of the overprint of the first generation of folds formed in the movement zone by a younger cleavage. Williams (1985) presents a similar observation on the nature of the transected folds documented by Stringer (1975) and Borradaile (1978) for New Brunswick, Canada. He reports that the transecting cleavage is in fact axial planar to a later fold generation, and an earlier cleavage axial planar to the transected folds is also observed locally. Williams (1985) therefore concludes that the 'transected' folds are the result of the overprinting of first generation folds by a later generation of structures.

Structures with the geometry of transected folds are also reported near the Indian Island Fault in north central Newfoundland (Williams & Urai 1989, fig. 1). A cleavage fans around folded veins but cuts obliquely across the fold axes. This geometry can be described as Δ-type transection following Borradaile (1978). The fanning of the cleavage may indicate contemporaneity of folds and the cleavage. Several lines of evidence, however, indicate that the relationship is due to overprinting and that the cleavage is older than veins and folds. Folding of the veins on an axis slightly oblique to the cleavage–vein intersection lineation has produced a transected fold geometry.

Transected folds have been widely documented in movement zones in the Caledonides (Sanderson *et al.* 1980, Soper 1986, Soper *et al.* 1987, Woodcock *et al.* 1988) and in the Appalachians (Stringer 1975, Borradaile 1978, Malo 1986, Williams & Urai 1989). Movement zones are zones of intense deformation where multiple generations of structures and overprinting relationships should be expected. Movement zones may also be late in the structural history of an area, and consequently overprinting between structures related to the movement zones and structures related to the early history of the area should be expected. Where structures with the geometry of transected folds were carefully examined in the Canadian Appalachians, an overprinting relationship between the folds and the cleavage was observed. Although these examples do not negate the possibility that true cleavage transected folds may exist, the geometry of transected folds by itself should not be used to infer a sense of movement since this geometry may be an overprinting.

Structural history of the transpression zone

The transpression zone is divided into low-strain zones and high-strain zones. The high-strain zones are located at the contact between lithological units and terranes of contrasting competence. The structural history of the low-strain zones and the high-strain zones are similar, but since the deformation is concentrated in the high-strain zones, more complex overprintings are observed in these. The structural evolution of the high-strain zone on Swan Island is shown schematically in Fig. 8. Initial shortening across the transpression zone produced asymmetrical S-verging F_2 folds (Fig. 8a). F_2 folds are interpreted as the product of deformation in a transpression zone on the basis of: (1) their regional distribution and concentration in the high-strain zones; and (2) the contrast in style between F_2 folds and the large open regional folds affecting the Dunnage Zone south of the Bay of Exploits.

F_2 folds were then rotated in a clockwise sense prior to being folded by NE-plunging dextral F_3 folds and overprinted by the regional S_3 cleavage (Fig. 8b). As the beds were steepened, their orientation became favorable for bedding-parallel strike-slip movement. The deformation became localized in the less competent argillite-rich horizons which resulted in the formation of type 1 shear zones and mélanges (Fig. 8c), which form part of the Boones Point Complex. The presence of the regional cleavage S_3 in the mudstone matrix between the boudinaged beds in the mélanges indicates that shear zones formed progressively with F_3 folding. Deformation was thereafter concentrated within the shear zones with formation of steeply-dipping F_4 folds.

Overturning of bedding may have occurred either following or late during F_3 folding and faulting (type 1 shear zones). VM shear zones probably formed to accommodate vertical extension generally associated with compression across a transpression zone. A cleavage S_{3B} with the morphology and orientation typical of the regional cleavage S_3 is associated with the VM shear zones. Thus two cleavages with similar orientation and morphology were generated during the structural evolution of the transpression zone.

The structures in the transpression zone overprint the youngest formation in the area which is late Llandoverian in age, and are overprinted in turn by a late Silurian (age—G. R. Dunning personal communication 1985) granodioritic intrusion. This constrains the timing of deformation in the movement zone to a 15–20 Ma period following the deposition of the late Llandoverian formation. The transpression zone is therefore a Silurian movement zone.

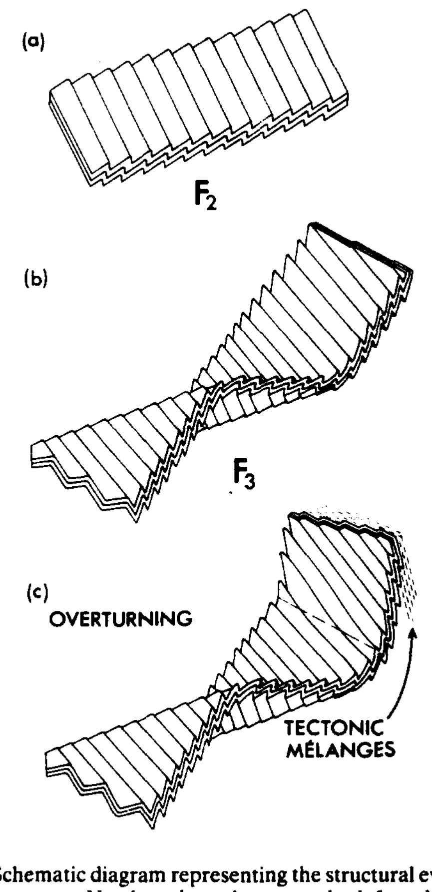

Fig. 8. Schematic diagram representing the structural evolution of the movement zone. North and south are on the left and right, respectively, of the diagram. (a) F_2 folds formed as a result of shortening across a N-dipping sequence. (b) Refolding of F_2 folds by NE-plunging dextral F_3 folds. (c) Overturning of the sequence. Tectonic mélanges and shear zones (represented by broken lines) formed during or closely following the steepening and overturning of the sequence.

CONCLUSIONS

The structural evolution of transpression zones in low-grade rocks has received considerable attention in recent years due to the recognition of the importance of transcurrent faulting in the tectonic history of an orogen. Structures in the Bay of Exploits, north central Newfoundland, were found to be the product of deformation in a dextral Silurian transpression zone. This study has focussed on the development of one specific transpression zone, but some of the conclusions reached may be of more general interest.

(1) The deformation in Notre Dame Bay transpression zone progressed from a regional ductile phase in low-strain zones to a more localized brittle–ductile phase in high-strain zones which resulted in the formation of mélanges. The deformation was complex and involved several generations of structures.

(2) Shear zones and mélanges formed during or closely following the steepening of bedding when the orientation of the beds became favorable for bedding-parallel strike-slip movement.

(3) Two cleavages, including the regional cleavage, having similar morphology and orientation with respect to bedding were generated during the structural evolution of the transpression zone.

(4) Transected folds may only be the result of the overprinting of the folds by a younger generation of cleavage.

Acknowledgements—I am grateful to Colleen Elliott, Patricia Flagler, Craig Place and Joseph White for their discussions and comments on this work. Sherri Townsend typed the manuscript and Angel Gomez drafted the figures. I would especially like to thank my supervisor, Paul Williams, for his encouragement and guidance. Funding was provided by an NSERC postgraduate scholarship, NSERC operating grant A8512, and EMR Research Agreement No. 71. This paper forms part of a Ph.D. dissertation which was undertaken at the University of New Brunswick.

REFERENCES

Antonuk, C. N. 1986. Geology, structure and microstructure of Farmer's Island and adjacent islands, north-central Newfoundland, with a discussion of the Dunnage mélange. Unpublished M.Sc. thesis, University of New Brunswick, Fredericton, Canada.

Arnott, R. J. 1983. Sedimentology of Upper Ordovician–Silurian sequences on New World Island, Newfoundland: separate fault-controlled basins? *Can. J. Earth Sci.* **20**, 345–354.

Arnott, R. J., McKerrow, W. S. & Cocks, L. R. M. 1985. The tectonics and depositional history of the Ordovician and Silurian rocks of Notre Dame Bay, Newfoundland. *Can. J. Earth Sci.* **22**, 607–618.

Blewett, R. S. & Pickering, K. T. 1988. Sinistral shear during Acadian deformation in north-central Newfoundland, based on transecting cleavage. *J. Struct. Geol.* **10**, 125–127.

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

Dean, P. L. 1978. The volcanic stratigraphy and metallogeny of Notre Dame Bay, Newfoundland. Memorial University of Newfoundland, Geology Report 7.

Dunning, G. R., Kean, B. F., Thurlow, J. G. & Swinden, H. W. 1987. Geochronology of the Buchans, Roberts Arm, and Victoria Lake groups and Mansfield Cove Complex, Newfoundland. *Can. J. Earth Sci.* **24**, 1175–1184.

Eastler, T. E. 1971. Geology of Silurian rocks, Change Islands and easternmost Notre Dame Bay, Newfoundland. Unpublished Ph.D. thesis, Columbia University, New York.

Elliott, C. G. 1985. Stratigraphy, structure, and timing of deformation of southwestern New World Island, Newfoundland. *Current Res. Geol. Surv. Pap. Can.* **85-1B**, 43–49.

Elliott, C. G. 1988. The depositional, intrusive and deformational history of southwest New World Island, and its bearing on orogenesis in central Newfoundland. Unpublished Ph.D. thesis, University of New Brunswick, Fredericton, Canada.

Elliott, C. G. & Williams, P. F. 1988. Sediment slump structures: a review of diagnostic criteria, and application to an example from Newfoundland. *J. Struct. Geol.* **10**, 171–182.

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

Helwig, J. A. 1967. Stratigraphy and structural history of the New Bay area, north central Newfoundland. Unpublished Ph.D. thesis, Columbia University, New York.

Helwig, J. A. 1970. Slump folds and early structures, Northeastern Newfoundland Appalachians. *J. Geol.* **78**, 172–187.

Heyl, G. R. 1936. Geology and mineral deposits of the Bay of Exploits area. *Newfoundland Dept. Nat. Res. Geol. Section Bull.* **3**.

Horne, G. S. 1968. Stratigraphy and structural geology of southwestern New World Island area, Newfoundland. Unpublished Ph.D. thesis, Columbia University, New York.

Horne, G. S. 1969. Early Ordovician chaotic deposits in the Central Volcanic Belt of northeastern Newfoundland. *Bull. geol. Soc. Am.* **80**, 2451–2464.

Jacobi, R. D. & Wasowski, J. J. 1985. Geochemistry and plate tectonic significance of the volcanic rocks of the Summerford Group, north-central Newfoundland. *Geology* **13**, 126–130.

Karlstrom, K. E., 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.

Kean, B. F., Dean, P. L. & Strong, D. F. 1981. Regional geology of the Central Volcanic Belt of Newfoundland. In: *The Buchans Orebodies: Fifty Years of Geology and Mining* (edited by Swanson, E. A., Strong, D. F. & Thurlow, J. G.). *Spec. Pap. geol. Ass. Can.* **22**, 65–78.

Knipe, R. J. & White, S. H. 1979. Deformation in low grade shear zones in the Old Red Sandstone, S.W. Wales. *J. Struct. Geol.* **1**, 53–66.

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

Malo, M. 1986. Stratigraphie et structure de l'anticlinorium d'Aroostook-Percé en Gaspésie, Québec. Unpublished Ph.D. thesis, Université de Montréal, Canada.

McKerrow, W. S. & Cocks, L. R. M. 1981. Stratigraphy of the eastern Bay of Exploits, Newfoundland. *Can. J. Earth Sci.* **18**, 751–764.

Means, W. D., Hobbs, B. E., Lister, G. S. & Williams, P. F. 1980. Vorticity and non-coaxiality in progressive deformations. *J. Struct. Geol.* **2**, 371–378.

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

Nelson, K. D. 1981. Mélange development in the Boones Point Complex, north-central Newfoundland. *Can. J. Earth Sci.* **18**, 433–442.

Neuman, R. B. 1976. Early Ordovician (Late Arenig) brachiopods from Virgin Arm, New World Island, Newfoundland. *Geol. Surv. Can.* **261**, 10–61.

Neuman, R. B. 1984. Geology and paleobiology of islands in the Ordovician Iapetus Ocean: review and implications. *Bull. geol. Soc. Am.* **95**, 1188–1201.

Nowlan, G. S. & Thurlow, J. G. 1984. Middle Ordovician conodonts from the Buchans Group, central Newfoundland, and their significance for regional stratigraphy of the Central Volcanic Belt. *Can. J. Earth Sci.* **21**, 284–296.

Pickering, K. T. 1987. Wet-sediment deformation in the Upper Ordovician Point Leamington Formation: an active thrust–imbricate system during sedimentation, Notre Dame Bay, north-central Newfoundland. In: *Deformation of Sediments and Sedimentary Rocks* (edited by Jones, M. E. & Preston, R. M. F.). *Spec. Publs geol. Soc.* **29**, 213–239.

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

Reusch, D. N. 1983. The New World Island Complex and its relationships to nearby formations, north-central Newfoundland. Unpublished M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, Canada.

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.

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

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

Soper, N. J., Webb, B. C. & Woodcock, N. H. 1987. Late Caledonian (Acadian) transpression in north-west England: timing, geometry and geotectonic significance. *Proc. Yorks. geol. Soc.* **46**, 175–192.

Stringer, P. 1975. Acadian slaty cleavage noncoplanar with fold axial surfaces in the Northern Appalachians. *Can. J. Earth Sci.* **12**, 949–961.

Stringer, P. & Treagus, J. E. 1981. Non-axial planar cleavage in the Hawick Rocks of the Galloway area, Southern Uplands, Scotland. *J. Struct. Geol.* **2**, 317–331.

Swinden, H. 1985. Geochemical setting of volcanogenic sulphide mineralization in the Wild Bight Group, Central Newfoundland. *Geol. Ass. Can. Fredericton Abs. w. Prog.* **10**, A60.

Strong, D. F. 1977. Volcanic regimes of the Newfoundland Appalachians. *Spec. Pap. geol. Ass. Can.* **16**, 61–90.

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

van der Pluijm, B. A. 1984. Geology and microstructures of eastern New World Island, Newfoundland and implications for the northern Appalachians. Unpublished Ph.D. thesis, University of New Brunswick, Fredericton, Canada.

van der Pluijm, B. A. 1986. Geology of eastern New World Island, Newfoundland: an accretionary terrane in the northeastern Appalachians. *Bull. geol. Soc. Am.* **97**, 932–945.

van der Pluijm, B. A. 1987. Timing and spatial distribution of the deformation in the Newfoundland Appalachians: a "multi-stage" collision history. *Tectonophysics* **135**, 15–24.

van der Pluijm, B. A., Karlstrom, K. E. & Williams, P. F. 1987. Fossil evidence for fault-derived stratigraphic repetition in the northeastern Newfoundland Appalachians. *Can. J. Earth Sci.* **24**, 2337–2350.

van der Pluijm, B. A. & van Staal, C. R. 1988. Characteristics and evolution of the Central Mobile Belt, Canadian Appalachians. *J. Geol.* **96**, 535–547.

Wasowski, J. J. & Jacobi, R. D. 1986. The tectonics and depositional history of the Ordovician and Silurian rocks of Notre Dame Bay, Newfoundland: Discussion. *Can. J. Earth Sci.* **23**, 583–586.

Williams, H. 1979. Appalachian orogen in Canada. *Can. J. Earth Sci.* **16**, 792–807.

Williams, P. F. 1976. Relationships between axial-plane folations and strain. *Tectonophysics* **22**, 311–324.

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

Williams, P. F., Elliott, C. G. & Lafrance, B. 1988. Trip B2. Structural geology and mélanges of eastern Notre Dame Bay, Newfoundland. Geol. Ass. of Canada, Memorial University of Canada, St. John's, 1988.

Williams, P. F. & Urai, J. L. 1989. Curved vein fibres—an alternative explanation. *Tectonophysics* **158**, 311–333.

Woodcock, N. H., Awan, M. A., Johnson, T. E., Mackie, A. H. & Smith, R. D. A. 1988. Acadian tectonics of Wales during Avalonia/Laurentia convergence. *Tectonics* **7**, 483–495.