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Brittle faulting in the Canadian Appalachians and the interpretation of reflection seismic data

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Abstract—Brittle faults in the Canadian Appalachians, mostly Late Devonian or younger in age, are divided into two groups on the basis of their movement vectors: a strike-slip group and a dip-slip group. The two groups are subdivided on the basis of orientation and sense of movement. They are interpreted in terms of Palaeozoic transpression and Mesozoic extension during Atlantic opening. The Palaeozoic faults comprise strike-slip and reverse faults which were active in Devono-Carboniferous times. Mesozoic strike-slip faults are interpreted as transfer faults and they are coupled with normal faults, many of which developed by reactivation of earlier reverse faults or steeply-dipping surfaces including bedding and earlier foliations. Together these Mesozoic faults comprise an adequate mechanism for crustal extension perpendicular to the orogen. Evidence for listric normal faults is rare in the older rocks of the orogen, and it is suggested that this is due to the adequacy of the deformation mechanisms afforded by the pre-existing planes of weakness.

The data presented demonstrate clearly that geological structures are commonly repeated at all scales from outcrop to regional. Following this principle, prominent shallowly-dipping reflectors imaged in the Lithoprobe East seismic profiles are interpreted as large-scale representatives of the Palaeozoic reverse faults, seen in outcrop, that were reactivated as normal faults during Atlantic opening. It is suggested that reverse faulting is to be expected as a deformation mechanism in general in the late stages of a collisional orogen.

We draw attention to the importance of the Mesozoic faults in modifying the attitude of earlier surfaces such a bedding, foliations and faults.

INTRODUCTION

The tectonic history of the Appalachian orogen is complex and is essentially one of several independent collisions between the pre-Palaeozoic North America, island arcs and parts of Gondwana during the Palaeozoic (Colman-Sadd *et al.* 1992, van Staal in press). Thrusting was an integral part of the various collisions that took place in the Ordovician and Silurian although convergence was oblique and resulted in significant orogen parallel motion (van Staal & Williams 1988, Williams *et al.* 1988, Lafrance & Williams 1992). Deformation continued into the Devonian and Carboniferous by which time in the northern Appalachians it was for the most part markedly transpressive (Hanmer 1981, Malo *et al.* 1992, Hibbard & Hall 1993). During the Mesozoic extension gave rise to the present Atlantic Ocean, essentially parallel to the Palaeozoic orogen (Tankard & Balkwill 1989 and other articles in the same volume).

Here we are concerned with the Canadian Appalachians and the story is essentially one of: (1) compressional and transcurrent faulting related to the late stages of the Palaeozoic shortening episode, and (2) extensional and transfer faulting related to the Mesozoic extensional episode. Many of the faults active in the Mesozoic were initiated in the Palaeozoic and simply reactivated during extension.

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The Canadian Appalachians are divided into a series of tectono-stratigraphic terranes which are elongate parallel to the orogen. The major terranes are shown in Fig. 1. The westernmost of these, the Humber terrane, represents the eastern margin of early Palaeozoic North America. The other terranes were accreted to it during the Palaeozoic. The boundaries of the terranes are composite structures formed during a long history of development and reactivation in which they were active both as shear zones and brittle faults. Sympathetic shear zones and faults are developed within the terranes, locally defining smaller terranes (e.g. Williams *et al.* 1988a, Williams *et al.* 1988b, Lafrance & Williams 1992), and the region is characterized by a lenticular disposition of rock units.

Continuity both across and along strike seems to be the exception rather than the rule. On a geological map, pre-Devonian rocks occur as lenses elongate parallel to the orogen. On the ground the same is commonly true at outcrop scale, and bedding and foliations are commonly steep. There are many major shear zones, many of which are steep. In contrast, Lithoprobe East seismic data reveal a structure dominated by shallowly-dipping reflectors (Fig. 2). These features generally have reasonable continuity across strike and one in particular is interpreted as having considerable continuity (Hall *et al.* 1990).

We suggest that many of the reflectors are late faults which cut across the earlier lenticular fabric. The latter is largely a pre-Late Devonian phenomenon that resulted from the oblique collision manifest as early ductile

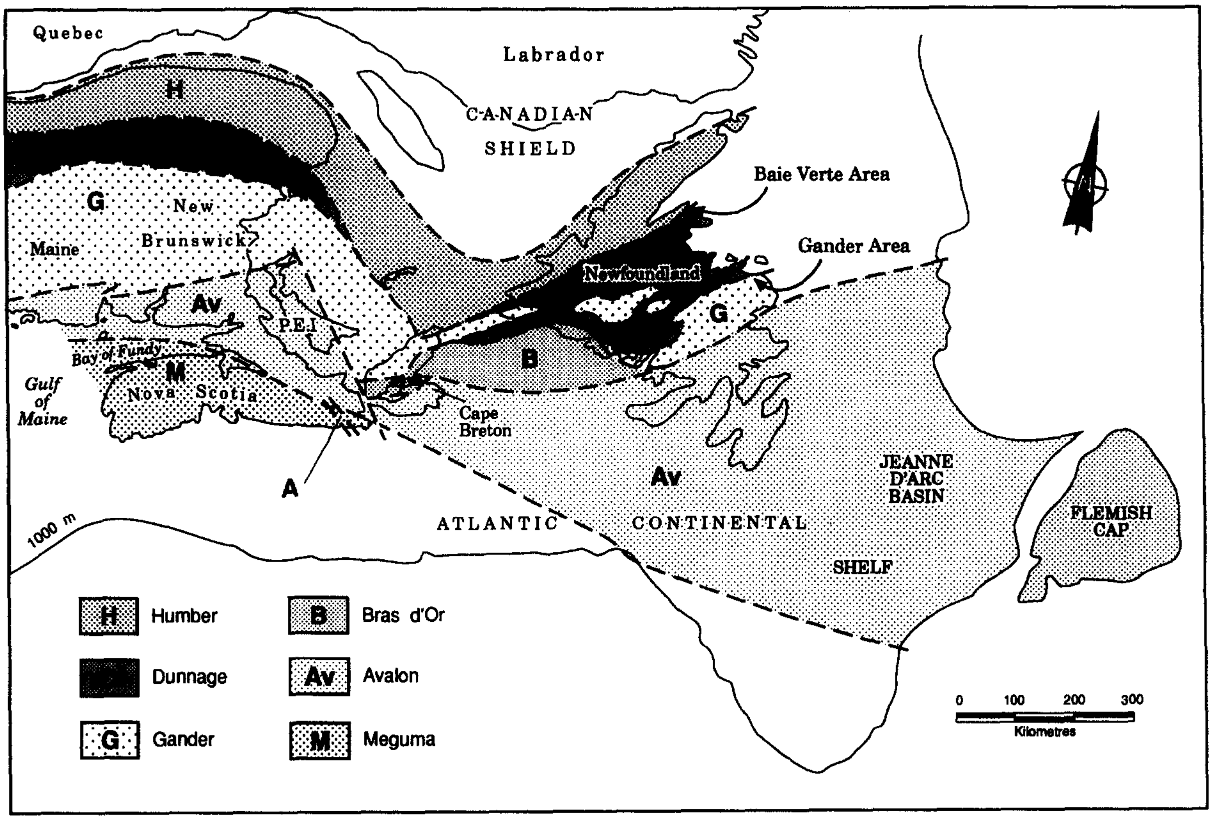


Fig. 1. Terrane map modified from Williams & Hatcher (1983), Barr & Raeside (1989) and Colman-Sadd *et al.* (1992). 'A' indicates the mega-kink represented in Fig. 8.

thrusting and later ductile transcurrent movement. The transcurrent movements continued into Carboniferous times but became progressively more brittle and less pervasive. Since the faults discussed here overprint the lenticular fabric, they are no earlier than Late Devonian. They would be best studied therefore in rocks of that age or younger. However, much of our data comes from older rocks where we can recognize the faults as young by overprinting relationships but cannot place an upper limit on their age.

Some examples of these faults are noted on existing maps and have been reported by various writers (e.g. Hatcher *et al.* 1988). They may, however, be more significant than generally acknowledged by workers studying Palaeozoic rocks. In this paper we draw attention to the significance of the faults, discuss their history and present a regional synthesis. Finally, we draw attention to the potentially profound significance of these faults for interpretation of northern Appalachian structure and, more specifically, for interpretation of the Lithoprobe East seismic data.

The transcurrent faults occur in many orientations and represent more than one period of deformation. In the following we divide them into two groups partly on the basis of orientation and divide dip-slip faults into reverse and normal.

TRANSCURRENT FAULTS: GROUP 1

The first group comprises brittle faults that follow earlier ductile structures, for example, the Belleisle Fault (see Fig. 3 for this and other named structures and places) in New Brunswick (e.g. Webb 1969, Bradley 1982, Leger 1986, Leger & Williams 1986), the Cobequid-Chedabucto Fault in Nova Scotia (Mawer & White 1987) and the Chanceport Fault in Newfoundland (Williams *et al.* 1988b, Lafrance 1990). They are generally moderately to steeply-dipping and have trends in the northeast quadrant that tend to group around northeast and east. Figures 4(a) & (b) include data for small scale examples from Newfoundland. Some, if not all, of the Group 1 faults had dip-slip components and were active in Devonian and later times (e.g. Gussow 1953, Belt 1968, Webb 1969, Ludman 1981, Bradley 1982, Johnson & Wones 1984, Leger & Williams 1986, Williams *et al.* 1988b, Lafrance 1990). Leger (1986) presented slickenside and fracture pattern evidence for a dextral transcurrent component of movement on NE-trending faults in southern New Brunswick as well as a significant dip-slip component in post Late-Devonian or Carboniferous times. Yeo & Gao (1987) demonstrated a complex Carboniferous history of transcurrent movement on the Hollow Fault of Nova Scotia and there is the possibility of still later reactivation, for example during opening of the Atlantic Ocean (cf. Plint & van de Poll 1984).

TRANSCURRENT FAULTS: GROUP 2

Faults belonging to the second group have trends in the northwest quadrant with the most common orientation being approximately northwest, except in parts of

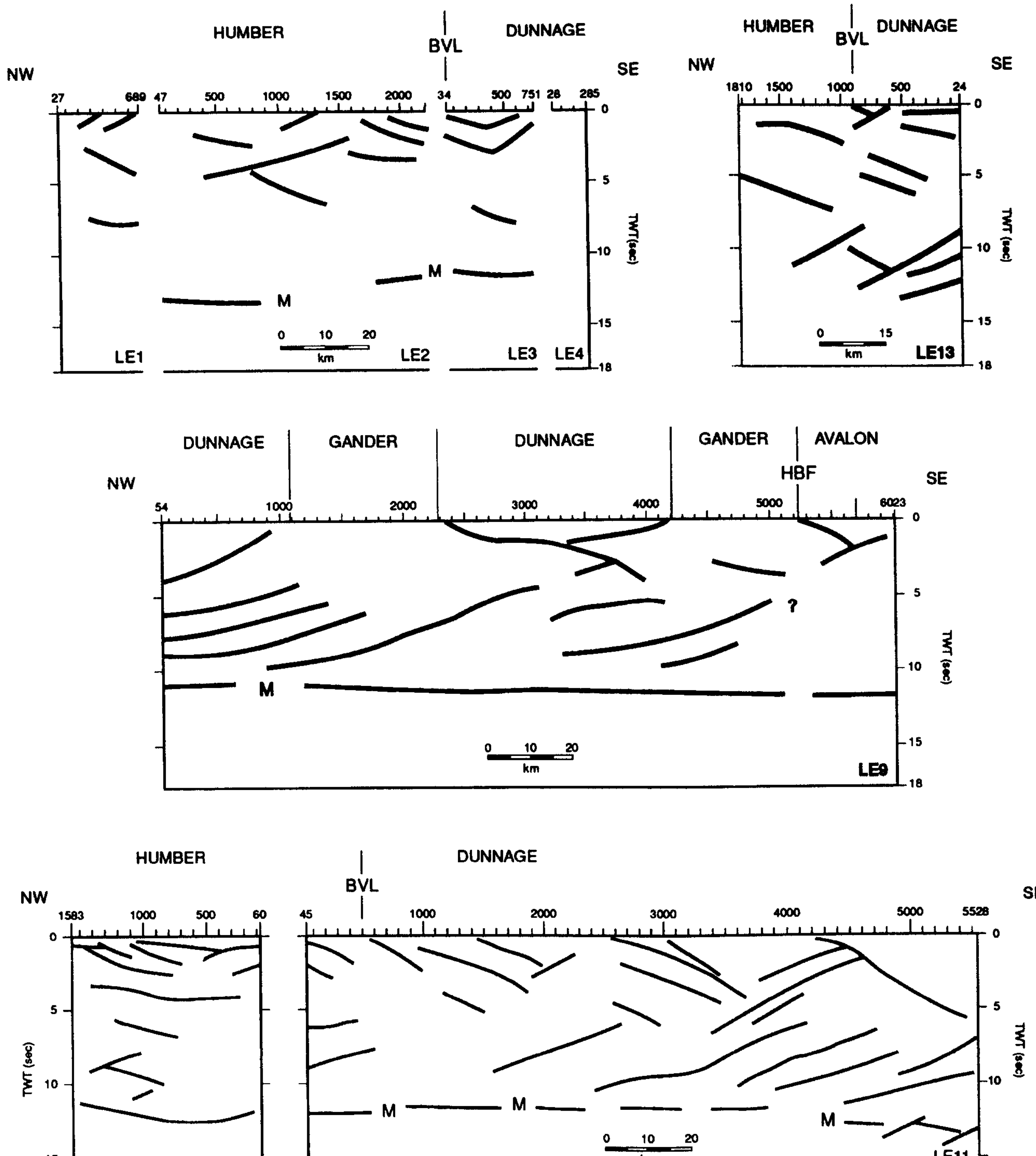


Fig. 2. Seismic profiles as interpreted by Quinlan *et al.* (1991) from Lithoprobe East (LE) reflection seismic survey data. LE numbers correspond to Lithoprobe East transect line numbers in Fig. 3. BVL and HBF show intersections of Baie Verte Line and Hermitage Bay Fault respectively with the transect lines. Terrane names as in Fig. 1.

Newfoundland where a W-trend is equally common (e.g data included in Figs. 4a & b). Faults of this group are generally younger than those described above and we interpret group 2 as representing a single generation of structures. Since Groups 1 and 2 overlap in orientation it is difficult or impossible to assign W-trending members to the appropriate groups on the basis of orientation alone. Somewhat arbitrarily, faults that are spatially associated with older shear zones of the same orientations are tentatively assigned to the first group and the others are assigned to the second.

Group 2 faults have been interpreted as transfer faults associated with the opening of the Atlantic (Leger & Williams 1986, Williams *et al.* 1988b, Williams & Hy 1990, Miller 1990). We present further evidence here in support of this interpretation and suggest that the faults operated as part of a larger system which is discussed later in the text.

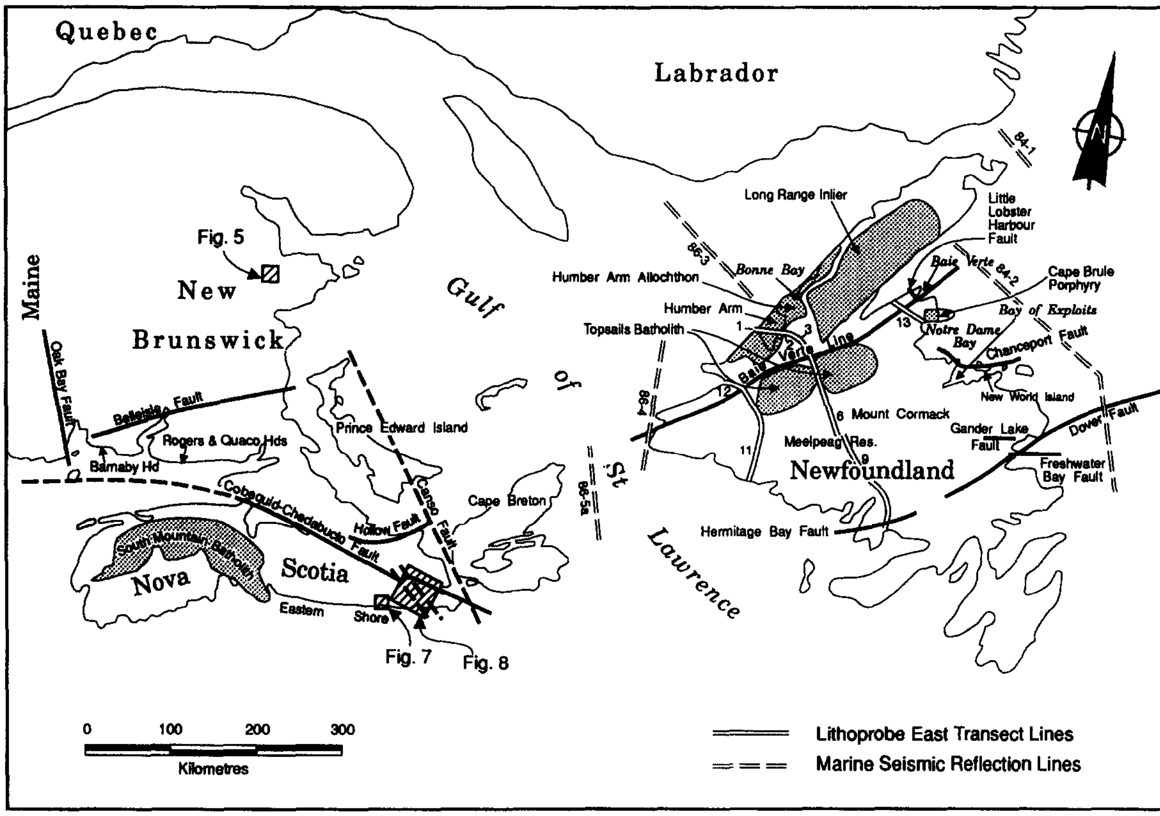


Fig. 3. Locality map showing Marine Seismic and Lithoprobe transect lines and places and features referred to in the text. Some geological features are stippled for clarity, but no correlation or similarity is implied.

Field occurrence

NW-trending group 2 faults are prominent from Maine to Newfoundland but are mostly recognized as breaks in outcrop and as lineaments on aerial photographs. In the Meguma Terrane of Nova Scotia they are so prominent as topographic features that their trend is readily recognized from a road map. In the structurally more complex terranes they are less obvious but nevertheless recognizable as lineaments.

In the Meguma Terrane, the lineaments trend approximately NW. Elsewhere they vary considerably in trend (approximately between west and north) which might be interpreted as indicating the presence of more than one generation of faults. However, satellite-image-analysis of part of the Bathurst mining camp, New Brunswick (Fig. 5) leads us to interpret the group as a single generation for the following reasons: (1) The faults can be sinuous with individual lineaments varying in trend by almost 45° (e.g. *X* and *Y* in Fig. 5), and there is a continual graduation from one extreme to the other. (2) There is overprinting between representatives of the different trends, but no consistent pattern. (3) Similar faults in Mesozoic rocks on the continental shelf east of Newfoundland, interpreted as transfer faults (Tankard & Welsink 1987), are normal to the ocean–continent boundary. Since the generalized boundary is curved approximately 90° (Fig. 1), the transfer faults fan sympathetically through a similar angle (Tankard & Welsink 1987), showing approximately the same variation as the faults in the Bathurst camp.

The same variation in trend is observed outside of the Bathurst camp, but the continuity of segments of different trend has not generally been recognized (we know of one example from New World Island). In the Dunnage Zone, for instance, both W-trending (some even slightly south of west) and more NW-trending faults are common but generally occur as discrete lineaments. Nevertheless, while keeping in mind the possibility of diachroneity, we interpret the two as coeval wherever the faults cut all other structures.

In continuous coastal exposure small-scale examples of the faults can be seen in their entirety and broader, longer ones are locally represented by slickensides on one or both sides of a break in outcrop. Slickensides striae are predominantly shallowly plunging but locally are steeply plunging; associated quartz fibres may vary continuously in orientation through as much as 90° from horizontal to vertical. Some of the best examples of these faults are seen in the Meguma Terrane (Williams & Hy 1990) where NW-trending faults are parallel or approximately parallel to *ac* joints (*ac* joints are perpendicular to the axis of regional folds) and can only be distinguished from the latter by virtue of the offset associated with them. The faults vary in morphology from arrays of vertical tension gashes, locally associated with kinkbands (Figs. 6a–c), through linked tension gashes (Fig. 6d), to simple vertical shear-surfaces (Fig. 6e). Locally, the displacement is more distributed and the faults are narrow vertical zones of spaced discontinuities and associated kinks and kinkbands (Fig. 6f). Here and elsewhere in Atlantic Canada the faults are

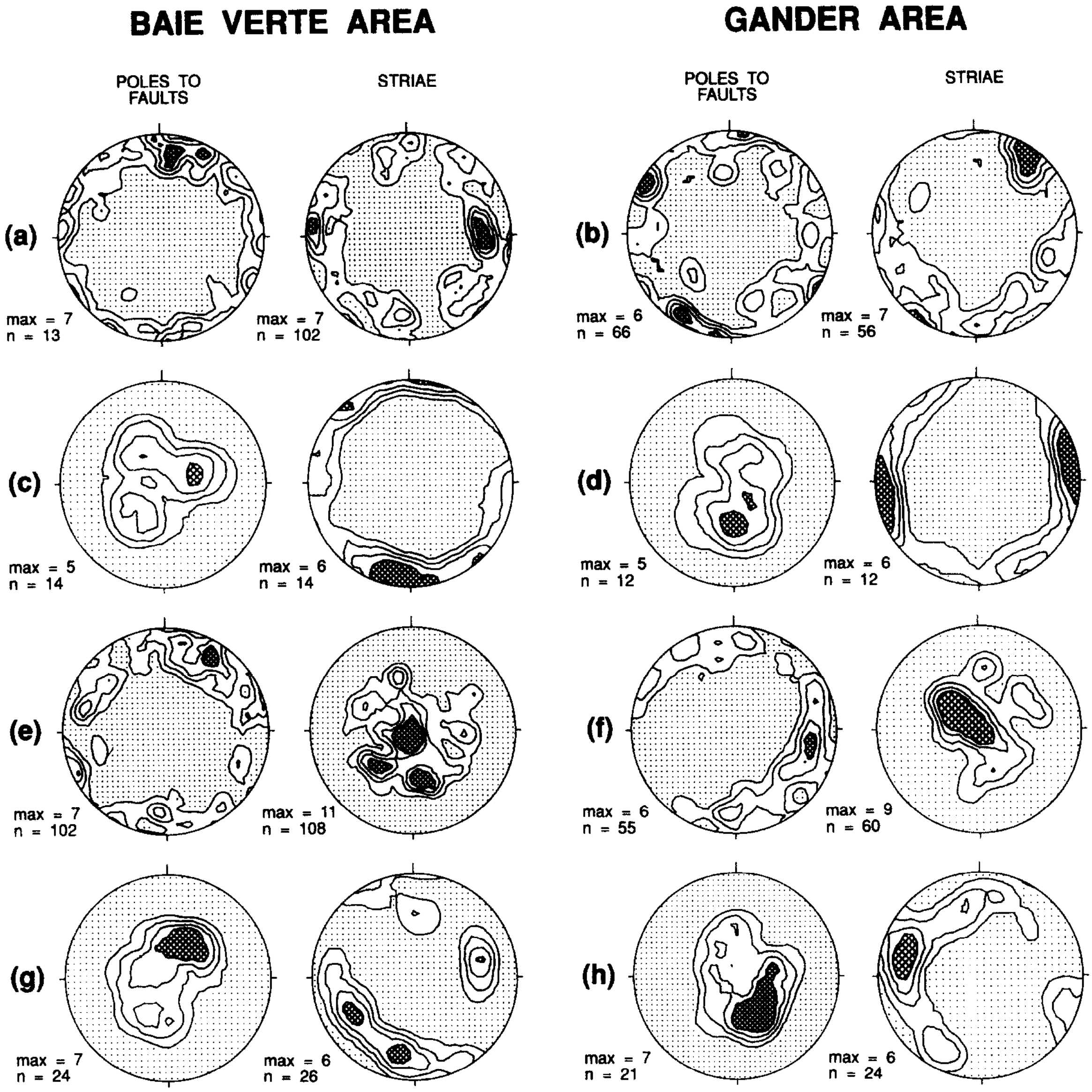


Fig. 4. Lower hemisphere equal area plots of poles to mesoscopic faults and slickenside striae. Measurements from along and adjacent to the Humber/Dunnage Zone boundary (Baie Verte Line) in the Baie Verte area and the northernmost segment of the Dunnage/Gander Zone boundary (Gander area) in Newfoundland (see Fig. 1 for locations). In general the sense of movement on these faults cannot be determined; only faults for which slickenside striae give the direction of movement are recorded. The data are plotted and contoured using the program and method of Starkey (1977) and are subdivided according to the orientation of the fault and the striae. If the pitch of the striae is <45° the fault is plotted as transcurrent and if >45° as dip-slip. In the Baie Verte area the faults locally cut Carboniferous rocks. (a)–(d) Transcurrent faults. (e)–(h) Dip-slip faults. (a) Steep transcurrent faults are predominantly W-striking Group 2 faults. Lesser concentrations with N–NE strikes and NW strikes are interpreted as Group 1 and 2, respectively. Group 1 faults are parallel to the terrane boundary. (b) NE-striking Group 1 transcurrent faults parallel to the terrane boundary and W- to NW-striking Group 2 transcurrent faults. (c) Shallowly dipping transcurrent faults with north–northeast movement directions approximately parallel to the Baie Verte Line are interpreted as flower structures associated with the Baie Verte Line that have taken up the transcurrent motion. (d) W-striking shallowly dipping transcurrent faults are interpreted as flower structures related to Group 2 transcurrent faults. (e) & (g) Dip-slip faults mostly have NW strikes and are therefore assumed to be related to the Group 2 transcurrent faults as flower structures. This interpretation fits in well with the fact that Group 2 faulting is strongly developed in the area [see (a)]. A lesser concentration of NE-striking faults is interpreted as flower structures associated with Group 1 transcurrent faults parallel to the terrane boundary. (f) & (h) Dip-slip faults with NE strikes are interpreted as flower structures associated with Group 1 transcurrent faults parallel to the terrane boundary.

commonly characterized by zones of more or less *in situ* breccia cemented by quartz which may be sulphide-bearing. Wherever the movement direction can be determined they are dominantly transcurrent.

Both the W- and NW-trending faults can be recognized as discontinuities on aeromagnetic total field and gravity maps (Miller 1987, 1988, 1990, Sean McDonald personal communication 1990). A clear example can be seen in the Gulf of St Lawrence where such a magnetic discontinuity (Geological Survey of Canada, Map NK/NL20-AM) is continuous with the Canso Fault, which forms the topographic lineament separating Cape Breton Island from mainland Nova Scotia. A very prominent NW-trending magnetic discontinuity interpreted as a fault traverses much of Newfoundland and is more or less coincident with the Lithoprobe East Transect Lines

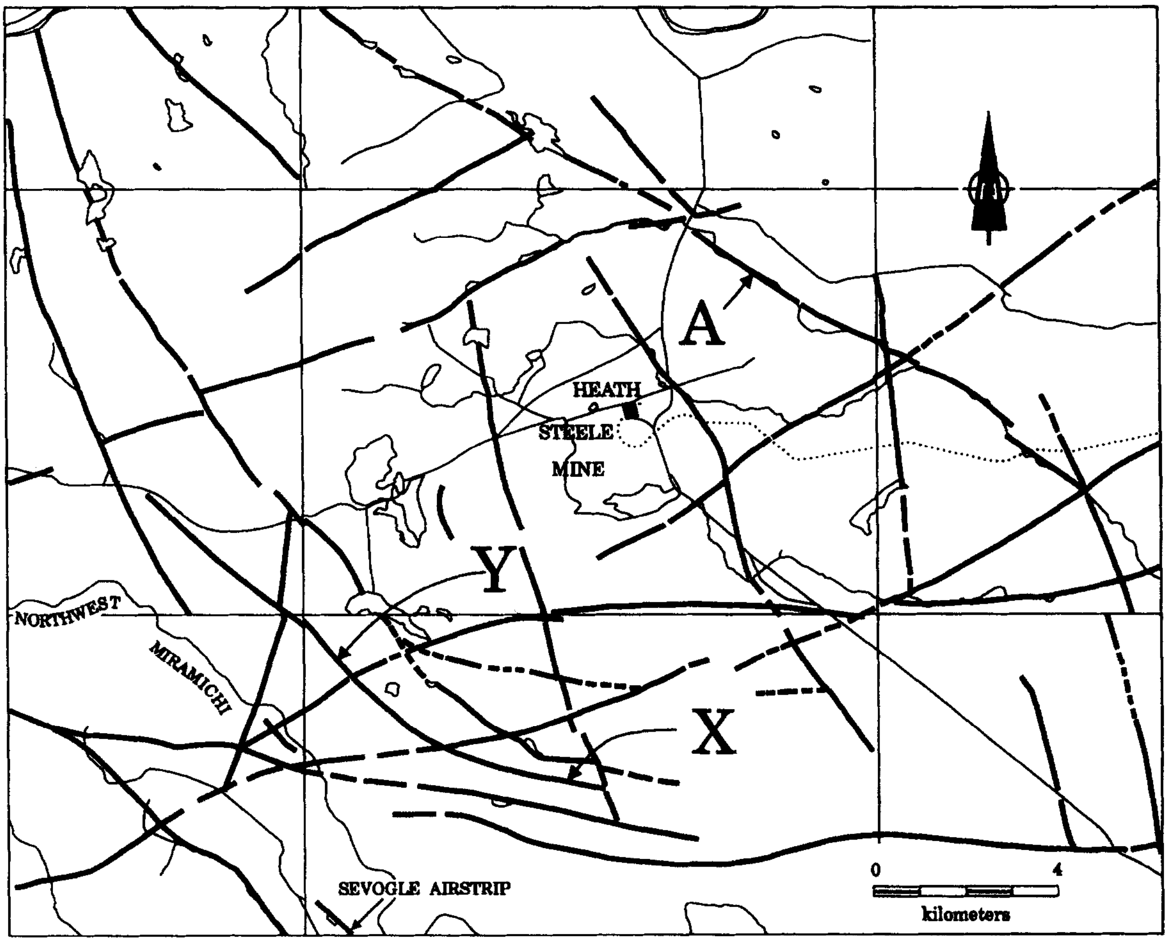


Fig. 5. Map of the Heath Steele Mines area (Bathurst mining camp) showing major lineaments extracted from Landsat MSS and TM data. (Image dates: MSS—October 1975 & TM—June 1987). After map prepared by J.G. Torrance (in Williams & McAllister 1989). Note the curvature in the lineaments (e.g. between *X* and *Y*) and the inconsistency in overprinting relationships between lineaments of various orientations. For location of map see Fig. 3.

3–9 (Fig. 3) (Miller personal communication 1990). This same feature coincides with a lineament visible on the satellite image and also on the geological map of Newfoundland (cf. Colman-Sadd *et al.* 1990). The northwestern end of this lineament is described as the Bonne Bay cross-strike discontinuity by Cawood & Botsford (1991); they interpret it as a transfer fault associated with rifting at the start of the Appalachian orogenic cycle.

The most spectacular geophysical expression of the NW-trending faults is seen on the aeromagnetic vertical gradient maps of Nova Scotia (e.g. Fig. 7 and Country Harbour map, Geological Survey of Canada, 1985). Some faults are sharply defined single features whereas others are broad and kinkband-like, which is precisely the variation recorded in the mesoscopic faults described above (Fig. 6). Figure 7 shows the Indian Harbour Fault which combines a sharp brittle feature to the north with a kinkband-like structure in the south. An even larger structure with a similar origin is seen on the tectonic map of Nova Scotia (Keppie 1982) where the southern edge of the St Marys Graben describes a kinkband-like structure (Figs. 1 and 8). Aeromagnetic vertical gradient trends in the Meguma rocks clearly follow the Carboniferous boundary. The boundaries of the kinkband coincide with the NW-striking Country Harbour and the New Harbour faults (Fig. 8). Both faults can be traced as lineaments on Airborne Radar images (unreferenced sample plate produced by EMR Canada combining radar images with a geological map) well into the Carboniferous, and the fault along the margin of St Marys Graben is overprinted by the kinkband. Thus the whole of the Cobequid-Chedabucto fault zone and the Carboniferous graben are overprinted by the NW-striking faults on a regional scale.

Sense and magnitude of movement

NW-trending faults in the Gulf of Maine have large horizontal separations (Hutchinson *et al.* 1988). They are sinistral and measurable in tens of kilometres at the northern end of the gulf and dextral with similar magnitude at the southern end of the gulf.

In New Brunswick, displacement magnitudes vary considerably, but for the most part horizontal separations are small relative to the length of the faults. In northern New Brunswick, one NW-trending lineament (A in Fig. 5.), which follows sections of the Tomogonops and Upsalquitch Rivers, can be traced for at least 120 km. The regional geological map (Davies 1977) shows no perceptible displacement across the lineament over most of its length. Locally it is shown as a fault with an

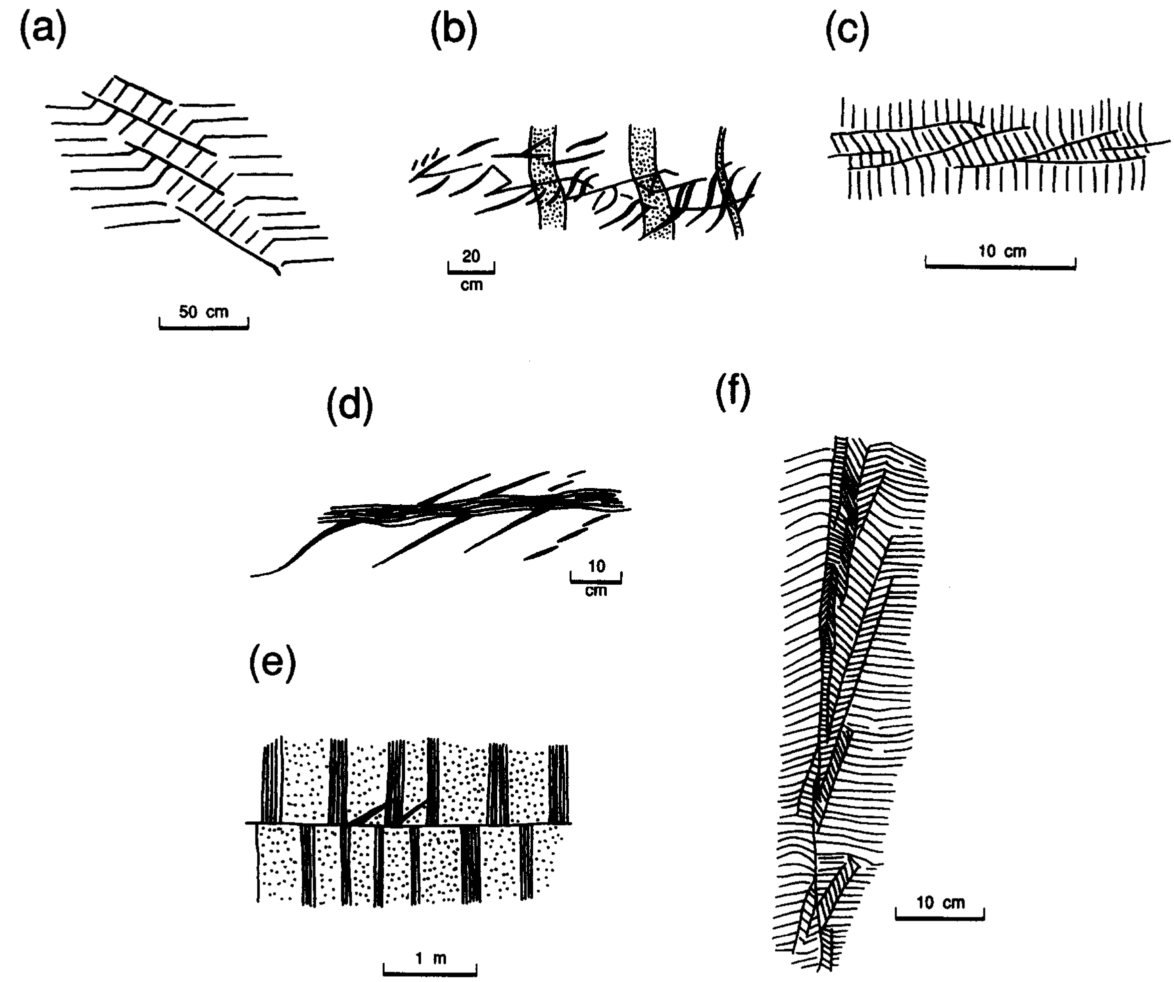


Fig. 6. Sketches of small-scale NW-striking transcurrent structures in turbidites from the Meguma Terrane, Nova Scotia. All are viewed on approximately horizontal surfaces. (a)–(c) Incipient faults with kinkband-like form and associated Riedel faults. Note that in (b) there is the beginning of a through-going fracture. (d) Fault sinistrally displacing initial Riedel faults; both are filled by vein quartz. (e) Steeply-dipping bedding displaced by a vertical fault. Two truncated quartz veins dipping to the left suggest sinistral offset as in (d). (a)–(d) are all small from coastal exposures on the Eastern Shore. (f) Small-scale fault and associated kinkbands from a large-scale fault zone at Forest Hill.

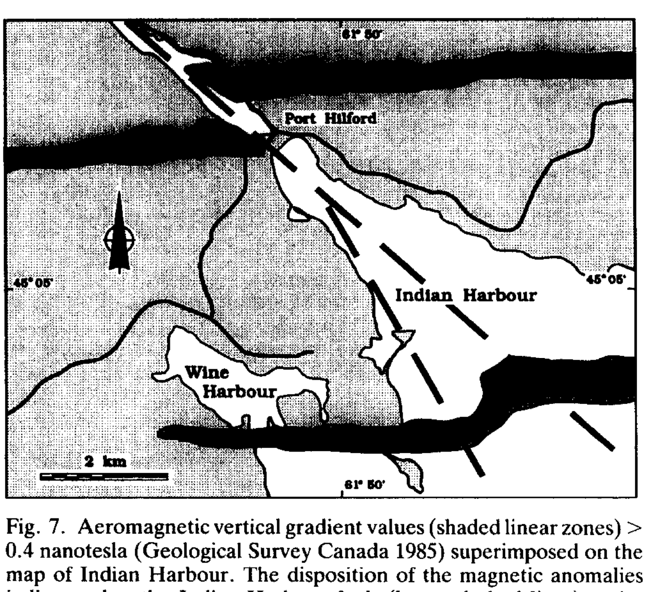


Fig. 7. Aeromagnetic vertical gradient values (shaded linear zones) > 0.4 nanotesla (Geological Survey Canada 1985) superimposed on the map of Indian Harbour. The disposition of the magnetic anomalies indicates that the Indian Harbour fault (heavy dashed lines) varies from a sharp feature in the north to a kinkband in the south. For location of map see Fig. 3.

apparently large but indeterminate displacement. This large displacement, however, is not supported by recent mapping by Cees van Staal (personal communication 1991). The lineament-forming structure was at one time exposed in a trench close to the Heath Steele Mines and the observed horizontal separation was only several metres (A. L. McAllister personal communication 1989). In contrast, the sinistral Oak Bay Fault in southern New Brunswick has a separation of about 50 km (Stringer & Burke 1985, McCutcheon & Robinson 1987).

Well constrained examples of the faults in the Meguma are mostly, but not exclusively, sinistral (e.g. Hwang 1990). Some of the macroscopic faults show a reversal in movement along their length (Williams & Hy 1990). Horizontal separation and fault length vary from centimetres to kilometres. On the Eastern Shore there is a fault with a horizontal separation of hundreds of metres at least every 10 km along the strike of the orogen (Williams & Hy 1990).

The controversial Canso Fault (Fig. 3) is said to have a minimum dextral horizontal separation in excess of 30 km and possibly as much as 200 km (McCutcheon &

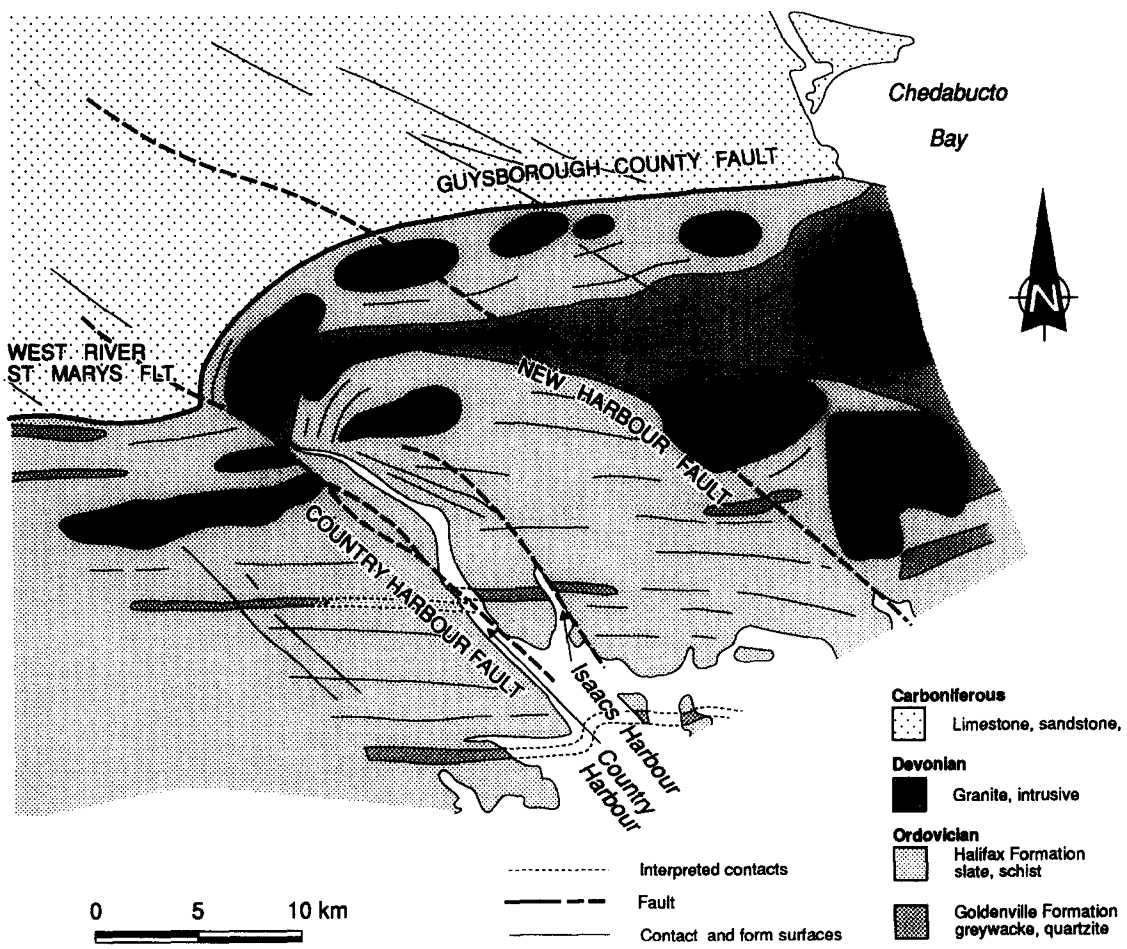


Fig. 8. Map of County Harbour showing a mega kinkband overprinting transposed layering and foliation in the Cobequid-Chedabucto shear zone. The shear zone is several kilometres wide in the area and includes all of the Devonian and Ordovician rocks shown. The kinkband also overprints the steep faults forming the edge of St Marys Graben in which the Carboniferous sediments occur. It is therefore younger than the Carboniferous sediments. Interpretive contacts are based on the vertical gradient trends (Geological Survey of Canada 1985). For location of map see Fig. 3.

Robinson 1987), but this rather tenuous interpretation is based on correlation of earlier shear zones across the fault. The picture is complicated by the proximity of the W-trending Cobequid-Chedabucto Fault, which may be responsible for some of the geological mismatch. Marillier *et al.* (1989) have pointed out that a NW-trending fault with a large displacement as proposed for the Canso Fault is not supported by their refraction seismic data. The latter show more or less continuous trends to the NE of Prince Edward Island across the extended line of the fault. However, the fault line is marked by a discontinuity in magnetic trends and it may be that its displacement in the area examined by Marillier *et al.* (1989) is too small to be resolved by their data. If this means a rapid change in the magnitude of the displacement along the length of the fault, it should be noted that such a change would be quite normal if it is a transfer fault (as described below). Thus it is possible that locally the Canso Fault is a major structure.

On New World Island in Newfoundland, NW-trending macroscopic faults are predominantly dextral and separations are small (tens of metres). W-trending faults in the Gander (Goodwin & O'Neill 1991) and Dover (Caron personal communication 1990) areas are also mainly dextral. A fault along Gander Lake displaces the Dunnage-Gander zone boundary approximately 5 km and the Dover Fault (the Gander-Avalon Zone boundary) is displaced approximately 3 km by a similar fault in Freshwater Bay (Caron personal communication 1990). Similarly, Miller *et al.* (1990) report W-trending dextral faults with horizontal separations up to 5 km in the Carboniferous of western Newfoundland.

In the Jeanne d'Arc basin of the Grand Banks, transfer faults strike northwesterly and are primarily dextral although horizontal separations of kilometres to tens of kilometres are both dextral and sinistral (Tankard & Welsink 1987).

Age of faults

The persistence of the fault-lineaments for tens of kilometres, in some cases crossing tectono-stratigraphic terrane boundaries, indicates that they are relatively young features. Wherever the faults are seen in outcrop they overprint all other structures. The faults cut the youngest rocks in the areas wherever they occur; they are seen in Devonian granitoids in New Brunswick (Cumming 1966, Hay 1967) and Nova Scotia (e.g. Horne *et al.* 1988, Williams & Hy 1990) and can be traced into the Carboniferous strata (e.g. Miller *et al.*

1990) in New Brunswick and Nova Scotia. They are known to cut Mesozoic rocks on the continental shelf (e.g. Tankard & Welsink 1987) and in the Gulf of Maine (Hutchinson *et al.* 1988). Significantly, they cut Triassic dykes but are also cut by them, indicating that they were active in the Triassic (Burke *et al.* 1973, Stringer & Burke 1985). Stringer & Burke (1985) suggested that one of these faults, the Oak Bay Fault, had a long history beginning in Devonian times, but Williams & Hy (1990) have suggested an alternative explanation and interpret it as a Mesozoic fault.

In Nova Scotia, veins and dykes of Devonian age have NW trends. This might also be interpreted as indicating an earlier age for the faults if it is assumed that the veins and dykes intruded along planes of weakness provided by the faults. However, although they share a common orientation and some of the veins are coincident with fault planes, the dykes and veins are generally simple tensional features with no apparent offset. This suggests that the age of the dykes and veins is different from the age of the faults since coincidence of orientation of tensional features and shear fractures would not be expected if they were coeval. Thus Williams & Hy (1990) interpret the veins as being related to earlier (Late Silurian or Early Devonian) *ac* joints, some of which were subsequently (Triassic or Jurassic) reactivated as faults. This same interpretation can be applied to the dykes and is supported by the observation that one of the dykes is locally cut by a NW-trending fault (Ruffman & Greenough 1990). This also explains the NW-trending segments of the contact of the 370–360 Ma old South Mountain batholith in Nova Scotia (cf. Horne *et al.* 1988) which is younger than the *ac* joints. Thus the evidence is consistent with a Mesozoic age for the faults.

Cawood & Botsford (1991) interpret the Bonne Bay cross-strike discontinuity as a product of deformation associated with rifting in the Early Cambrian at the start of the Appalachian orogenic cycle. Their interpretation is based on differences in stratigraphy and structural vergence across the fault. They suggest that a fault-formed topographic feature influenced sediments and therefore stratigraphy. Structural differences are interpreted in terms of repeated reactivation of the fault. It is not clear to us that the alternative possibility of stratigraphic and structural differences due to post-sedimentation folding and faulting has been adequately dismissed; we would like to see more conclusive structural evidence for the age of the fault. However, the foreland as represented by the Humber Zone Grenvillian basement is the one place where such an early fault might be preserved and reactivated. Elsewhere in the orogen later deformation is too intense for preservation to be likely, especially where transcurrent motion is involved. If the Bonne Bay structure is early it is possible that it was reactivated during the Mesozoic in the Humber Zone and spread from there across the rest of Newfoundland.

In summary then, the second group of transcurrent faults comprises predominantly NW- and W-trending structures of probable Mesozoic age, though there may be some earlier structures with similar orientations. The large faults rarely outcrop and are recognized mainly as narrow gaps in coastal exposure, as lineaments on aerial photographs and on aeromagnetic maps. Locally, the NW-striking faults are shown to have reactivated earlier *ac* joints and this may be generally true. These NW-striking faults are mostly sinistral and the W-striking faults are mostly dextral, but both vary and the sense of displacement can vary even within a single fault.

DIP-SLIP FAULTS

Dip-slip faults are common in the Canadian Appalachians and they can be demonstrated to have formed over a long period (cf. Hatcher *et al.* 1988). Because many of the faults do not outcrop and many are parallel to the general trend of the orogen, it is difficult to group them into generations. Our observations are mostly of small examples that are largely or completely contained within an outcrop. Despite these problems, there is a pattern and it is possible to develop a consistent working hypothesis which we present here. There are many faults for which we cannot determine the movement sense, but here we are concerned primarily with those with demonstrable slip-senses.

Reverse faults

Reverse faults occur throughout the region as essentially planar, brittle structures that overprint the earlier ductile structures and commonly occur in conjugate pairs. They may not all belong to the same period of deformation, but examples known to us are mostly consistent with an approximately W–NW shortening direction of Carboniferous or younger age (see Leger 1986). It is possible that many are related to Group 1 Carboniferous transcurrent faults as positive flower structures (Leger 1986, Nance 1987).

The examples that we are familiar with in Nova Scotia and Newfoundland are all small separation faults (e.g. Fig. 9). It is possible that larger structures do occur, but they are generally indistinguishable from other types of faults with the same trend.

In Newfoundland, the faults commonly cut the ductile shear zones in Bay of Exploits (e.g. Figs. 9c & d) and are oriented such that the horizontal intersection axis of conjugate pairs lies in the shear zone foliation. In some areas there are also conjugate faults with steeply plunging axes (see van der Pluijm *et al.* 1987), suggesting that they may be related to late transcurrent movement. Late ductile movement in the shear zones is believed to have occurred in the Late Silurian (Lafrance 1990), and the latest brittle movement could be of the same age or younger (Elliott 1988). The trend of the intersection axes, being dependent on the shear zone orientation, varies considerably, but generally lies in the northeast quadrant. Similar structures with a N-strike are reported from the Humber Arm allochthon of western Newfoundland (Bosworth 1985) where they are not spatially associated

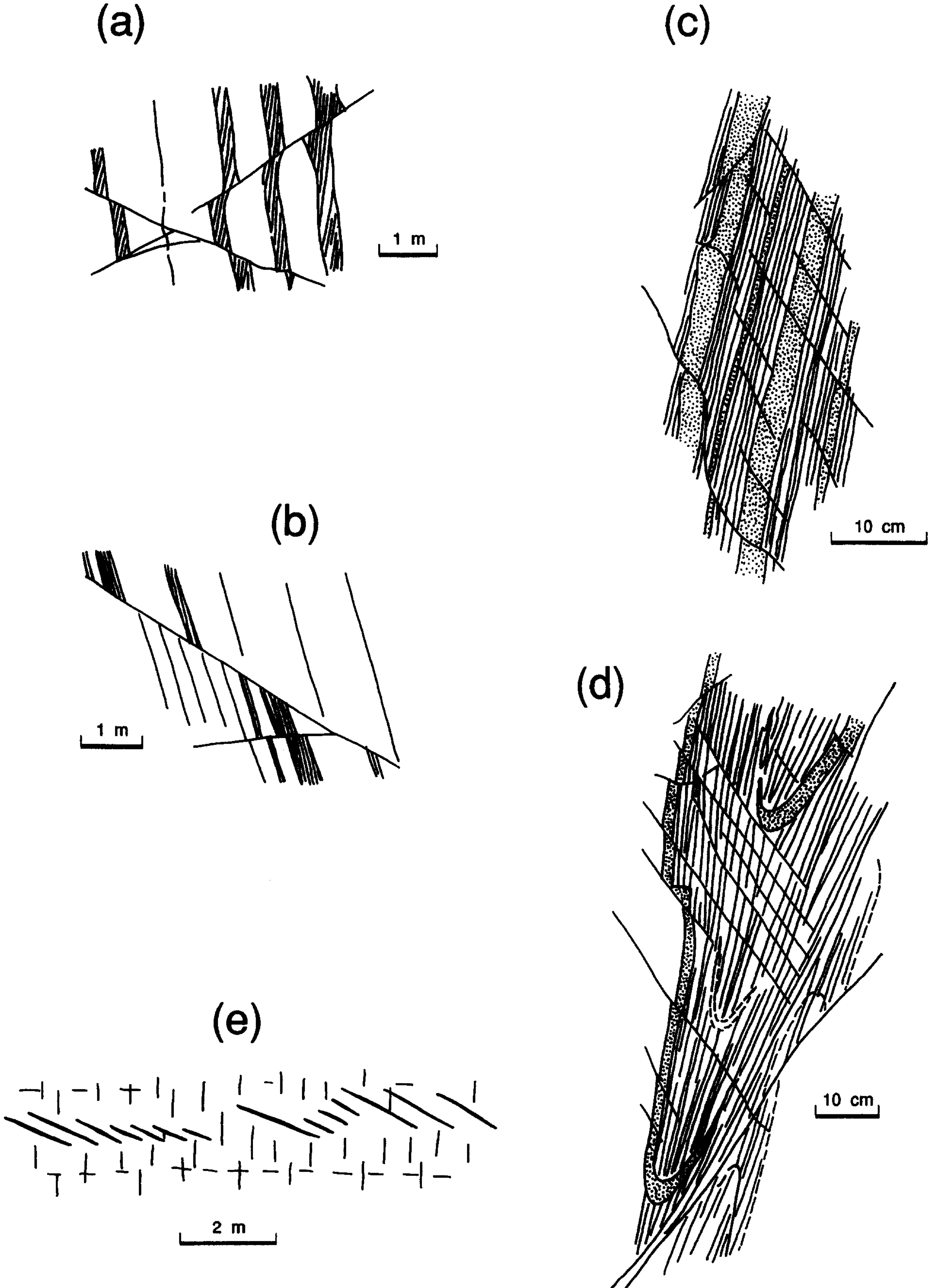


Fig. 9. Sketches of small-scale NE-striking dip-slip faults from the Meguma Terrane and Newfoundland. (a) & (b) Reverse faults in turbidites in road metal quarries along Highway 102 north of Halifax Airport. Both are viewed, on approximately vertical surfaces, looking northeast. (c) & (d) Reverse faults in transposed turbidites in the Chanceport shear zone, Bay of Exploits, Newfoundland. Both are viewed on approximately vertical surfaces, looking west. Note minor normal displacement on some of the faults suggesting reactivation; they may all have been reactivated but the net slip is normal on only a few. (e) Array of shallowly-dipping fractures, in massive sandstone in the Meguma Terrane, Eastern Shore, Nova Scotia, viewed on an approximately vertical surface, looking northeast. Bedding is poorly developed and is horizontal. there is a well developed, approximately vertical cleavage. The fracture array extends to either end of the outcrop and the few individual fractures for which there are markers show normal movement.

with older shear zones. They are the youngest structures recorded there (except for normal faults) and post-date Taconic thrusting, but their age is otherwise unconstrained. Slickenside striae indicate transcurrent as well as dip-slip movement on these faults (Bosworth 1985).

Along and adjacent to the northernmost segment of the Dunnage-Gander Zone boundary there are many NE-striking dip-slip faults (Figs. 4f & h). Both reverse and normal motions have been recorded, but mostly the sense of movement is unknown. We interpret these faults as positive flower structures associated with Group 1 transcurrent faults along the terrane boundary. This interpretation is supported by the observation that many of the faults have a large component of transcurrent movement associated with them (transcurrent and dip-slip components can be equal). Reverse movements are consistent with the interpretation and normal movements can be explained by later reactivation as in the Carboniferous structures discussed below. In the Baie Verte area adjacent to the Dunnage-Humber Zone boundary dip-slip faults of the same orientation (Figs. 4e & g) are recognized even though other orientations are dominant there. These faults show all the same features as those along the Dunnage-Humber Zone boundary discussed above and interpreted in the same way. Shallowly dipping transcurrent faults (Fig. 4c) in the Baie Verte area may also be related to the same transcurrent motions.

In Nova Scotia dip-slip faults known to us trend NE and, as in Humber Arm, are not localized in older zones (Figs. 9a & b). In New Brunswick the faults are larger but have similar orientations. In the vicinity of Quaco Head there are two E-dipping thrusts in Carboniferous rocks (Plint & van de Poll 1984). One, the Quaco Head Fault, was active as a thrust in late Visean and Westphalian times; the other, the Rogers Head fault, was also active as a thrust in Westphalian times. Both were reactivated as normal faults in Jurassic times (Plint & van de Poll 1984). A similar thrust fault located on Barnaby Head emplaces older granitic mylonites onto Carboniferous sedimentary rocks (Variscan Front of Rast and coworkers, e.g. Rast & Grant 1973, Rast & Dickson 1982). The fault is brittle, but formed parallel to a pre-existing mylonitic foliation. Other westerly directed thrusts are known in the Carboniferous of New Brunswick (e.g. Rast & Grant 1973, Ruitenberg & McCutcheon 1982, Nance 1987) and it is possible that they generally represent positive flower structures associated with transcurrent movement on an extension of the Cobequid Fault running parallel to the Bay of Fundy. In this context it is significant that Plint & van de Poll's (1984) Westphalian age for the thrusting on Quaco and Rogers Heads is coeval with the transcurrent movement on the Cobequid Fault (Yeo & Gao 1987).

Normal faults

Dip-slip faults, mostly steep, generally appear to be late relative to all other structures in the same outcrops, but most cannot be dated. It is possible that there is more than one period of deformation represented. However, since many of the faults cut Late Carboniferous rocks and since similar faults are known to cut Mesozoic rocks, we interpret them as being related to Atlantic opening.

Steep dip-slip faults striking parallel to the general trend of the orogen are not uncommon in Carboniferous and older rocks along the south coast of New Brunswick. Many of them can be shown to be normal faults. The Carboniferous thrust faults on Quaco and Rogers Heads are mostly shallowly dipping but they were reactivated as normal faults in Triassic or Jurassic times (Plint & van de Poll 1984).

In Nova Scotia many small (decimetre to metre scale), essentially planar fractures occur in arrays, commonly with a decimetre-scale spacing. Individual fractures dip SE or NW and trend approximately NE. It is generally impossible to determine either the sense of movement on individual fractures or the overall configuration of the arrays. Locally, however, arrays of fractures with normal displacements define a horizontal or shallowly dipping zone (Fig. 9e). These are apparently Riedel fractures associated with low-angle, incipient detachment faults. The consistency in the orientation of the fractures indicates that they are younger than the folds in the area which are generally believed to be Silurian or early Devonian in age.

In Newfoundland, steeply-dipping normal faults have been recorded in the Humber Arm allochthon and, as stated above, some are believed to be reactivated 'thrust' faults (Bosworth 1985). Similarly, some individual faults in the arrays of reverse faults in the Bay of Exploits show normal displacement (Figs. 9c & d), suggesting local reactivation. Steep and shallow dip-slip faults (Figs. 4e–h) have already been described from the northern ends of the Baie Verte Line and the Dunnage/Gander Zone boundary region and, as stated above, the sense of movement on these faults is generally unknown; the few faults for which the sense can be determined are roughly equally divided between reverse and normal faults, with one notable exception; where movement sense can be determined, steep W-dipping, N-trending dip-slip faults in the Gander area (included in Fig. 4f) are dominantly normal faults (Goodwin & O'Neil 1991). One of these faults is slightly offset by a W-trending, steeply-dipping dextral transcurrent fault; the normal fault may either be older than or approximately coeval with the dextral fault. In most cases, however, the relative ages of the faults cannot be determined.

Normal faults trending predominantly in the northeast quadrant and dipping both SE and NW are common in Mesozoic sediments on the Canadian continental shelf (e.g. Tankard & Welsink 1989, Tankard *et al.* 1989, Welsink *et al.* 1989, Keen *et al.* 1991b). On land, Triassic rocks occur in half-graben throughout the Appalachians (e.g. Nadon & Middleton 1985, Costain & Çoruh 1989). These Mesozoic structures are related to Atlantic opening and represent crustal scale extension; they must therefore have a counterpart in the basement rocks. We suggest that the counterpart is to be found in the steep and shallow dip-slip faults described here.

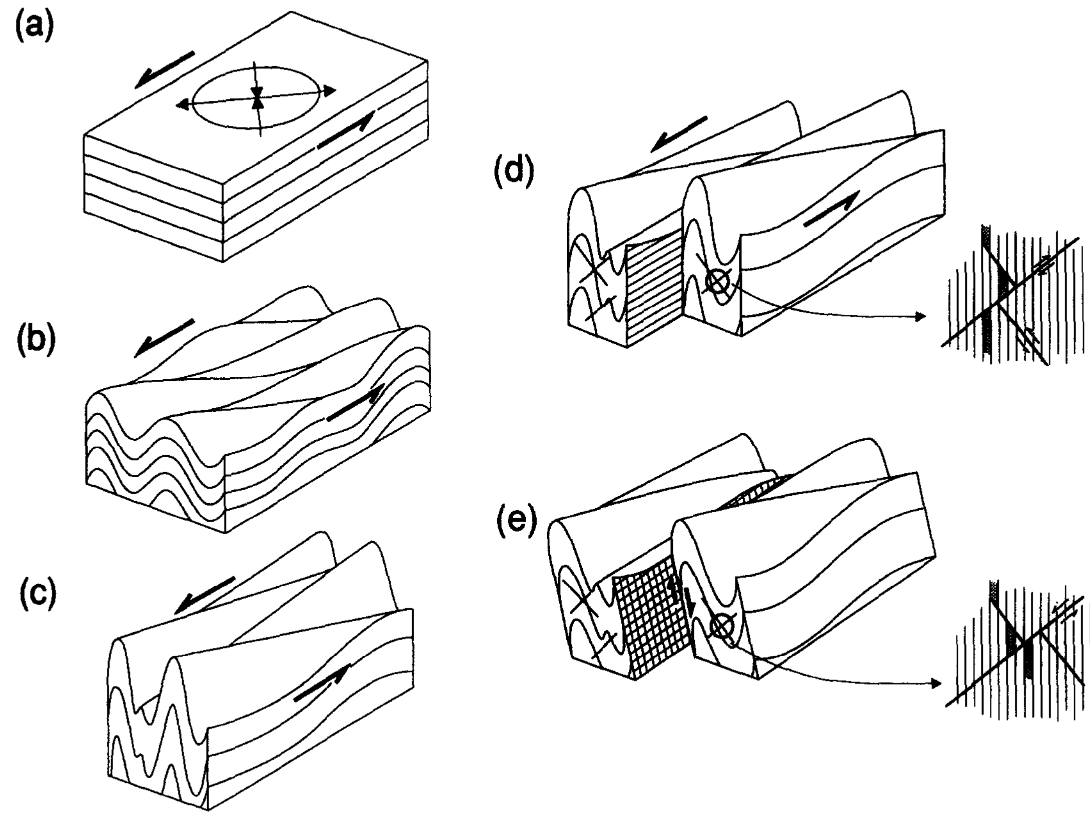


Fig. 10. Possible sequence of structural events in the Meguma terrane. (a)–(d) represents Palaeozoic transpression and (e) represents Mesozoic extension. Insets in (d) and (e) show development of Carboniferous reverse faults and their reactivation as normal faults in the Mesozoic. See text for discussion.

DISCUSSION

History and geometry of the faults

From published work and the work reported here, we recognize a pattern in the brittle faulting history which is consistent with all the observations. Transcurrent faults striking in the northeast quadrant and reverse faults are coeval and were active from Devonian at least until Late Carboniferous times. They apparently represent a continuation of earlier ductile transpressive conditions that persisted at least until the end of the Silurian (e.g. Elliott *et al.* 1991, Lafrance & Williams 1992) and locally as late as Carboniferous (Yeo & Gao 1987). Transcurrent faults striking in the northwest quadrant and normal faults also appear to be coeval and are mostly of Mesozoic age. They form a transfer-extensional fault pair (Gibbs 1984, Lister *et al.* 1986) and are related to the opening of the Atlantic.

Palaeozoic

In the context of Devono-Carboniferous transpression, boundinaged bedding-parallel veins and pressure shadows lying in the axial plane foliation indicate extension parallel to the NE-trending Silurian–Early Devonian fold hinges in the Meguma Terrane (unpublished work by P. F. Williams). Thus both the folding and the faulting that followed it are compatible with a transpressional event during the Devonian and Carboniferous in the Meguma Terrane and probably throughout the region. We suggest that (1) these folds formed at an angle to the orogen in a shear environment that progressively rotated them towards parallelism with the orogen, and (2) transcurrent faults developed as bedding was rotated by folding and shear into a suitable orientation for layer-parallel slip. In such an environment, reverse faults would be expected to become progressively more common as tightening of folds and rotation of bedding into the shear plane orientation rendered folding and oblique shear on steep planes less viable as a shortening mechanism. This interpretation is summarized in Fig. 10; the final product is steeply-dipping bedding, steeply-dipping transcurrent faults and a conjugate array of reverse faults.

We suggest that this might be a general model for the history of a collision. Initially when bedding is horizontal, thrusting and folding with or without transcurrent motion (depending on whether the collision is orthogonal or oblique) are the preferred deformation mechanisms. As bedding and early foliations are transposed into vertical attitudes by folding new mechanisms have to develop and the conjugate reverse faults might be expected to form right across the orogen.

Mesozoic

The extensional faults that we associate with the opening of the Atlantic are mostly steep or are reactivated reverse faults (less steep). We interpret the steep faults as domino-type extensional features (cf.

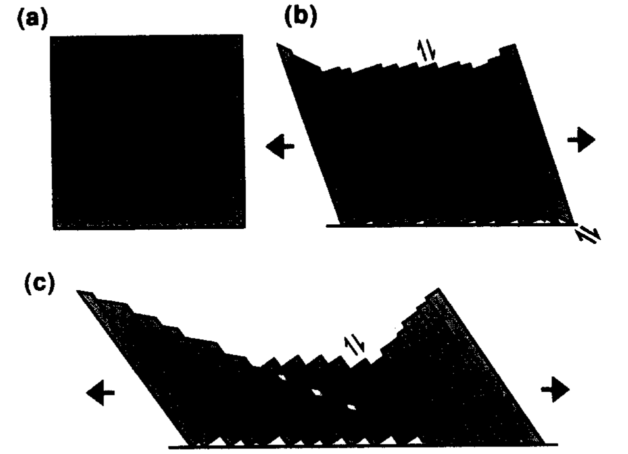


Fig. 11. A model for extension of crust that has shallowly dipping faults and a vertical planar anisotropy. It is assumed that both the faults and the vertical anisotropy are planes of weakness and therefore available slip-systems to accommodate the deformation; (a)–(c) represents progressive horizontal extension.

Williams & Hy 1990). This interpretation is essentially the same as that proposed by Tankard & Welsink (1987) for the continental shelf around the Hibernia oilfield on the basis of seismic data. In their interpretation, the steep faults sole in a listric normal fault. Similar detachments might be expected throughout the region, but from our surface observations we do not know of any large brittle faults that are likely candidates. However, Roberts & Williams (1993) have shown that the shallowly-dipping ore body in the Denison Potash Mine, New Brunswick is a mylonite zone on which the hanging-wall moved to the southeast. Evaporites are of course capable of ductile flow under conditions that would result in brittle behaviour in most other rocks. Further, because of their weakness and subhorizontal attitude in the Denison Mine area they would channel any detachment faulting. One possible interpretation of the structure, therefore, is that the evaporites acted as a ductile detachment fault during Atlantic opening (Roberts & Williams 1993).

The lack of subhorizontal detachment faults may be a sampling problem or it may be that such faults are rare. If rare, a possible reason is that there were already adequate deformation mechanisms without the development of new faults; a combination of a steep planar anisotropy (layering, foliation and transcurrent faults and shear-zones) and moderately dipping reverse faults comprise adequate planes of weakness for plane-strain extension (Fig. 11).

Many of the transfer faults have small horizontal separations, and it is likely that vertical displacements are even smaller. This may explain why their effect is generally ignored. However, the faults are sufficiently numerous that their cumulative effect may be large even where there are no individual faults with large displacements.

Many transfer faults may have been ignored because there are markers that cross them, suggesting that there is no significant displacement associated with them. However, it is important to realize that the displacement can vary rapidly along transfer faults because of their interaction with coeval normal faults (Fig. 12). A consequence of this is that a major fault may show only minor displacement along much or even all of its length.

For example, the major lineament approximately coincident with lines 3–9 of the Lithoprobe East seismic transect (Fig. 3) is *ca.* 200 km long, is clearly visible on aerial photographs, crosses almost the full width of the Humber and Dunnage Zones, is visible as a strong lineament on aeromagnetic maps (H. G. Miller personal communication 1990) and is visible as a break in various lithological units. Nevertheless it has no obvious regional-scale displacement. However, it may be responsible for the straight line termination of the Long Range Inlier in the Humber Zone (as discussed above) and the southern contact of the Topsails batholith in the Dunnage Zone (Fig. 3) (Colman-Sadd *et al.* 1990). There may be many locally significant displacements which vary in both sense and magnitude and therefore do not integrate into a significant regional displacement as shown diagrammatically in Fig. 13.

The characteristic variation in magnitude and sense of displacement along transfer faults is potentially important in explaining other observed features in the region. A prominent example in the Baie Verte peninsula is the broadening of two map units (Birchy Complex and Rattling Brook Group) northeastward across the NW-trending Little Lobster Harbour Fault (see Hibbard 1982). A possible interpretation is shown in Figs. 12(c) & (d). This fault coincides with strong aeromagnetic (H. G. Miller personal communications 1990) and photo-lineaments which extend right across the Baie Verte peninsula. The lineament also coincides with the approximately 15 km long NW-trending straight contact of the Cape Brule porphyry.

Two general conclusions are supported by our observations. First, the apparent displacement along *individual* faults may not accurately reflect the importance of the fault *systems* to which they belong. In particular, apparent displacement of markers varies along transfer faults and may be quite small. The entire transfer-normal fault system however, may accommodate significant deformation and may have a profound influence on the present orientation of earlier structures. Second, it is well known that anisotropy influences deformation. We have presented a model for accommodation of crustal extension by a mechanism other than that involving the classic pair of transfer faults and listric normal faults. In the model presented (Fig. 11) listric normal faults are replaced by moderately dipping faults (which exploit pre-existing reverse faults) operating in concert with steeply dipping slip surfaces (in this case, pre-existing foliation surfaces and/or faults).

Correlation of structural observations and Lithoprobe East seismic data

The most prominent features in the upper levels of the Lithoprobe East seismic reflection sections are discontinuities that pitch in the plane of the section at approxi-

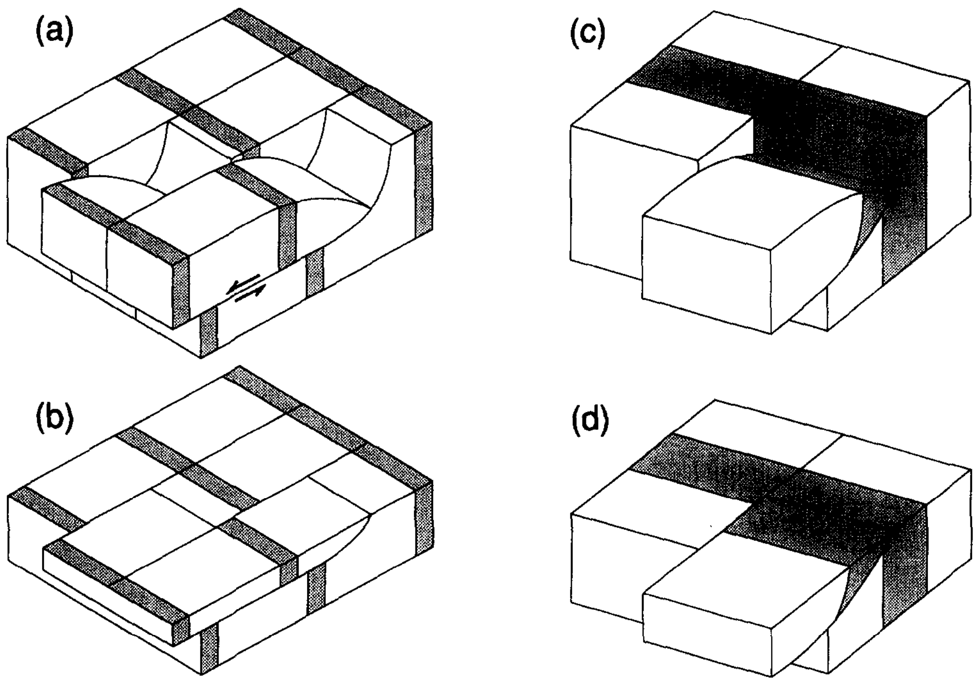


Fig. 12. (a) & (b) Diagram showing how displacement can vary rapidly along a transfer fault where the motion has been transferred from one listric fault to an adjacent one. (b) shows the outcrop appearance after erosion. (c) & (d) diagram showing how the outcrop width of a given unit may increase significantly across a transfer fault. (d) shows outcrop appearance after erosion. See text for further discussion.

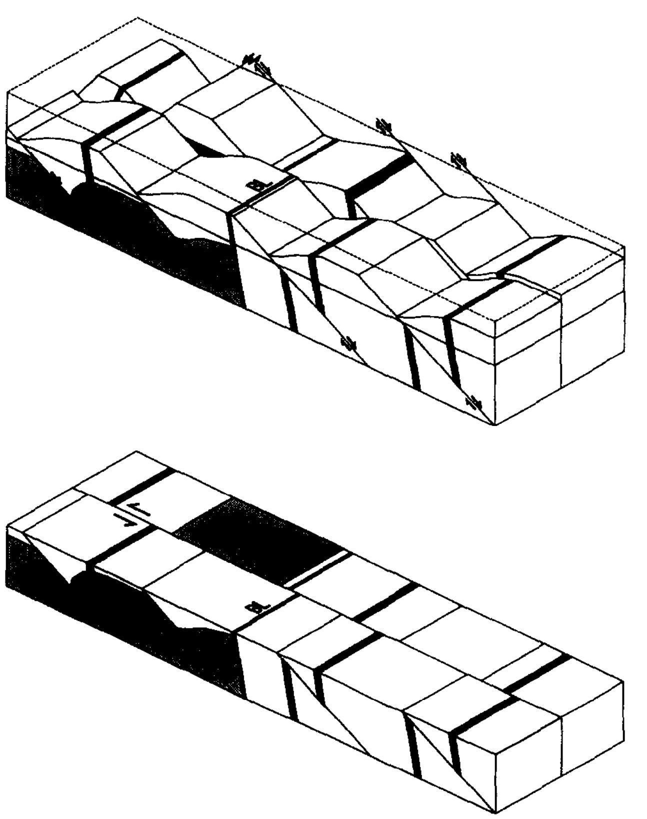


Fig. 13. Array of transfer and normal faults showing irregular variation (compare markers) in displacement along the length of the transfer fault. The lower block shows the outcrop appearance after erosion. Extension of the crust in this diagram is assumed to be achieved according to the model illustrated in Fig. 11. Note that although the principal movement along the transfer fault is transcurrent, movement vectors will vary considerably in orientation within the fault plane. See text for further discussion. 'BL' represents a major vertical boundary such as the Baie Verte Line as interpreted here.

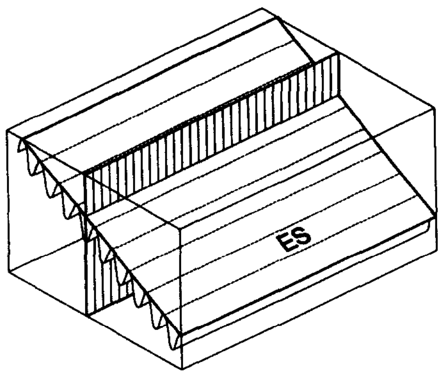
mately 30°. Mostly these are apparent dips. However, one of the structures is interpreted (Hall *et al.* 1990) as striking approximately NE (perpendicular to the transect line) so that the observed pitch represents a true dip, suggesting that the others may also represent true dips also, a least to a first approximation. One of the surprises of the Lithoprobe data has been that, with the exception of the Dover Fault, none of the steeply-dipping shear zones or faults associated with major tectonostratigraphic boundaries such as the Baie Verte Line show up on the sections. Instead, the shallowly dipping features may cross the downward projection of such boundaries.

Surface evidence indicates that both the lineament-forming major ductile faults and most major lithological boundaries are generally steep. For example at Mt Cormack, Newfoundland the thrust associated with a tectonic window dips steeply (in excess of 45°) away from the window on either side (Colman-Sadd & Swinden 1984). In other words, the structure that gives rise to the window is very tight and it is notable that it is overprinted by a reverse fault with a dip of the order of 30° (Colman-Sadd & Swinden 1984), consistent with our interpretation of the overall structural history of the region.

To reconcile these observations with the seismic reflection data we have to assume that (1) there is a marked, ubiquitous change in structure with depth, (2) the structure seen at the surface is simply noise on a much larger structure which is imaged by the seismic reflection data, or (3) the structures observed in the sections are young enough to overprint the steep structures observed at the surface.

The first of these options is unlikely. It would require

(a)



(b)

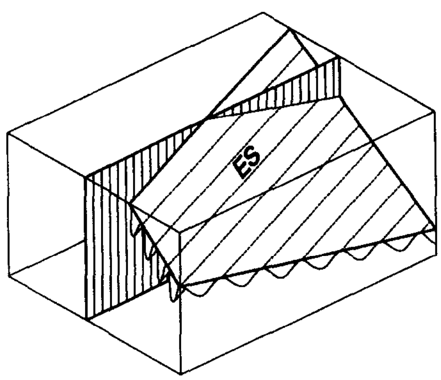


Fig. 14. ES is an enveloping surface to a given folded horizon. The traces of the fold hinges are shown on the enveloping surfaces as broken lines. The vertically hatched surface represents a vertical fault. See text for discussion.

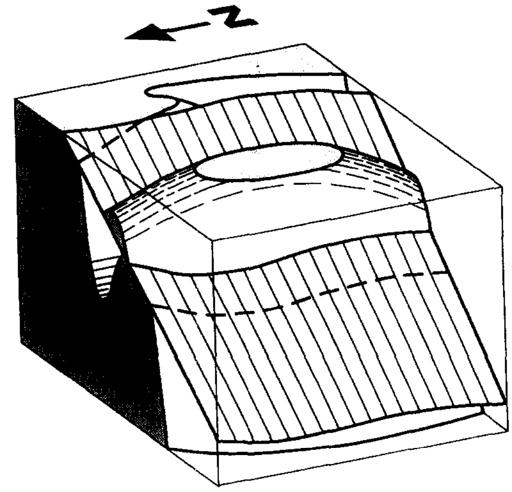


Fig. 15. Cartoon illustrating a possible relationship between the Dunnage/Gander Zone boundary and the shallowly-dipping reflector that coincides with the northern end of the boundary. The stippled areas represent the Gander Terrane and the surface distribution is simplified from that shown in Fig. 1 (the two windows of Gander in the Dunnage Zone (Fig. 1) are represented by a single window here). See text for further discussion.

that the change occurs between the surface and a travel-time depth of 2 s throughout the area covered by the on-land survey. Such an interpretation has no geological merit.

The second possibility at first sight seems geologically more reasonable than the first. The shallow reflectors could be enveloping surfaces to major lithological packages. The folds, shear zones and/or faults observed on the surface could then be structures that transposed all mesoscopic surfaces into a steep attitude while preserving the gross distribution of the enveloping surface. Such a structure is commonly seen on a mesoscopic scale (Hobbs *et al.* 1976, p. 253).

This interpretation is geometrically sound but is geologically unsatisfactory. It requires the major boundaries represented by the shallow reflectors to be older than all the mesoscopically observed structures; the boundaries would need to be related to the initial terrane accretion. Then, since the shallow reflectors are seen on the seismic profile (Fig. 2) to cross the vertical downward projection of features such as the Baie Verte Line, it would mean that such features as the Baie Verte Line caused no significant vertical separation of the earlier markers responsible for the shallow reflections (see Fig. 14a). In geometrical terms this is a reasonable explanation, but since the surface expression of such combined shear zones and faults is persistent for considerable distances (>350 km for the Baie Verte Line) and is parallel to the overall strike, the intersection of the fault surface and the shallowly dipping reflector would have to be horizontal (Fig. 14a), otherwise the trace of the fault would not be parallel to the overall strike. Certainly we would see a sinuous outcrop convergent with the major fault lineaments for some of the enveloping surfaces, if they existed (Fig. 14b).

It could be argued that the Dunnage-Gander Zone boundary is such an enveloping surface, as it is sinuous and irregular in outcrop over much of its length. In the northeast the boundary coincides reasonably well with the interpreted position of the westerly dipping reflector (Hall *et al.* 1990) mentioned above. Toward the south however, where the boundary between Dunnage and Gander Zone rocks becomes much more complex and sinuous, the reflector does not follow it. For example, Gander Zone rocks outcrop in the vicinity of Meelpaeg Reservoir (see Colman-Sadd *et al.* 1990), but there is no shallowly-dipping reflector coincident with their boundary. Figure 15 shows these relationships in a very simplified cartoon. For the purpose of representation, the boundary between the Dunnage and Gander zones is shown as a simplified enveloping surface. The plunges have been assumed so as to place the Dunnage over Gander rocks (see Colman-Sadd & Swinden 1984). We suggest that the boundary is probably segmented in detail and stepped across a series of crustal slices bounded by anastomosing, generally NE-trending, steeply dipping, ductile and brittle faults, reflecting the structure as it was in Late Devonian times. Such an interpretation of the Dunnage Zone as a completely shredded zone is consistent with our observations in Notre Dame Bay (Williams *et al.* 1988b) and with the conclusions of Williams (1989) based on fossil distribution in the central Dunnage Zone. The interpretation presented in Fig. 15 requires that the Dunnage-Gander Zone boundary be older than the shallow reflector and suggests that the latter is a late fault rather than an enveloping surface to older structures.

The third possibility is consistent with Fig. 15 and is geologically the most reasonable. For example, if the reflectors seen in the seismic sections are large examples of the Late Carboniferous thrusts or reverse faults they would cut the ductile faults and all contacts steepened by earlier folding. They could even cut some or all of the Carboniferous transcurrent faults since we have no evidence for or against the reverse faulting having persisted longer than the transcurrent movement. If the convergence direction associated with the transpressive regime became more nearly orthogonal to the orogen through time, then the reverse faulting would be expected to outlast the transcurrent movement. Interpreting the reflectors as reverse faults has geological merit. Their geometry, even to their occurrence as what could be interpreted as conjugate pairs (Fig. 2, see also fig. 3 Line 11 in Hall *et al.* 1990), is identical to what we see on a small scale, and the age of the faults is such that we might expect even large-scale structures to be preserved, especially if they were reactivated as normal faults during Atlantic opening as observed on a smaller scale. Alternatively, they could all be normal faults associated with Atlantic opening as suggested recently for similar reflectors identified in transects in the British Isles (Blundell 1990). However, the shallow dips fit better with reverse rather than normal faulting; since we see reverse faults of similar orientation in the region and since the normal faults that we do see are commonly steeper, we favour the reverse fault origin with probable reactivation during normal faulting.

Keen *et al.* (1991a) have interpreted such faults in the Bay of Fundy and Gulf of Maine as thrusts. They present two possible interpretations. According to one (their fig. 10a) the Avalon Terrane is first thrust southeastwards over the Meguma Terrane and then internal northwesterly directed thrusts develop within the Avalon Terrane as it is thrust in the same direction over the 'Central Block'. The movement is interpreted as Carboniferous in age but it is suggested that the northwesterly directed thrusts may reactivate earlier structures. According to the interpretation, movement on each of the two thrusts at the base of the Avalon Terrane must have exceeded 50 km and the Avalon Terrane itself must have been quite narrow (*ca.* 100 km). We know of no evidence that Carboniferous thrusting was associated with such large scale movement in this part of the Appalachians. It can be argued that it is necessitated by transcurrent movement on the Cobequid-Chedabucto Fault but that argument is predicated on the assumption that the Meguma Terrane behaved as a rigid, non-rotating block. If that constraint is relaxed, there is no reason for large-transport thrusts to accommodate the transcurrent movement.

In Keen *et al.*'s (1991a) second interpretation (their fig. 10b) Meguma Terrane basement is thrust northwesterly, on the Fundy Fault, over Avalon Terrane; the latter is still separated from the 'Central Block' by a parallel thrust. The southeasterly directed thrust (their shear zone B) is interpreted as a late structure with minor displacement which would fit well with the Carboniferous thrusts as we know them. The northern limit of the Meguma Terrane in this interpretation is speculative and it can be argued that the boundary is an undetected steeply-dipping feature that is cut by shear zone B. This would be consistent with our interpretation.

The reflectors were previously interpreted as thrusts (Stockmal *et al.* 1987) that were thought to be early structures associated with Siluro-Devonian movement. Our objection to that interpretation is that for early thrusts to persist as planar markers on the scale of the reflectors would be inconsistent with surface observations that show contacts generally steep and thrusts repeated by folding and faulting (e.g. Colman-Sadd & Swinden 1984, van Staal & Williams 1988).

As a general point we suggest that where reflectors are a product of deformation it is likely that they will represent late structures since early structures are inevitably modified by later tectonic events. Such modification is likely to render them poor as reflectors unless it simply involves reactivation.

Interpreting the shallow reflectors as late faults provides a possible explanation for the lack of seismic evidence for most of the vertical faults. As noted previously the Dover Fault is the only vertical fault recognized in the seismic sections. It is not imaged directly, but is recognizable as a zone with no through-going structures and, primarily, as a break in a strong 'E-reflector' coincident with the Moho (Keen *et al.* 1986) under the Gander Zone. It is neither cut by any of the reflectors that we interpret as reverse faults nor does it cut any. If one of the thrust reflectors did cut the Dover Fault, the latter would be much more difficult to recognize. It may be that other vertical faults persist to considerable depth, but that the presence of late shallow structures obscures them. In this interpretation, the difference between the Dover Fault and the Baie Verte Line, for example, is that either (a) fortuitously no late faults intersect the Dover Fault or (b) late movement on the Dover Fault is younger than the reverse faults. Either way we suggest that the vertical faults seen on the surface may persist to depth but are only recognizable where they are not overprinted by later reverse and normal faults and where good displaced shallow markers exist.

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