Current Research (2003) Newfoundland Department of Mines and Energy
Geological Survey, Report 03-1, pages 1-13

STRUCTURAL RELATIONSHIPS SOUTH OF GRAND LAKE, NEWFOUNDLAND

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ABSTRACT

The Cabot Fault Zone and the Little Grand Lake Fault in western Newfoundland experienced a complex structural history. A schematic deformational history for both structures is proposed, based on field studies combined with preliminary isotopic age-dates (U–Pb and $^{40}Ar–^{39}Ar$). During the Early to Middle Ordovician, the Little Grand Lake Fault accommodated north-directed thrusting of the Dashwoods Subzone over the Notre Dame Subzone. Thrusting of the Dashwoods Subzone on top of the Humber Zone started during the Middle Ordovician. This phase of the deformation in the Cabot Fault Zone truncates the Little Grand Lake Fault and therefore postdates the initial movement accommodated by the Little Grand Lake Fault. The Middle Ordovician thrusting phase initially emplaced the Disappointment Hill complex (DHC) westward onto the internal Humber Zone, based on our interpretation that the DHC has a Dunnage Zone provenance. During the Late Ordovician, Dashwoods Subzone rocks moved down relative to the Humber Zone along the Cabot Fault Zone, while thrusting continued in the internal Humber Zone. West-directed thrusting in the Humber Zone lasted at least until the Late Silurian and created a metasedimentary thrust stack in the internal Humber Zone. During the Late Silurian–Devonian (?), oblique dextral strike-slip movements in the Cabot Fault Zone, truncated the earlier formed thrust stack, and moved the Dunnage Zone down relative to the Humber Zone.

Agmatites, tonalites and gabbros of the DHC show an affinity with the Ordovician igneous complexes that occur in the Dashwoods and Notre Dame subzones, such as the Southwest Brook complex. If they are indeed related, this would imply that the DHC is an allochtonous outlier of the Dunnage Zone in the Humber Zone, comparable with the Coney Head complex in northern Newfoundland.

INTRODUCTION

The Humber Zone has been interpreted to represent the preserved remnants of the Laurentian continental margin (e.g., Williams, 1979; Cawood *et al*., 1994; Waldron and van Staal, 2001). The internal Humber Zone is the eastern, more intensely deformed and metamorphosed part of the Humber Zone. South of Grand Lake, this zone consists of metasedimentary rocks that are structurally interleaved with (meta)plutonic rocks of the Disappointment Hill complex (DHC), anorthosite and related rocks of the Steel Mountain complex and Meso- to Neoproterozoic(?) granitoid gneissic rocks (Currie and van Berkel, 1992). The metaplutonic rocks have classically been interpreted to represent Proterozoic basement to the metasedimentary rocks of the internal Humber Zone. Some of these units have been intruded by schistose amphibolitic dykes, plausibly related to the late Neoproterozoic Long Range Dykes (Owen and Greenough, 1994; Wendland, 2002). The metasedimentary rocks in the eastern part of the Humber Zone comprise metapsammites, metapelites, calcareous schists and marble, probably correlatives of the Mount Musgrave Group and Fleur de Lys Supergroup, and are exposed in a series of thrust sheets (*see also* Cawood and van Gool, 1998). A leucocratic peralkaline granite (Hare Hill granite, 608 ± 4 Ma, U–Pb on zircon, Currie *et al*., 1992) is situated in the centre of the internal Humber Zone, however, its relationship with the DHC and other units of the internal Humber Zone are uncertain because most of the contacts are faulted (*see also* Owen and Currie, 1991).

The western part of the Dunnage Zone has been subdivided into the Notre Dame and Dashwoods subzones (Figure 1), separated by the Little Grand Lake Fault. The Notre

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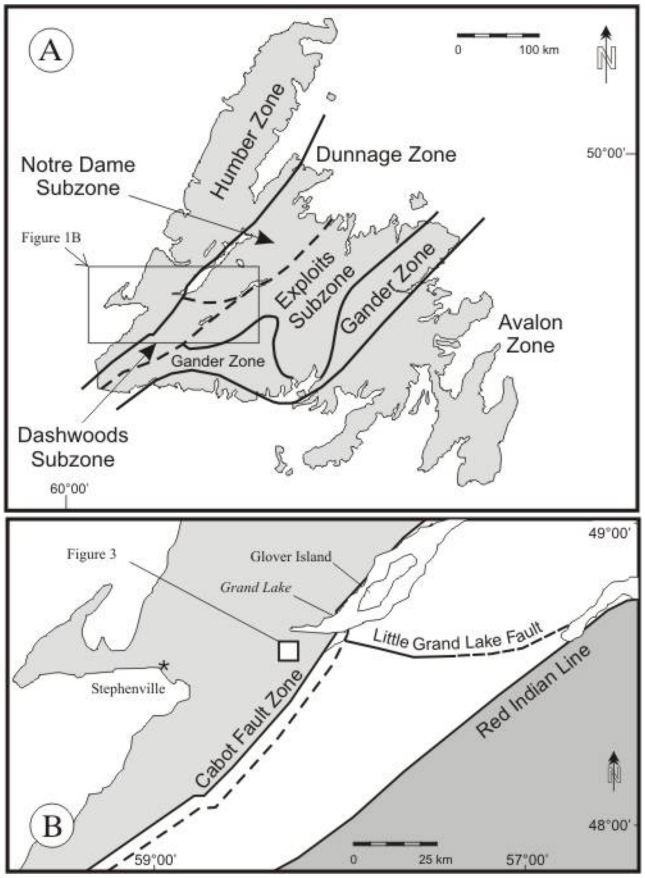

Figure 1. *(A) Map of Newfoundland showing the main tectono-stratigraphic subdivisions; (B) map showing the main structures in southwestern Newfoundland and the location of Figure 3.*

Dame Subzone mainly contains low-grade (greenschist) metamorphosed rocks of the ophiolitic Grand Lake complex (490 ± 4 Ma, U–Pb on zircon, Cawood *et al*., 1996) and overlying Glover Group (Knapp, 1982; Szybinski *et al*., 1995) and the weakly to undeformed and metamorphosed Topsails Igneous suite, consisting of the Glover Island granodiorite (440 ± 2, U–Pb on zircon, Cawood *et al*., 1996), the Rainy Lake complex (438 ± 8 Ma, U–Pb on zircon, Whalen *et al.*, 1987), the Topsails granite (429 ± 3 Ma, U–Pb on zircon, Whalen *et al*., 1987) and associated rock units. In contrast, the Dashwoods Subzone consists mainly of deformed medium- to high-grade metamorphic rocks, including migmatized metasedimentary rocks, mafic–ultramafic ophiolitic remnants, agmatites made up of mafic schollen surrounded in a leucocrate matrix of tonalite, variably foliated Ordovician granodiorite plutons, para- and orthogneissic rocks of the Cormacks Lake complex and weakly or undeformed Silurian mafic to felsic plutons (e.g., the Main Gut gabbro; Herd and Dunning, 1979; Dunning *et al*., 1989; Currie and van Berkel, 1992; Whalen, 1993).

There exists a contrast in metamorphic grades and absolute ages between the Humber Zone and adjacent subzones of the Dunnage Zone in western Newfoundland. Cawood *et al*. (1994) argued that rocks in the internal Humber Zone are mainly affected by Early Silurian lower amphibolite-facies metamorphism and deformation. In contrast, the main deformational and high-grade (up to granulite-facies) metamorphic phase in the Dashwoods Subzone is generally believed to be pre-Silurian (Taconic; e.g., Dunning *et al.*, 1989; Currie and van Berkel, 1992). Pre-Silurian rocks of the Notre Dame Subzone have been metamorphosed under greenschist-facies conditions, but the timing of deformation and metamorphism is not well documented. Nevertheless, the absence or weak nature of a deformational and metamorphic overprint in the Early Silurian Topsails Igneous suite, (440 to 427 Ma), combined with the presence of a pre-Silurian unconformity does suggest that the main tectono-metamorphic event in the Notre Dame Subzone is pre-Silurian (*see also* Whalen *et al*., 1987).

During the 2002 field season, work has continued on the nature of the Cabot Fault Zone and the Little Grand Lake Fault (Figure 1) in western Newfoundland to better understand the regional geology of the area. The (brittle-) ductile deformation in the Cabot Fault Zone has been bracketed between Early Ordovician and latest Silurian (Dunning *et al.*, 1990; Cawood *et al.*, 1994). Post-Salinian movement did occur in the Cabot Fault Zone (page 527 - 537 *in* Williams, 1995; Cawood *et al*., 1996), but this dominantly brittle deformation phase is not a main focus of this study. Instead, this study is focused on the brittle–ductile to ductile elements formed in response to deformation on the (sub)zone bounding structures. The kinematic history has to be resolved in order to present a tectonic model for the interaction between the Humber and Dunnage zones in western Newfoundland.

STRUCTURES

THE CABOT FAULT ZONE

In the literature, several different terms are used to describe the Cabot Fault Zone (Figure 1b). Wilson (1962) named the structure, the Cabot Fault Zone, and defined it as a northeast-trending, narrow belt of large faults in western Newfoundland that continues southward all the way into New England. It is a fault zone with a long and complex movement history. However, the term Cabot Fault Zone is restricted (by some authors) to structures related to brittle Carboniferous movement(s) and associated opening of the Deer Lake and Bay St. George ‘strike slip’ basins (page 527-537 *in* Williams, 1995; Cawood *et al*., 1996). In southwestern Newfoundland, the term Long Range Fault has been introduced to describe the boundary between the Humber Zone and the Dashwoods Subzone (Piasecki, 1988; Brem *et al.*, 2002) and in some reports the term Long Range (Cabot) Fault is used (e.g., Piasecki, 1988; Whalen *et al.*, 1993). Generally, both the term Long Range Fault and Cabot Fault are interchangeable (e.g., throughout Williams, 1995). A third term used to describe the structure is the Baie Verte (-Brompton) Line, which is defined as, "a narrow structural zone marked by discontinuous mafic–ultramafic plutons" (Williams and St. Julien, 1982). For southwestern Newfoundland, the Baie Verte–Brompton Line is interpreted to coincide with the Cabot Fault Zone / Long Range Fault (e.g., Figure 1 *in* Fryer *et al*., 1997). Herein, the term Cabot Fault Zone is used as defined by Wilson (1962) to describe the structure that separates the Humber Zone from the Dunnage Zone (Figure 1).

In the study area, the Cabot Fault Zone is up to a few kilometres wide and is characterized from west to east by a greenschist-facies mylonite zone, a cataclastic deformation zone and an amphibolite-facies high-strain tectonite zone (Brem *et al.*, 2002).

Throughout the study area, muscovite–chlorite greenschist-facies mylonites are developed in the western part of the Cabot Fault Zone (*see also* Brem *et al*., 2002). Foliations strike north-northeast and dip subvertically to the east or west (Figure 2a). Mineral lineations and stretching lineations (e.g., elongate quartz-aggregates) plunge about 30° south (Figure 2b). Locally, some more steeply plunging lineations or northeast-plunging lineations occur, but these deviations do not appear to be systematic. Shear-sense indicators such as winged objects (Plate 1) and shear bands indicate an oblique (dextral and west-side up) movement.

The cataclasite zone (described in Brem *et al.*, 2002) is best developed in a sliver of Glover Group volcanic rocks, which have been deformed in the Cabot Fault Zone, south of Grand Lake (*see also* Currie and van Berkel, 1992). The cataclasite is also developed east of the Steel Mountain anorthosite, but to a minor extent. Generally, the cataclasite always occurs east of, and in close proximity to, the mylonites (Brem *et al.*, *op. cit.*). Centimetre to decimetre-scale striated surfaces and slickensides are abundant within this zone, but all structures are oriented in such a way that stress tensor analysis is not possible.

The amphibolite-facies tectonite zone in the Cabot Fault Zone is commonly developed in the Dashwoods Subzone, approximately 1.5 to 2 km east of the Humber Zone–Dunnage Zone boundary. The tectonite zone (up to 200 m wide) is commonly characterized by strongly foliated tonalite having numerous mafic xenoliths (Plate 2) that are intruded by granodiorite. This granodiorite is generally xenolith-free, coarser grained and less deformed than the tonalite and has been interpreted to have intruded the tonalite during deformation (Plate 3). The tectonite zone strikes northeast and dips steeply (~85°) southeast having stretching lineations plunging down-dip (Figures 2c and d). Shear-sense indicators such as cm-scale winged objects, sheath folds (Plate 4) and S-folds indicate both a normal and a reverse sense of movement, which suggests a complex deformational history for this tectonite.

THE LITTLE GRAND LAKE FAULT

The Little Grand Lake Fault as defined by Piasecki *et al.* (1990) is an east–west-trending structure in the Dunnage Zone that separates the Notre Dame Subzone to the north from the Dashwoods Subzone to the south. The southern part of the Notre Dame Subzone consists of volcanic and sedimentary rocks of the Glover Group, which have been metamorphosed under greenschist-facies conditions, and weakly deformed and metamorphosed rocks of the Topsails Igneous suite. The northern part of the Dashwoods Subzone consists of granodioritic plutons, migmatitic metasedimentary rocks and gabbroic agmatites (Whalen, 1993).

The main movement along the Little Grand Lake Fault is northward-directed thrusting, which is observed on localized, steep south-dipping east–west-striking ductile shear zones within a 1.5-km-wide zone in the Dashwoods Subzone rocks (Figure 2e). This narrow belt of multiple shear zones represents the boundary between the two subzones (Piasecki *et al.*, 1990). The widest shear zone, located nearest to Little Grand Lake, is at least 40 m wide. Each shear zone is characterized by S-C' mylonites and shear bands. The foliation is mainly defined by muscovite crystals and the stretching lineation by quartz (and to a lesser extent feldspar)-aggregates. Mineral and stretching lineations are downdip closest to Little Grand Lake (at UTM - 444263E 5378402N)[2], but to the south, the plunge swings east and the angle of the plunge gradually shallows and eventually becomes subhorizontal (Figure 2f; UTM 442824E 5377877N). Strain intensity decreases toward the south, where mylonite zones become sparser and narrower. A possible interpretation of this strain pattern is a triclinic transpressional shear-zone model as described by Lin *et al.* (1998), but more data is needed to test this hypothesis. Shear-sense indicators show a clear south-side up movement, and having a dextral component where lineations have an oblique plunge. These shear zones are best developed in

[2] All UTM grid references are in NAD 27 coordinates.

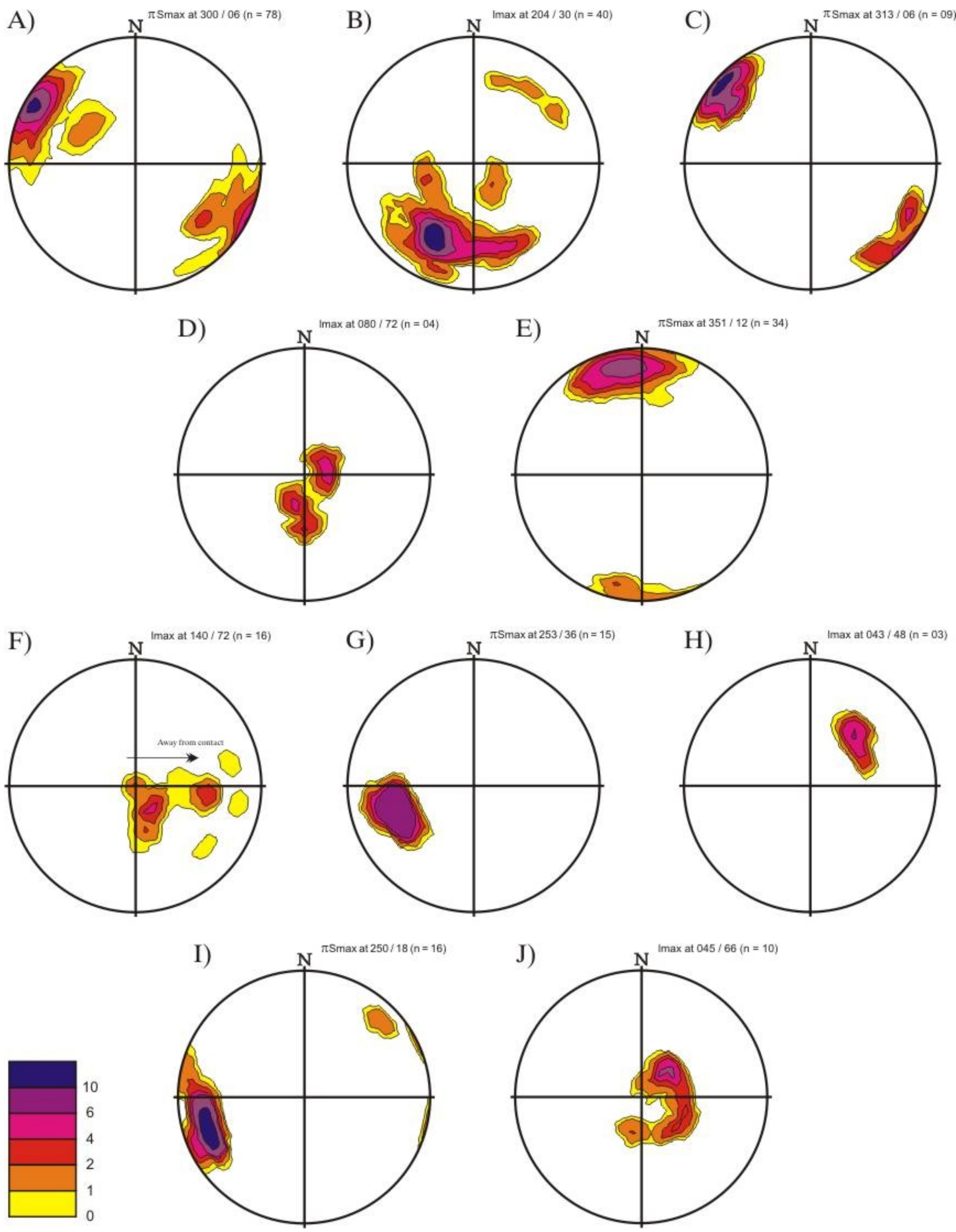

Figure 2. *Equal area lower hemisphere projection of poles to foliation and lineations of the structures discussed in the text:(A) poles to the foliation along the greenschist-facies mylonites of the Cabot Fault Zone; (B) mineral and stretching lineations of the greenschist-facies mylonites of the Cabot Fault Zone; (C) poles to foliation of the amphibolite-facies high-strain zone in the Cabot Fault Zone; (D) stretching lineations in the amphibolite-facies high-strain zone in the Cabot Fault Zone; (E) poles to foliation along the Little Grand Lake Fault; (F) mineral and stretching lineations in mylonites of the Little Grand Lake Fault; (G) poles to the foliation in rocks of the Caribou Brook Shear Zone; (H) mineral and stretching lineations in rocks of the Caribou Brook Shear Zone; (I) poles to the foliation in rocks of the Tulks Pond Shear Zone; and (J) stretching and mineral lineations in rocks of the Tulks Pond Shear Zone. Density contours using the Small Circle method with a grid resolution of 15.*

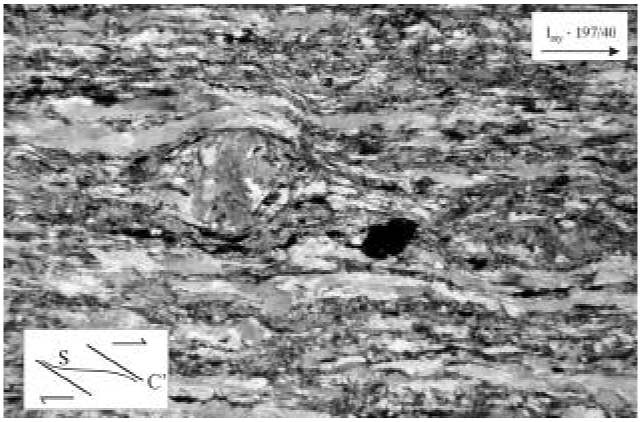

Plate 1. *A photomicrograph of an S-type Kfs-porphyroclast and shear bands in a greenschist-facies mylonite from the Cabot Fault Zone, indicating oblique (dextral and W-side up) movement (UTM - 423805E 5375634N). Photograph is taken with P-CS (Sénarmont Compensator). The field of view is approximately 4 mm.*

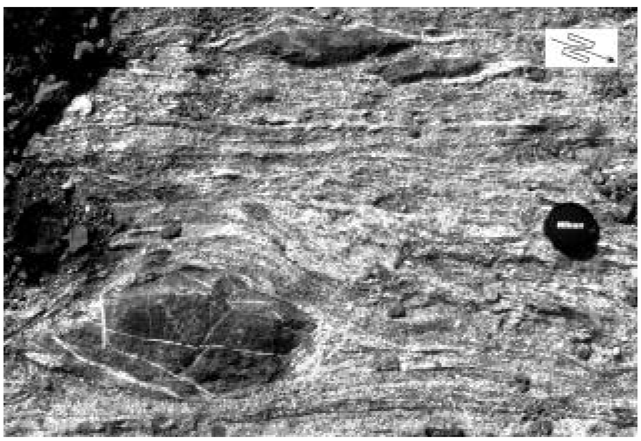

Plate 2. *Field photograph showing an outcrop of the amphibolite-facies high-strain zone with xenoliths in the Cabot Fault Zone (UTM - 393834E 5338563N).*

granitoid rocks (*see also* Whalen *et al.*, 1993), but smaller mylonitic zones also occur within the mafic agmatites. The Little Grand Lake Fault could not be traced west of the Cabot Fault Zone and hence, is probably truncated by it.

A second set of less penetrative structures has been observed in the area around Little Grand Lake. A granite of the Topsails Igneous suite (about 429 Ma; Whalen *et al*., 1987), typical of the Notre Dame Subzone, has been faulted and juxtaposed with a granodiorite pluton that is considered to be part of the Dashwoods Subzone (Whalen *et al*., 1993; Whalen, 1993). A fault breccia has developed in the lower part of the Topsails granite (UTM - 460510E 5384054N), but the sense of movement on the fault is poorly constrained, due to the lack of kinematic indicators. Considering its dip of about 40°N (Currie and van Berkel, 1992; Whalen *et al*., 1993) and the absence of Topsails granites in the Dashwoods Subzone, the fault probably accommodated a south-directed reverse sense of movement.

Plate 3. *Field photograph of syntectonic intrusion of granodiorite into the amphibolite-facies high-strain zone of the Cabot Fault Zone (UTM - 393679E 5338678N).*

Several workers have ascribed both movements that occur in the Little Grand Lake area to take place on the Little Grand Lake Fault (Currie and van Berkel, 1992; Whalen *et al*., 1993). However, the attitude of the fault systems (subvertical for the south-over-north ductile movement and dipping 40° north for the brittle–ductile north-over-south movement), suggests that they represent two different sets of structures, which are merely spatially related or coincidental at the surface. This interpretation is further supported by the observation that several north-over-south thrusts comparable with the latter movement exist in the southern part of the Notre Dame Subzone as far as 10 km north from the Little Grand Lake Fault (Z.A. Szybinski, S.J. Piercey and G.A. Jenner, unpublished GSC map).

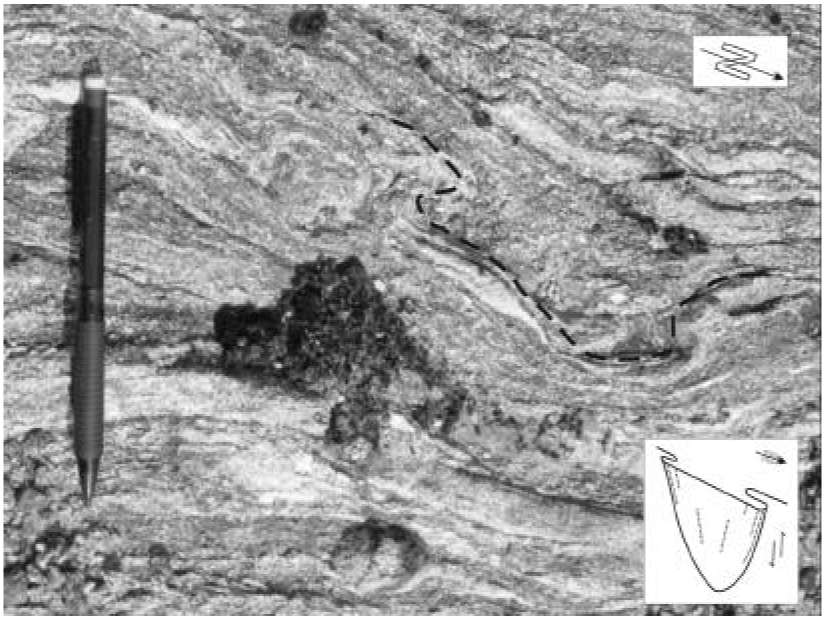

Plate 4. *Field photograph of a sheath fold (just right of the centre) in the amphibolite-facies high-strain zone of the Cabot Fault Zone. The sheath fold indicates a west-side up movement, as depicted in the sketch (cf. Figure 5.9 in Passchier and Trouw, 1996) (UTM - 393834E 5338563N).*

Considering the observations and structural relationships described above, the term Little Grand Lake fault should be restricted to the ductile south-over-north movement, which brings up Dashwoods Subzone rocks relative to those of the Notre Dame Subzone (cf. Piasecki *et al.,* 1990). The brittle north-over-south movement remains unnamed at present.

THE CARIBOU BROOK SHEAR ZONE, TULKS POND SHEAR ZONE AND DISAPPOINTMENT HILL COMPLEX

Two high-strain zones bound the Disappointment Hill complex (Currie and van Berkel, 1992) in the internal Humber Zone. These high-strain shear zones are herein termed the Caribou Brook Shear Zone and the Tulks Pond Shear Zone (Figure 3).

The Caribou Brook Shear Zone strikes northwest–southeast and juxtaposes igneous and metamorphic rocks of the DHC with relatively undeformed anorthositic gabbro of the Steel Mountain complex (Figure 3). A cross-section through the shear zone is exposed at the end of a lumber road (UTM - 415179E 5381974N), 2 km south of Disappointment Hill. Here, the shear zone is approximately 100 m wide and is prevalently developed in the Steel Mountain anorthosite. The main foliation within the shear zone dips 54° toward the east-northeast and the mineral (hornblende) and stretching (feldspar-aggregates) lineations are downdip (Figures 2g and h). The strain progressively increases toward the upper part of the shear zone, near the contact with rocks of the structurally overlying DHC. Shear-sense indicators, such as deformed veins (Plate 5) and shear bands indicate a reverse, east-side-up movement, i.e., the DHC is thrust over the Steel Mountain complex. The northwestern continuation of the Caribou Brook Shear Zone is most probably truncated by a north–south-striking brittle fault that bounds the DHC to the west (Figure 3). This brittle fault is poorly exposed, but is apparent topographically and on aerial photographs as a lineament in the form of a steep cliff next to a valley. The southeastern continuation of the Caribou Brook Shear Zone is constrained, based on exposures of anorthosite (Steel Mountain complex) to the southwest and tonalite, gabbro and diorite exposures (DHC) to the northeast (*see* Figure 3).

The Tulks Pond Shear Zone is a north-northwest–south-southeast-striking structure that extends for several kilometres south of

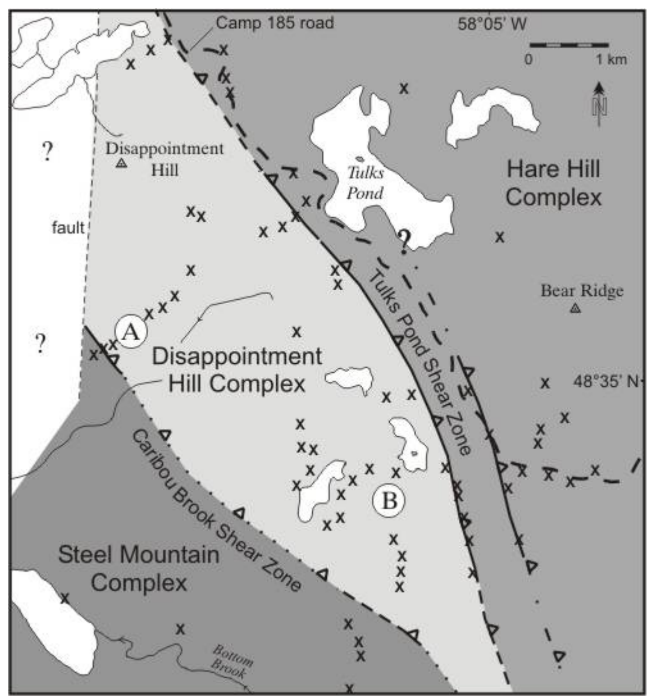

Figure 3. *Simplified geological map of the area surrounding the Disappointment Hill complex. Disappointment Hill complex consists of metaplutonic rocks; Hare Hill complex consists of alkali granites and granitoid gneisses; Steel Mountain complex consists of anorthosites and gabbros. Topographical nomenclature after Martineau (1980). (A) occurrence of agmatite (further explanations in text); (B) occurrence of a ca. 431 Ma gabbro; and (x) outcrop.*

Plate 5. *Field photograph of a sheared and boudinaged quartz vein within the Caribou Brook Shear Zone indicating east-over-west shearing: rocks of the Disappointment Hill complex are thrust westward over anorthositic rocks of the Steel Mountain complex (UTM - 415179E 5381974N).*

Plate 6. *Field photograph of an S-C' mylonite in the Tulks Pond Shear Zone showing an east-side up movement (UTM - 418148E 5382741N).*

Tulks Pond. Topographically, this zone forms a lineament along a ridge west of the Camp 185 road (Figure 3). It juxtaposes granites and granitoid gneisses of the Hare Hill complex (Currie and van Berkel, 1992) and gabbros and tonalites of the DHC (Figure 3). The main foliation strikes north-northwest and dips steeply east (Figure 2i). Mineral and stretching lineations plunge steeply (Figure 2j). Shear-sense indicators such as shear bands (Plate 6) and drag structures indicate a reverse, east-side-up sense of movement. This indicates that the granitoid rocks of the Hare Hill complex are thrust over the DHC. The Tulks Pond Shear Zone has not been observed northwest of Tulks Pond (Figure 3). To the south, the structure is assumed to continue at least up to Bottom Brook.

The DHC (Figure 3), comprising an area of 20 km^2, consists mainly of (meta)plutonic rocks such as tonalite, granodiorite, granitoid gneiss, gabbro and charnockite (*see also* Owen and Currie, 1991; Currie and van Berkel, 1992). A distinctive agmatite is exposed in the western part of the DHC (Figure 3, location A; Plate 7a) where xenoliths of mafic (gabbroic) to ultramafic (pyroxenite) composition occur within a coarse crystalline granodiorite matrix having distinctive blue quartz crystals. The xenoliths vary in size (< 50 cm) and shape (elongate-angular to rounded) and locally constitute up to 75 percent of the rock. Several of the xenoliths contain a fabric that does not continue into the matrix and is thus interpreted to have formed before the intrusion of the granodiorite. This lithological association closely resembles the Southwest Brook complex in the Dashwoods Subzone (Dunning *et al*., 1989) and other agmatites exposed in the northern and central part of the Dashwoods Subzone (Plates 7b and c). Furthermore, Whalen *et al.* (1997) emphasized the strong geochemical and isotopic similarities between the Ordovician Cormacks Lake charnockite in the Dashwoods Subzone (460 ± 10 Ma, U–Pb on zircon, Currie

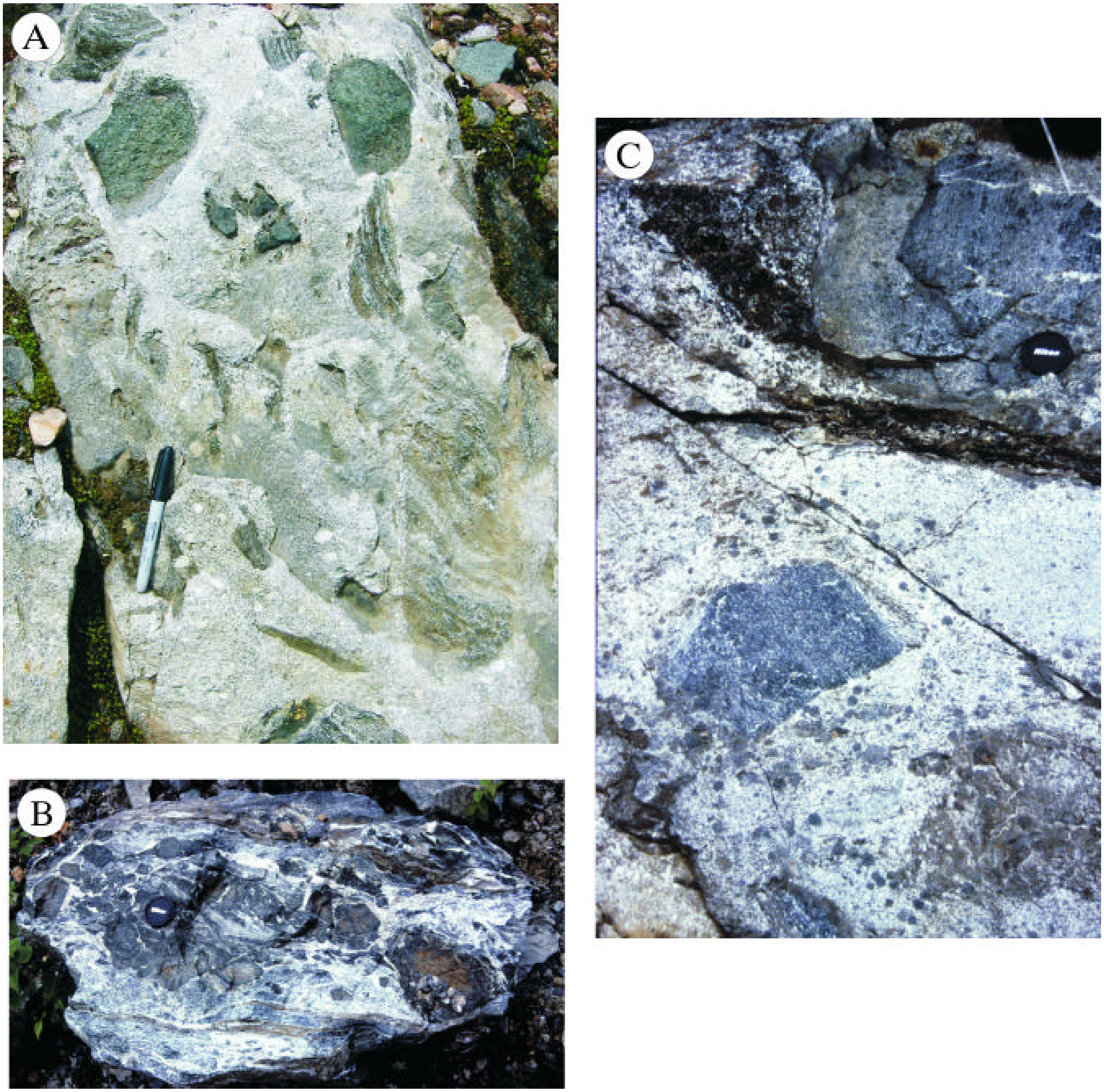

Plate 7. *Three agmatites that are comparable in lithology and xenolith content and that are possibly genetically related: (A) Field photograph of an agmatite exposed in the Disappointment Hill complex (UTM - 415713E 5382145N); (B) Field photograph of an agmatite exposed in the northern part of the Dashwoods Subzone (UTM 433176E 5381720N); (C) Field photograph of an agmatite exposed in the central part of the Dashwoods Subzone (UTM 383524E 5318911N).*

et al., 1992) and the charnockite in the DHC. If this correlation is correct, it implies that the DHC is an exotic slice of Dashwoods in the Humber Zone.

DISCUSSION

The Little Grand Lake Fault is truncated by the Cabot Fault Zone to the west and presumably the Star Lake Shear Zone (i.e., the Lloyds River Fault in Whalen *et al*., 1993) to the east. Exposure east of Little Grand Lake is poor and therefore this supposition could not be verified in the field. The inferred position of the fault between Little Grand Lake and the Star Lake Shear Zone is therefore solely based on the associated contrast in magnetic anomalies shown by the aeromagnetic map, which clearly outlines the surface trace (Figure 4) of the Little Grand Lake Fault. However, imme-

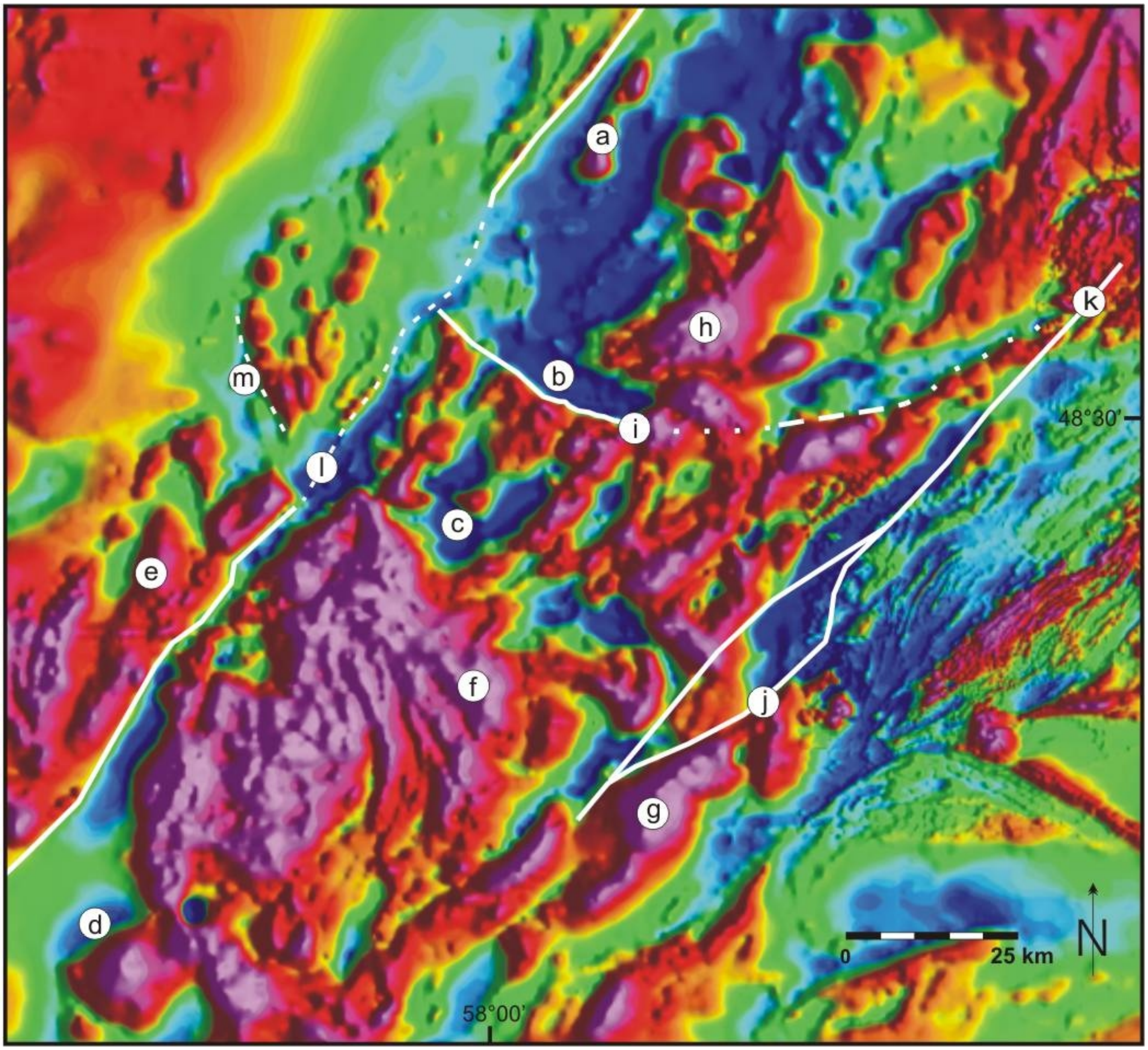

Figure 4. *Simplified map of aeromagnetic anomalies in the Grand Lake - Little Grand Lake area: (a) Glover Island; (b) Little Grand Lake; (c) Island Pond; (d) Mica Pond; (e) Steel Mountain Complex; (f) Cormacks Lake Complex; (g) Annieopsquatch Ophiolite Complex; (h) Topsails granite; (i) Little Grand Lake Fault; (j) Lloyds River Fault - Red Indian Line; (k) Star Lake Shear Zone; (l) Cabot Fault Zone, and (m) Caribou Brook Shear Zone.*

diately east of Little Grand Lake, its trace is obscured by a high positive anomaly. The anomaly is most probably due to the occurrence of a granodiorite pluton (*see* Whalen, 1993) that either cuts the Little Grand Lake Fault or makes a small jog. Granodiorite and tonalite plutons of the Dashwoods Subzone generally (but not necessarily all) have an Early to Middle Ordovician age (488 to 456 Ma Dunning *et al*., 1989; Dubé *et al*., 1996; Whalen *et al*., 1997). The age of the Glover Group in the adjacent Notre Dame Subzone is Arenig (about 480 Ma), based on a graptolite fauna found in black shales of the Glover Group (Williams, 1992; Swinden *et al*., 1997). East of Little Grand Lake, the trace of the Little Grand Lake Fault curves anticlockwise toward the Star Lake Shear Zone (Figure 4), but does not occur east of this major structure. The Star Lake Shear Zone is the northeasterly extension of the Lloyds River Fault (Figure 4; Lissenberg and van Staal, 2002) and the earliest known age of movement on the Lloyds River Fault is about 468 Ma (^{40}Ar–^{39}Ar on hornblende, C.J. Lissenberg and V.J. McNicoll, unpublished data, 2002). Considering the ages and relationships described above, the age of north-directed thrusting accommodated by the Little Grand Lake Fault is constrained between about 480 and 468 Ma. This age constraint contrasts with the interpretation by Whalen *et al*. (1993), who favoured a Silurian age for the movement on the Little Grand Lake Fault.

An upper intercept U–Pb age of 1498 +9/-8 Ma of the DHC charnockite was interpreted by Currie *et al.* (1992) to represent the time of its intrusion. The charnockitic mineralogy of the felsic body was interpreted to form as a result of emplacement under granulite-facies conditions. This led to correlations of the DHC with the (dominantly) Mesoproterozoic Long Range Inlier in western Newfoundland (Owen and Currie, 1991). The upper and lower intercepts at present are unreliable, because the regression line has a poor fit, suggesting significant complexity in the zircon systematics. Alternatively, this date could be an age of inheritance. Moreover, the strong resemblance of part of the DHC, such as the agmatites that are exposed in the southwestern corner of the DHC, to the Middle Ordovician rocks of the Southwest Brook complex (Dunning *et al*., 1989) of the Dashwoods Subzone, and the allochtonous nature of the DHC, in general, suggest that the DHC could be a transported slice of mainly Paleozoic Dashwoods Subzone rocks, comparable with the Coney Head complex in northern Newfoundland (Dunning, 1987).

A Dunnage Zone provenance of the DHC is also consistent with a preliminary $^{40}Ar–^{39}Ar$ hornblende age of about 431 Ma of a hornblende–gabbro exposed in the eastern part of the DHC (B on Figure 3; $^{40}Ar–^{39}Ar$ on hornblende, A.G. Brem and V.J. McNicoll, unpublished data, 2002). Early Silurian hornblende-bearing gabbro and diorite are common in the western part of the Dunnage Zone (e.g., Main Gut gabbro, Rainy Lake complex), but are very rare in the Humber Zone. One known exception is the ca. 431 Ma Taylor Brook gabbro (Heaman *et al*., 2002), which intrudes the allochthonous Humber Zone basement exposed in the Long Range Inlier and was transported to the west during the Late Silurian and Devonian (Cawood and Williams, 1988; Waldron *et al*., 1998). Testing our hypothesis that the DHC has a Dunnage Zone provenance is currently in progress by means of geochemical and geochronological studies.

The early movements in the Cabot Fault Zone, which are accommodated by the amphibolite-facies tectonites, are constrained by the age of the syn-kinematic granodiorite described above (Plate 3) and by a syntectonic pegmatite dyke that intrudes similar gneissose granodioritic rocks. The pegmatite intruded during oblique dextral strike-slip (Plates 3 and 4 *in* Brem *et al*., 2002) and yielded a lower intercept age of about 456 Ma (U–Pb on zircon, A.G. Brem and D.W. Davis, unpublished data, 2002), which suggests that ductile shearing in the Cabot Fault Zone postdates the Early to Middle Ordovician movements deduced for the Little Grand Lake Fault. This interpretation is consistent with earlier conclusions that the Little Grand Lake Fault is truncated by the Cabot Fault Zone.

The trace of the Cabot Fault Zone is clearly visible on the aeromagnetic map (Figure 4). West of Glover Island, the western part of the Cabot Fault Zone is characterized by a positive magnetic anomaly that can be traced along-strike for the entire length of the study area. Orientations of the foliations and lineations in this muscovite-rich granite are similar to those present in the greenschist-facies mylonites of the Cabot Fault Zone (Figure 2a; Brem *et al*., 2002). The metasedimentary rocks and the DHC in the internal Humber Zone experienced thrusting during the Early to Late Silurian (Cawood and van Gool, 1998; Brem *et al*., 2002). These west-vergent thrust faults are truncated by the Cabot Fault Zone, which suggests that the Cabot Fault Zone accommodated significant post Early Silurian deformation.

SUMMARY OF THE MOVEMENT HISTORY

The following kinematic history is proposed for the Humber Zone–Dunnage Zone boundary: (1) In the Early to Middle Ordovician, the Dashwoods Subzone moved upward relative to the Notre Dame Subzone by thrusting along the Little Grand Lake Fault. This resulted in the juxtaposition of medium- to high-grade rocks of the Dashwoods Subzone with lower grade rocks of the Notre Dame Subzone; (2) In the late Middle Ordovician (ca. 465 to 460 Ma), strong deformation and metamorphism occurred in the Cabot Fault Zone and in the Dashwoods Subzone. No evidence of high-grade metamorphism is present in the Humber Zone. Thus, the high-strain tectonite zone in the Cabot Fault Zone probably accommodated uplift of the Dashwoods Subzone relative to the Humber Zone. This time of deformation is concurrent with west-vergent thrusting that led to the emplacement of high-level allochtons in the Humber Zone (Waldron *et al*., 1998). Speculating, this event also led to the initial emplacement of parts of the Dunnage Zone (DHC ?) above rocks of the internal Humber Zone. At the beginning of the Late Ordovician (ca. 456 Ma, Caradoc) there is a change in sense of movement accommodated by the Cabot Fault Zone to oblique dextral, bringing rocks of the Humber Zone up relative to the Dashwoods Subzone; (3) Thrusting in the internal Humber Zone continued until at least the Late Silurian, during which, thrust sheets having metasedimentary rocks are emplaced farther westward (*see also* Waldron *et al.*, 1998; Cawood and van Gool, 1998). Final emplacement of the DHC probably also took place during this period; and (4) During the Late Silurian–Devonian (?) the Cabot Fault Zone accommodated progressively more oblique dextral strike–slip movement, presumably after formation of a steeply dipping crustal-scale fault. This phase of faulting, which moved the Dunnage Zone down relative to the Humber Zone, cut through the earlier formed thrust stack of the Dunnage and Humber zones and produced (rootless?) outliers of Dunnage Zone rocks, such as the DHC, in the Humber Zone.

ACKNOWLEDGMENTS

The work forms part of a project funded by the Geological Survey of Canada's Targeted Geoscience Initiative (TGI) 000018. It is also supported by an NSERC grant to Lin. The first author wishes to thank Melissa Battler for her assistance during the field season. Sally Pehrsson and Johan Lissenberg are thanked for fruitful discussions concerning the geology and giving insights in their data. The latter is also thanked for his contribution to Figure 4. Sally Pehrsson, John Percival and Baxter Kean are thanked for critically reading the manuscript. This is Geological Survey of Canada contribution # 2002 177.

REFERENCES

Brem, A.G., Lin, S. and van Staal, C.R.
2002: Humber Zone - Dunnage Zone relationships and the Long Range Fault, south of Grand Lake, western Newfoundland: preliminary results: *In* Current Research. Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report: 02-1, pages135-144.

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

Cawood, P.A., and van Gool, J.A.M.
1998: Geology of the Corner Brook - Glover Island region, Newfoundland. Geological Survey of Canada, Bulletin No. 427, 96 pages.

Cawood, P.A., van Gool, J.A.M. and Dunning, G.R.
1996: Geological development of Eastern Humber and Western Dunnage zones: Corner Brook - Glover Island region, Newfoundland. Canadian Journal of Earth Sciences, Volume 33, pages 182-198.

Cawood, P.A. and Williams, H.
1988: Acadian basement thrusting, crustal delamination, and structural styles in and around the Humber Arm Allochthon, western Newfoundland. *In* Geology (Boulder), volume 16, pages 370-373.

Currie, K.L. and van Berkel, J.T.
1992: Notes to accompany a geological map of the southern Long Range, southwestern Newfoundland. Geological Survey of Canada, Report: 91-10, 13 pages.

Currie, K.L., van Breemen, O., Hunt, P.A. and van Berkel, J.T.
1992: Age of high-grade gneisses south of Grand Lake, Newfoundland. Atlantic Geology, Volume 28, pages 153-161.

Dubé, B., Dunning, G.R., Lauzière, K. and Roddick, J.C.
1996: New insights into the Appalachian Orogen from geology and geochronology along the Cape Ray fault zone, southwest Newfoundland. Geological Society of America Bulletin, Volume 108, pages 101-116.

Dunning, G.R.
1987: U–Pb geochronology of the Coney Head complex, Newfoundland. Canadian Journal of Earth Sciences, Volume 24, pages 1072-1075.

Dunning, G.R., O'Brien, S.J., Colman-Sadd, S.P., Blackwood, R.F., Dickson, W.L., O'Neill, P.P. and Krogh, T.E.
1990: Silurian orogeny in the Newfoundland Appalachians. Journal of Geology, Volume 98, pages 895-913.

Dunning, G.R., Wilton, D.H.C. and Herd, R.K.
1989: Geology, geochemistry and geochronology of a taconic batholith, southwestern Newfoundland. Transactions of the Royal Society of Edinburgh: Earth Sciences, Volume 80, pages159-168.

Fryer, B.J., Greenough, J.D. and Owen, J.V.
1997: Iapetus Ocean floor stuffed into a suture zone: xenolith ND isotopic evidence for Dunnage-equivalent basement in central Newfoundland. Canadian Journal of Earth Sciences, Volume 34, pages 1392-1400.

Heaman, L.M., Erdmer, P. and Owen, J.V.
2002: U–Pb geochronologic constraint on the crustal evolution of the Long Range Inlier, Newfoundland. Canadian Journal of Earth Sciences, Volume 39, pages 845-865.

Herd, R.K. and Dunning, G.R.
1979: Geology of Puddle Pond map area, southwestern Newfoundland. *In* Current Research, Part A. Geological Survey of Canada, Paper 79-1A, pages 305-310.

Hibbard, J.
1983: Kinematics of Acadian deformation in the Northern and Newfoundland Appalachians. Journal of Geology, Volume 102, pages 215-228.

Knapp, D.A.
1982: Ophiolite emplacement along the Baie Verte–Brompton line at Glover Island, western Newfoundland. Unpublished Ph.D. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, Canada, 338 pages.

Lin, S., Jiang, D. and Williams, P.F.
1998: Transpression (or transtension) zones of triclinic symmetry: natural example and theoretical modelling. Geological Society, Special Publication No. 135, pages 41-57.

Lissenberg, C.J. and van Staal, C.R.
2002: The relationships between the Annieopsquotch ophiolite belt, the Dashwoods block and the Notre Dame arc in southwestern Newfoundland. *In* Current Research. Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report: 02-1, pages 145-153.

Martineau, Y.
1980: The relationships among rock groups between the Grand Lake thrust and Cabot Fault, west Newfoundland. Unpublished M.Sc thesis, Memorial University of Newfoundland, St. John's, Newfoundland, Canada, 150 pages.

Owen, J.V. and Currie, K.L.
1991: The Disappointment Hill complex: Proterozoic granulites in southwestern Newfoundland. Transactions of the Royal Society of Edinburgh, Earth Sciences, Volume 82, pages 55-63.

Owen, J.V. and Greenough, J.D.
1994: Recrystallization of quartzofeldspathic rocks in a Paleozoic thrust belt, Grand Lake, Newfoundland; implications for garnet-biotite thermometry. Lithos, Volume 33, pages 225-239.

Passchier, C.W. and Trouw, R.A.J.
1996: Microtectonics. Springer-Verlag, Berlin, 289 pages.

Piasecki, M.A.J.
1988: Strain-induced mineral growth in ductile shear zones and a prelimenary study of ductile shearing in western Newfoundland. Canadian Journal of Earth Sciences, Volume 25, pages 2118-2129.

Piasecki, M.A.J., Williams, H. and Colman-Sadd, S.P.
1990: Tectonic relationships along the Meelpaeg, Burgeo and Burlington LITHOPROBE transects in Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 90-1, pages 327-339.

Swinden, H.S., Jenner, G.A. and Szybinski, Z.A.
1997: Magmatic and tectonic evolution of the Cambrian-Ordovician Laurentian margin of Iapetus: Geochemical and isotopic constraints from the Notre Dame subzone, Newfoundland. *In* The Nature of Magmatism in the Appalacian Orogen. *Edited by* A.K. Sinha, J.B. Whalen and J.P. Hogan. Geological Society of America, Memoir No. 191, pages 337-366.

Szybinski, Z.A., House, S. and Jenner, G.A.
1995: Stratigraphy and structure of the Glover Group, Grand Lake - Little Grand Lake area, Newfoundland. *In* Current Research, Part E. Geological Survey of Canada, Report 1995-E, pages 245-251.

Van Berkel, J.T. and Currie, K.L.
1988: Geology of the Puddle Pond (12 A/5) and Little Grand Lake (12 A/12) map areas, southwestern Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division, Report 88-1, pages 99-107.

Waldron, J.W.F., Anderson, S.D., Cawood, P.A., Goodwin, L.B., Hall, J., Jamieson, R.A., Palmer, S.E., Stockmal, G.S. and Williams, P.F.
1998: Evolution of the Appalachian Laurentian margin; Lithoprobe results in western Newfoundland. Canadian Journal of Earth Sciences, Volume 35, pages 1271-1287.

Waldron, J.W.F. and van Staal, C.R.
2001: Taconian orogeny and the accretion of the Dashwoods block: a peri-Laurentian microcontinent in the Iapetus Ocean. Geology, Volume 29, pages 811-814.

Wendland, O.
2002: Analysis of amphibolite dykes and granitoid rocks from the southern Long Range mountains, Newfoundland Canada: geochemical signature determination and tectonic discriminaton. Unpublished B.Sc. thesis, Keele University, Stoke-upon-Trent, England, 50 pages.

Whalen, J.B.
1993: Geology, Little Grand Lake, Newfoundland. Geological Survey of Canada, Open File Report: 2736, 1 sheet.

Whalen, J.B., Currie, K.L. and Piasecki, M.A.J.
1993: A re-examination of relations between Dunnage subzones in southwest Newfoundland. *In* Current

Research, Part D. Geological Survey of Canada, Paper 93-1D, pages 65-72.

Whalen, J.B., Currie, K.L. and van Breemen, O.
1987: Episodic Ordovician-Silurian plutonism in the Topsails Igneous Terrane, western Newfoundland. Transactions of the Royal Society of Edinburgh, Volume 78, pages 17-28.

Whalen, J.B., van Staal, C.R., Longstaffe, Gariépy, C., and Jenner, G.A.
1997: Insights into tectonostratigraphic zone identification in southwestern Newfoundland based on isotopic (Nd, O, Pb) and geochemical data. Atlantic Geology, Volume 33, pages 231-241.

Williams, H.
1979: Appalachian orogen in Canada. Canadian Journal of Earth Sciences, Volume 16, pages 792-807.

1995: Geology of the Appalachian-Caledonian orogen in Canada and Greenland. Geological Survey of Canada, Geology of Canada No. 6, 944 pages.

Williams, H. and St. Julien, P.
1982: The Baie Verte-Brompton line in Newfoundland and regional correlations in the Canadian Appalachians. *In* Current Research, Part A. Geological Survey of Canada, Paper 78-1A, pages 225-229.

Williams, S.H.
1992: Lower Ordovician (Arenig - Llanvirn) graptolites from the Notre Dame Subzone, central Newfoundland. Canadian Journal of Earth Sciences, Volume 29, pages 1717-1733.

Wilson, T.
1962: Cabot Fault, an Appalachian equivalent of the San Andreas and Great Glen Faults and some implications for continental displacement. Nature, Volume195, pages 135-138.