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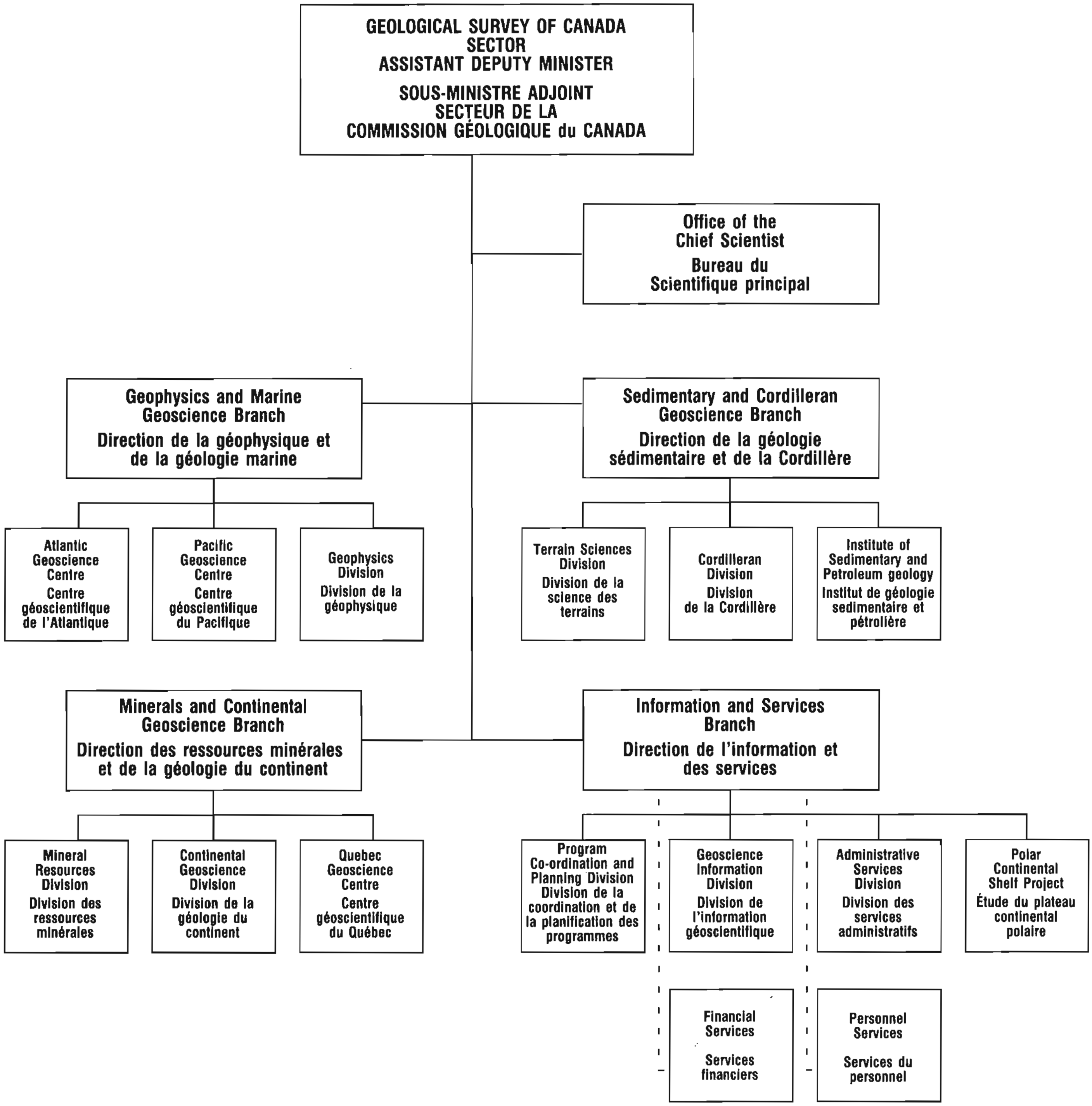
Cover description

Upright, shallowly east plunging F_2 folds in mylonitic gabbros of the ophiolitic Devereaux Formation near Pointe Verte, northern New Brunswick. These highly strained gabbros form the structurally lowest part of an allochthonous fragment of back-arc oceanic crust. Photo by C.R. van Staal and J.P. Langton.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Geology of the southwest arm of Grand Lake, western Newfoundland[1]

M.A.J. Piasecki[2]

Piasecki, M.A.J., Geology of the southwest arm of Grand Lake, western Newfoundland; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 1-8, 1991.

Abstract

Along the southwest arm of Grand Lake a gneiss complex of presumed Grenville affinity contains a marble-quartzite unit and is cross-cut by a mafic dyke swarm. Overlying metasedimentary cover rocks, presumed correlative to the Fleur de Lys Supergroup, contain a complex metaconglomerate, variable in composition and structural style, which is interpreted as a locally developed basal deposit.

East-over-west Paleozoic ductile shearing, accompanied and overlapped by greenschist to amphibolite-grade metamorphism, retrogressed the gneiss complex and decoupled basement and cover. This sector of the Steel Mountain terrane is allochthonous, thrust northwest over the Humber Platform sub-zone from an original site east of the Long Range fault, and telescoped into a stack of at least three thrust slices.

Résumé

Le long du bras sud-ouest du lac Grand, un complexe gneissique qui semblerait présenter certaines affinités avec des roches du Grenville contient une unité à marbre et quartzite, et est recoupé par un essaim de dykes mafiques. Les roches métasédimentaires qui le recouvrent, probablement corrélatives du Supergroupe de Fleur de Lys, contiennent un métaconglomérat complexe, de composition et de style structural variables, que l'on interprète comme étant un dépôt basal accumulé localement.

Un cisaillement ductile survenu d'est en ouest au Paléozoïque, accompagné et suivi d'un métamorphisme allant du faciès des schistes verts au faciès des amphibolites, a causé un rétromorphisme du complexe gneissique ainsi que le désaccouplement du socle et de la couverture sédimentaire. Ce secteur du terrane de Steel Mountain est allochtone, et a été charrié vers le nord-ouest au-dessus de la sous-zone de la plate-forme de Humber, à partir d'un site initial situé à l'est de la faille de Long Range, et a formé par télescopage un empilement d'au moins trois nappes de charriage.

[1] Contribution to the Canada-Newfoundland Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada, Continental Geoscience Division.

[2] Department of Geology, University of Keele, Keele, Staffordshire ST5 5BG

INTRODUCTION

Grand Lake straddles the Humber-Dunnage zone boundary (Williams, 1979; Williams et al., 1989), and its southwestern arm provides a section across the eastern flank of the Humber zone. This section spans the Cambrian(?) Grand Lake Brook Group of the Humber Platform subzone (Williams and Cawood, 1988), and gneisses and metasedimentary rocks of the Steel Mountain sub-zone (van Berkel et al., 1986; Currie and Piasecki, 1989). It crosses the Grand Lake thrust, the boundary between the two sub-zones of the Humber zone, and terminates in the complex Long Range (Cabot) fault zone, the Humber-Dunnage boundary (Piasecki et al, 1990). This report considers the tectonic and metamorphic history of rocks of this 11 km section across the Steel Mountain sub-zone.

THE GNEISS COMPLEX

The gneiss complex exposed along Grand Lake (Fig. 1) comprises a series of well-banded, light coloured granitic (sensu lato) hornblende-biotite gneisses interbanded with minor psammitic migmatitic paragneisses, which commonly contain "dents de cheval" feldspars up to 3 cm in size. Locally the complex contains massive feldspathic quartzite, marble, and semipelitic paragneiss. The gneisses commonly contain necked and boudinaged bands of mafic amphibolite gneiss from 3 cm to scores of metres thick. Gneisses and amphibolites share the same complex fabric, including rootless intrafolial folds of gneissosity, indicating more than one phase of isoclinal folding and gneissification. These rocks are interlayered with abundant, less banded granitoids of nebulitic aspect that share some elements of the composite foliation of the gneisses from which they appear to have been derived, but grade locally into weakly foliated potassic granite and pegmatite. The youngest members of the complex form minor bodies of hybrid microdiorite which cross-cut the gneiss (locality b in Fig. 1) but are foliated parallel to the composite gneissic foliation. This microdiorite contains prominent grains of sphene mantled by feldspar.

The gneiss complex contains, in addition to local patches of migmatitic paragneiss and thin schlieren of migmatized quartzite, a mappable unit of massive quartzite and marble with skarn, exposed in two vertical belts separated by gneiss (Fig. 1). Very coarse quartzite of the western belt becomes migmatitic toward the west and passes into a quartz-rich nebulitic granite gneiss. The contact is transitional and devoid of mylonite fabrics, but the quartzite is disrupted into large blocks in a quartzitic matrix, and the gneiss is intensely folded. The quartzite is in contact with a coarse marble which is bounded to the east by granitic gneiss. Marble boundaries lack mylonitic fabrics, but have the appearance of old fault-zones filled with metamorphosed fault breccias. Adjacent to these contacts the marble becomes fragmented and develops diopside-amphibole-epidote-plagioclase-idocrase-carbonate skarns. Thin veins of white biotite granite are associated with these contacts and may be responsible for skarn formation. The main body of marble is broken into large skarn-bounded blocks.

The eastern belt of quartzite and marble is sheared and thinner than the western belt. Massive quartzite is interleaved with thin units of coarse marble that are disrupted into large ovoid blocks bounded by anastomosing

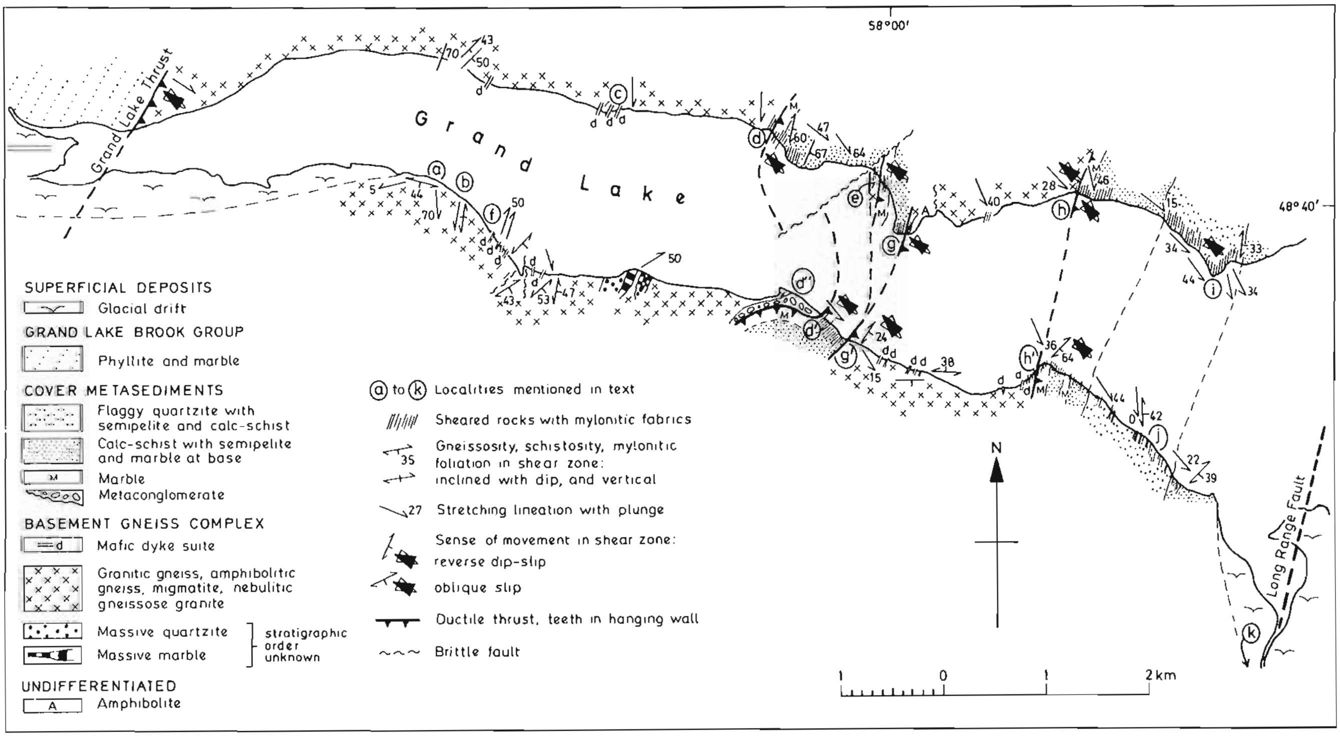


Figure 1. Simplified geology along the shore of the southwestern arm of Grand Lake. 'a' to 'j' refer to localities mentioned in the text

fault breccias which have been converted to skarn. One of the marble units is mylonitic and contains an ellipsoidal mass of quartzite 8 m thick.

THE DYKE SUITE

The gneiss complex contains a suite of northeast-southwest trending, variably foliated, mafic dykes, now metamorphosed to biotite amphibolites. The dykes cut the gneissose fabric and isoclinal folds of the gneisses at high angles, but do not appear to be present in the overlying cover sequence. Most of the dykes, which range in thickness from 10 cm to over 2 m, preserve relicts of chilled margins and magmatic textures in their cores, and many small apophyses within the gneiss remain unfoliated. Some of the dykes appear to have been intruded into hot gneisses and exhibit the "Sederholm effect" (melted gneiss injects the dyke).

THE COVER ASSEMBLAGE

The cover metasediments are non-gneissose, non-migmatitic and separated from basement by spectacular ductile mylonites which mark a zone of layer-parallel décollement and thrusting. A typical sequence, in ascending order, comprises marble and calc-silicate marble followed by calcareous and pelitic schist, followed by calcareous semipsammites which grade upward into well bedded quartzite with intervals of semipelitic and calcareous schist. Although many exposures of the metasedimentary rocks are mylonitic, this sequence is repeated in each tectonic sheet, and probably represents an original stratigraphic succession.

In most thrust sheets, marble marks the base of the cover succession and appears to have localized the décollement surface. In the structurally lowest sheet a spectacular metaconglomerate (Fig. 1, d'') occurs between the gneisses and the mylonitic marble. This unit has been traced for 4 km south of Grand Lake (Currie, 1987), but has proved difficult to interpret because of its variability in composition and structural style. At Grand Lake, metaconglomerate can be seen in situ between localities d' and d'' in Figure 1, and as large slumped blocks for a kilometre west of these localities. Where the unit is not migmatized, the matrix is psammitic, locally bedded, and contains thin stringers of minute magnetite grains, probably originally heavy mineral layers, which may be isoclinally folded. However most exposures have been converted to feldspar-porphyroblastic or stromatic migmatites, or permeated by granite to the extent that the matrix has taken on the appearance of massive granite (Fig. 2a, b, and c). The metaconglomerate contains matrix-supported cobbles of granite (Fig. 2a, d) and granitic and psammitic gneisses (Fig. 2d), many of which were mylonitized prior to incorporation into the conglomerate (Fig. 2a, b, c and d). It also contains angular fragments that suggest a tectonically disrupted rock (Fig. 2e).

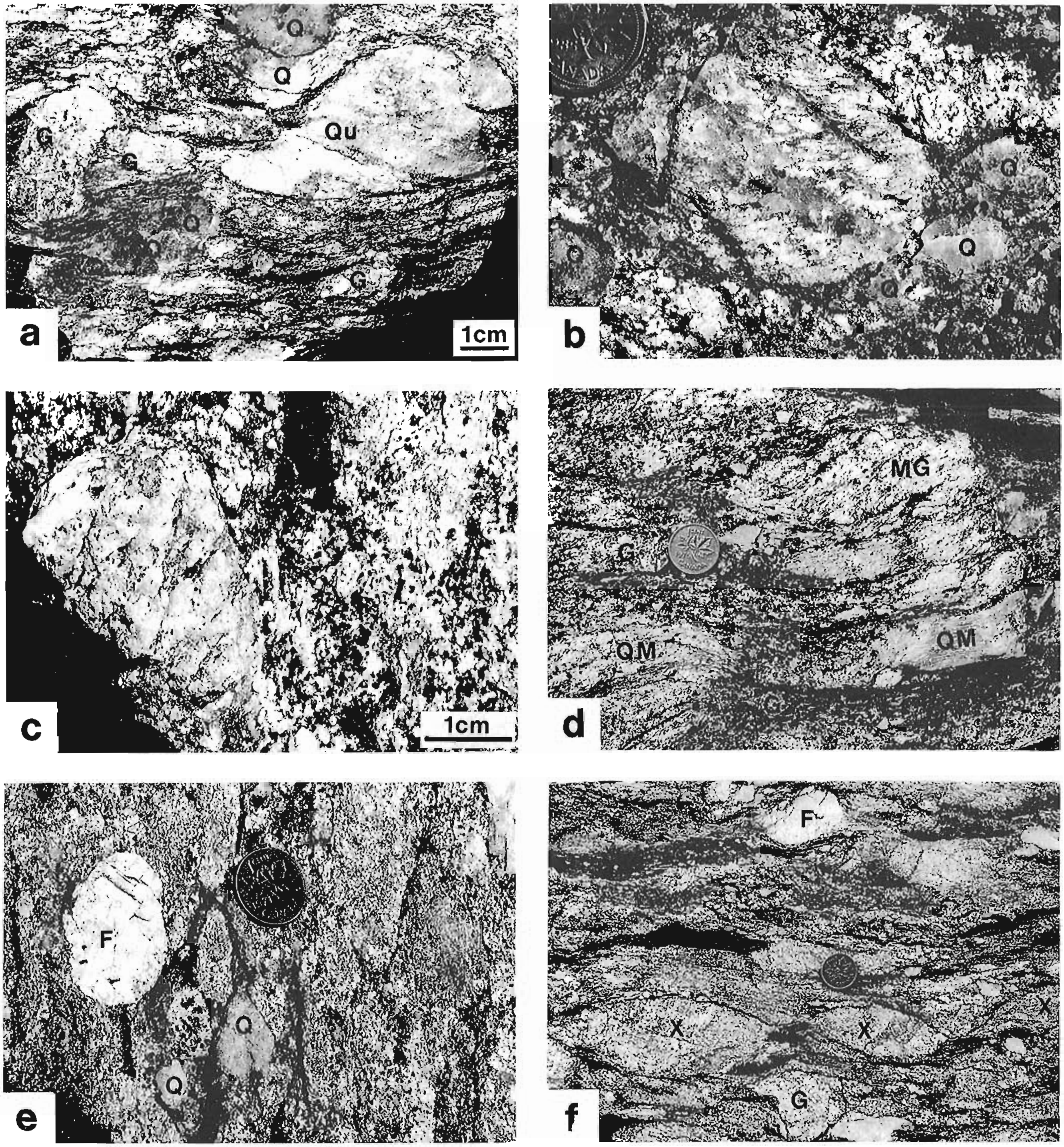
Most outcrops and blocks of metaconglomerate underwent inhomogeneous ductile shearing subsequent to migmatization (Fig. 2a, c and d), producing mylonite along ductile shears greater than 1 cm thick which anastomose around lenticles of less sheared matrix. Cobbles of basement rocks exhibit old fabrics partially transposed along their margins into the younger shear foliation of the matrix (Fig. 2d). Mylonitization was followed by a phase of inhomogeneous extension which pulled the rock apart into fragments from 1 to 50 cm long, with a second generation granular matrix sufficiently mobile to fill gaps between fragments (Fig. 2e, 3a). During this process, elongate clasts necked and boudinaged (Fig. 2d, f), locally producing a crude layering defined by trails of fragments which may vary from old matrix to disrupted granitic or migmatitic clasts.

Ubiquitous, rounded, pebble-like fragments of pure quartz (Fig. 2e) have proved particularly difficult to interpret. Locally they represent rodding and pull-apart of syn-shearing quartz veins formed in the plane of the mylonitic fabric (Piasecki, 1988). These fragments define a stretching lineation parallel to that in the adjacent mylonitic metasediments.

THRUSTING

The southwest arm of Grand Lake lies within the Steel Mountain terrane (Currie and Piasecki, 1989), which here forms a stack of at least three allochthonous, southeast-dipping thrust sheets whose bases lie on, from west to east, the Grand Lake thrust, and thrusts at localities e and g-g' in Figure 1. Each sheet exhibits the gneiss-cover sequence, and a decoupling of basement and cover along a décollement thrust (at d-d', e and h-h' in Fig. 1). This stack of thrust sheets is truncated on the east by the Long Range (Cabot) fault zone which brings low-grade volcanic rocks of the Dunnage zone (Glover Group) against the metasedimentary rocks. The Long Range fault forms a steep, major lineament with a complex history of ductile to brittle movements. On lumber road exposures near Lewaseechjeech Brook (locality k in Fig. 1), older mylonitic fabrics indicating sinistral and vertical movements are overprinted by vertical, brittle movements downthrowing to the west.

Within the thrust stack, basement gneiss distant from the main thrusts contains minor zones of protomylonite and blastomylonite with stretching lineations oriented north-south to northwest-southeast. At major thrusts, mylonitic zones are thicker and more intensely sheared, with stretching lineations generally trending northwest-southeast. Cover rocks between thrust sheets of gneiss are mylonitized for 200 m or more from the tectonic contact, and in the structurally highest sheet an apparent thickness of more than 1 km of quartzite with semipelitic and calcareous intercalations is inhomogeneously sheared to mylonite and blasto-mylonite. Stretching lineations, defined by mineral alignment and rodding of syn-shearing quartz veins, plunge gently to moderately southeast. Trends vary from 123° to 165° with most between 135-145°. Variations of up to 20° are common in exposures free of small folds, suggesting local variations in direction of motion during episodes of movement (c.f. Piasecki, 1988). Small, almost isoclinal, syn-shearing asymmetric folds with one short limb, fold earlier-formed increments of the mylonitic foliation. Their axial surface



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foliation corresponds to later increments of the composite fabric. Post-shearing, open upright folds and crenulations trend approximately normal to the stretching lineation, modifying their plunge, but not their trends.

Fabrics reflecting sense of motion (kinematic markers) abound within the shear zones. They include S-C fabric (Fig. 3b), shear bands, rotated porphyroclasts, granulated porphyroclasts with "tails" of granular aggregate, vergences of small syn-shearing folds, and sense of maximum curvature of fold axes (Berthe et al., 1979; Simpson and Schmid, 1983; Piasecki, 1988). These indicators consistently suggest layer-parallel ductile overthrusting to the northwest. Apparent movements in the opposite sense are indicated in the tectonically highest slice (at i in Fig. 1). Later, more brittle northwest-directed movement is indicated by minor shears, quartz and carbonate-filled tension fractures, and curvilinear axes of late crenulation folds. Transcurrent, layer-parallel motion is indicated within the gneisses of the tectonically lowest western slice, commonly related to north-south lineations (localities a, b, c and d in Fig. 1), but also associated with northwest-southeast lineations (localities c and f).

Figure 2. Features of the Grand Lake metaconglomerate

a) Cobbles of granite (G), sheared quartzite (Qu) and unsheared quartzite (Q) in a mylonitic granitic matrix. Goose Hill.

b) Clast of sheared granite (centre) in a coarse, relatively little deformed granitic matrix with large quartz "studs" (Q) (Goose Hill granite). In the clast an earlier mylonitic foliation (left to right) has been modified by later minor shears marked by (top left to bottom right). Goose Hill.

c) Clast of sheared granite (centre left) in a sheared migmatitic matrix. Shear fabric within the clast (granulated, recrystallized quartz "plates" trending top right to bottom left) is discordant to the younger mylonitic foliation of the matrix. Goose Hill.

d) Clast of mylonitized gneiss (MG) with sigmoidal foliation in blastomylonite matrix derived from migmatitic psammite. Near the clast margins, old mylonitic foliation (top right to bottom left) has been partly transposed into left to right foliation of the matrix. Apparent sense of movement in matrix is right lateral. Smaller clasts of granite (G) lie below the mylonitic clast. Two tabular clasts of quartzite mylonite (QM) may be separated fragments of one clast. Grand Lake.

e) Metaconglomerate with pebbles(?) of feldspar (F) and quartz (Q). The matrix shows an early stage of being pulled apart into fragments bounded by a second generation of quartz, here associated with magnetite. Grand Lake.

f) Sheared metaconglomerate. Under coin an originally tabular clast of quartzofeldspathic gneiss (X) has been pulled apart into three fragments that have slipped on shear bands and back-rotated into the form of "asymmetric" boudins. The matrix is of sheared granite-soaked psammite. Deformed white pebble (F) is of feldspar, and of granite at (G). Both right lateral and left-lateral layer-parallel movements appear to be indicated within the composite fabric.

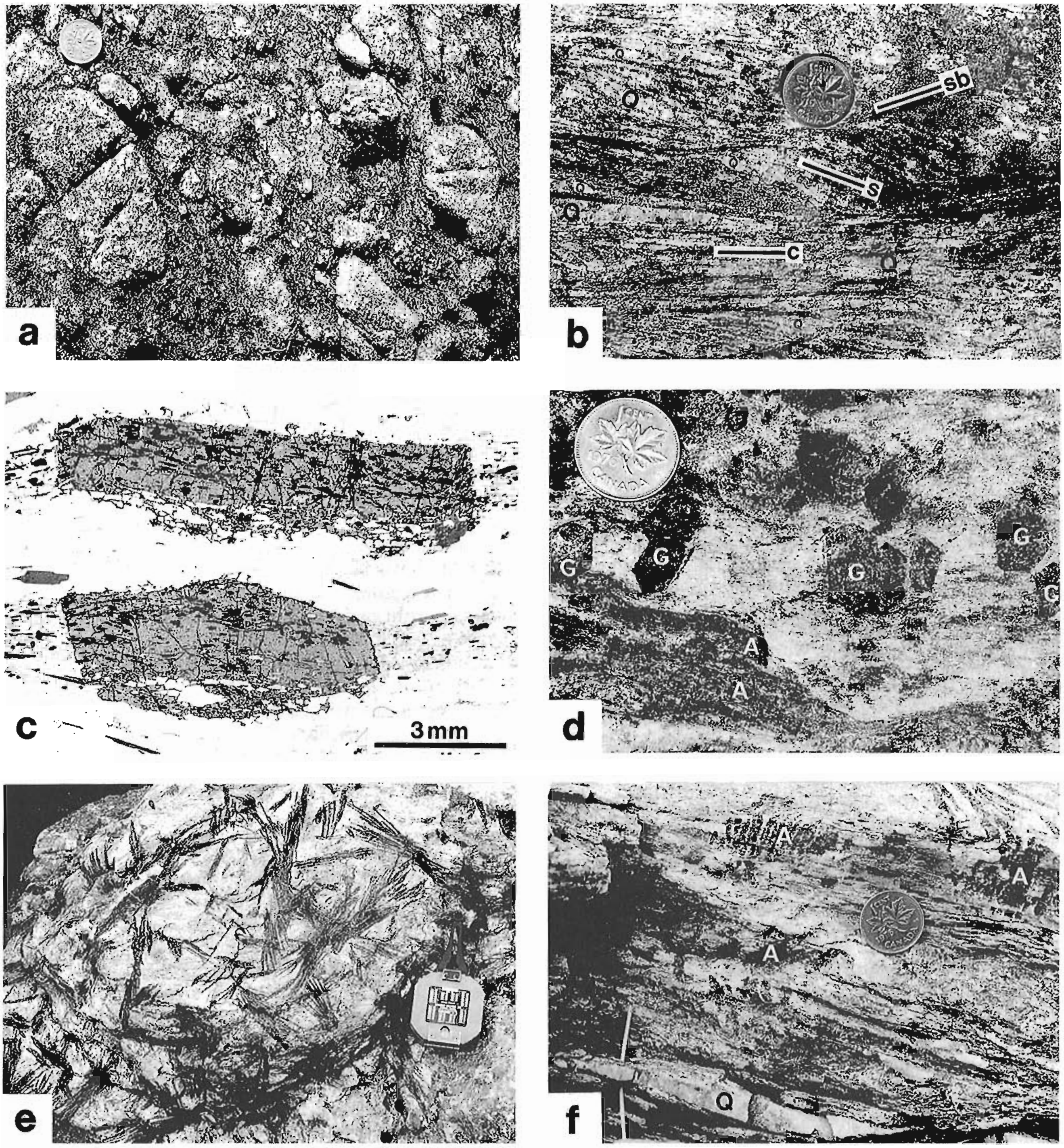
METAMORPHISM

The gneiss complex, including the dyke suite, has been pervasively retrogressed to upper greenschist or low amphibolite facies. Soda-rich plagioclase is charged with minute prisms and needles of epidote, in some cases amounting to 15 per cent by volume. Ferro-magnesian layers contain epidote, commonly zoned with Fe-rich cores, in amounts approaching that of biotite and amphibole. Garnets show partial retrogression to biotite internally charged with sphene, and sphene forms rims on opaques. These rocks characteristically contain large intergrowths of biotite, sphene, epidote, opaques and plagioclase in various proportions, locally mantled by coronas which appear to define outlines of earlier, higher-grade minerals. Regional retrogression preceded or was syntectonic with early phases of ductile thrusting, as demonstrated by the moderate straining and granulation of soda-rich plagioclase in mylonitic zones.

The cover sequence experienced syn-shearing metamorphism of at least greenschist grade, as indicated by the presence of deformed biotite and muscovite porphyroblasts. In some mylonites tabular garnets grew both along the C-fabric and normal to it, indicating growth controlled by hydraulic fracturing related to the waning shearing stress field (Temperly, 1990). Locally such garnets have back-rotated on shear bands confirming their late tectonic character. Late tectonic to post-shearing metamorphism of sheared pelites resulted in spectacular garnets reaching 2 cm in diameter (Fig. 3d), commonly accompanied by staurolite and kyanite. These porphyroblasts have been affected by late crenulation, indicating late tectonic growth. Calcareous schists recrystallized into striking "garbenschiefer" with large garnets and rosettes of blue-green amphibole growing across the mylonitic fabric (Fig. 3e and f).

DISCUSSION

The age of the gneiss complex can be constrained between the 617 Ma age of the Hare Hill granite (van Berkel and Currie, 1988) and the 1498 +/ − 8 Ma age of the Disappointment Hill granulite complex (Currie et al., in press). In lithology, fabric and dyke complex the gneisses resemble the Long Range Inlier (Owen and Erdmer, 1989), and clearly form part of the "Grenville" basement to the Humber zone. The Grand Lake section differs in its pale grey to pink colouration, which may reflect retrogression from higher (granulite-facies?) metamorphic grade. Retrogression of the gneisses and deformation and metamorphism of the cover sequence are attributed to northwest-directed ductile thrusting, associated either with mid-Ordovician emplacement of the Humber Arm allochthon (Williams, 1985; Cawood and Williams, 1988) or younger (Silurian?) events (Currie, 1987; Currie and Piasecki, 1989). This event reworked the basement, transposing older northwest to northerly trends in the gneisses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The thrust sheets observed in the Steel Mountain sub-zone must have been transported to the northwest from a region southeast of the Long Range fault, now occupied by Notre Dame and Dashwoods (Central Gneiss) sub-zones of the Dunnage zone (Piasecki et al., 1990). This suggests the presence of "Grenville-type" basement beneath the exotic Dunnage zone, as proposed on seismic evidence by Keen et al. (1986), and implies major telescoping of the upper crust during the extended ductile to brittle evolution of the Long Range fault zone.

The metaconglomerate is interpreted to be a locally developed basal member of the cover succession. It must have been laterally discontinuous, for it does not appear in the structurally higher, more easterly sheets. Following migmatization, the unit was involved in thrusting. Rheological differences between competent metaconglomerate and ductile marble would have localised greatest shearing strain in the overlying marble horizon, and hence in the location of the thick mylonite zone above the metaconglomerate. Alternately the metaconglmerate could form part of the basement sequence. This model is not favoured because the metaconglomerate does not exhibit the structural-metamorphic complexity of the gneisses. The metaconglomerate appears to have been migmatized during emplacement of granite at Goose Hill. This emplacement therefore constrains the age of deposition of the cover sequence and the subsequent ductile shearing. The Goose Hill granite is currently assumed to be into northeasterly "Appalachian" trends. Metamorphism, reaching kyanite-staurolite grade, accompanied and outlasted deformation. Metaconglomerate in the cover sequence shows that extension followed the ductile thrusting, but it is not clear whether one protracted event, or two, successive events were involved.

Figure 3. a) Goose Hill granite pulled apart into separated fragments surrounded by granulated granite. Probably originally a sheet of granite in the conglomerate. Goose Hill.

b) S-C foliation and low-angle shear bands (sb) in mylonitic quartzite of the cover sequence. Note swarm of synshearing quartz plates (Q) in the composite mylonitic C foliation. Sense of transport is left-lateral. Grand Lake.

c) Tabular garnets overgrowing mylonitic foliation in semipelitic mylonite of the cover assemblage. Grand Lake

d) Euhedral garnet (G) and amphibole (A) overgrowing mylonitic foliation in semipelitic mylonite. Cover assemblage. Grand Lake.

e) Mylonitic calc-schist recrystallized to "garbenschiefer" with rosettes of amphibole (+/− garnet) in a matrix of muscovite, biotite quartz and carbonate. Grand Lake.

f) Amphibole (A) overgrowing mylonitic foliation in calc-schist. A quartz plate (Q) in the foliation has slipped on a shear band and back rotated, indicating left-lateral slip. Grand Lake.

Precambrian on the basis of correlation with the nearby Hare Hill granite, but this correlation has not been demonstrated isotopically.

ACKNOWLEDGMENTS

I am indebted to K.L. Currie and J.T. van Berkel for their encouragement and support of this project and many discussions of the geology, and to D. Machin and D. Fox for their assistance in the field.

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A note on the stratigraphy and significance of the Martinon Formation, Saint John, New Brunswick

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Currie, K.L., A note on the stratigraphy and significance of the Martinon Formation, Saint John, New Brunswick in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 9-13, 1991.

Abstract

Large new road cuts reveal that the Martinon Formation, which consists almost entirely of turbiditic siltstone, contains huge (tens of metres) blocks of marble from the underlying Green Head Group, all of which lie at about the same stratigraphic horizon. Each block shed a train of calcareous breccia conglomerate toward the southwest. These data, together with previous observations of intrusion of the Martinon Formation by late Precambrian plutons, show that the Martinon Formation is of Precambrian age, but younger than the Green Head Group. Siltstone similar to the Martinon Formation occurs farther northwest in a volcano-sedimentary sequence probably correlative to the Late Precambrian Coldbrook Group. These observations support the hypothesis that the Saint John region forms a single terrane dissected by faults, rather than a collage of tectonic fragments.

Résumé

De grandes tranchées creusées pour la construction de nouvelles routes révèlent que la Formation de Martinon, qui se compose presque entièrement de siltstone turbiditique, contient d'énormes blocs de marbre (ayant des dizaines de mètres)du Groupe de Green Head sous-jacent, qui se trouvent tous à peu près sur le même horizon stratigraphique. À partir de chaque bloc se répand, vers le sud-ouest, une traînée de conglomérat calcareux bréchique. Ces données, jointes aux observations précédentes relatives à une intrusion dans la Formation de Martinon de plutons du Précambrien tardif, montrent que la Formation de Martinon date du Précambrien, mais est plus récente que le Groupe de Green Head. Il existe plus loin au nord-ouest dans une séquence volcanosédimentaire, probablement corrélative du Groupe de Coldbrook du Précambrien tardif, un siltstone semblable à celui de la Formation de Martinon. Ces observations confirment l'hypothèse selon laquelle la région de St-John formerait un seul terrane disséqué par des failles et ne serait pas un collage de fragments tectoniques.

INTRODUCTION

Although the stratigraphy of the Saint John region of southern New Brunswick has been studied in detail for more than 100 years, many unresolved problems remain due to structural complexity and recurrence of similar lithologies in different stratigraphic units. One such problem concerns the Martinon Formation, long thought to form a single area of homogeneous black siltstone in the western outskirts of Saint John (Fig. 1). Alcock (1938) included these rocks as part of the Precambrian Green Head Group, which typically consists of marble and quartzite, but locally contains beds of black pelitic siltstone up to 30 cm thick. Leavitt and Hamilton (1962) recognized the anomalous character of this great mass of siltstone relative to the rest of the Green Head Group, and proposed the name Martinon Formation, but left the assignment to the Green Head Group unchanged. Wardle (1977), with the aid of an almost continuous road-cut section across the formation, showed that it overlies the typical Green Head Group, and forms the core of a gently west-plunging syncline, intruded at its western end by granitoid plutons later dated by Poole (1980) at about

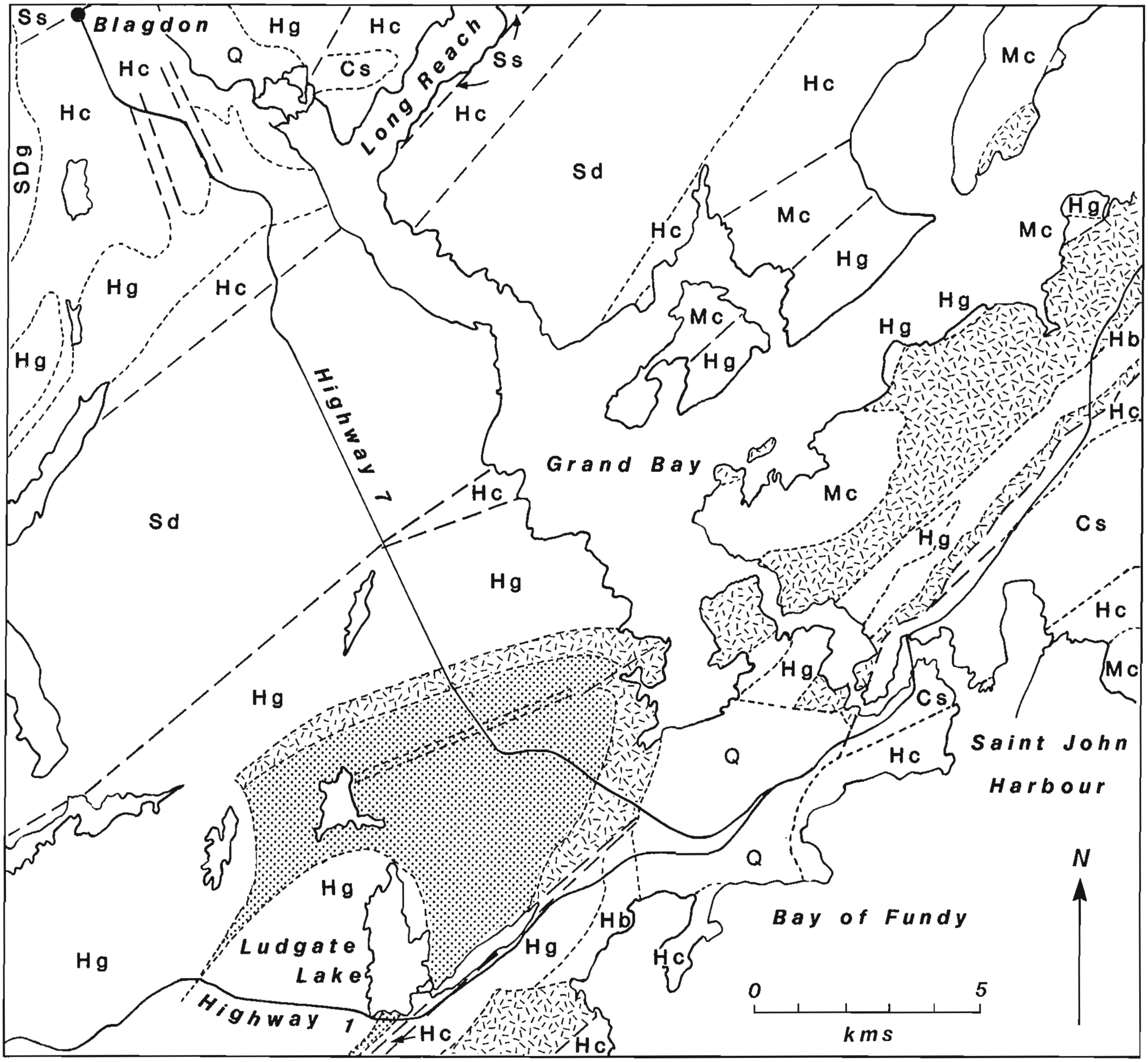

Figure 1. Simplified geological map of the Saint John region showing the location of the Martinon Formation (stippled) and the Green Head Group (hachured). The chain of giant marble blocks in the Martinon Formation is marked by dashed lines. Other units are lettered: H_b Late Precambrian orthogneiss H_g Late Precambrian granitoid plutons; H_c Late Precambrian volcanic and sedimentary strata; C_s Cambrian sedimentary strata; S_s Silurian sedimentary and volcanic strata; S_d Silurian dyke complex, SD_g Devonian granite (Mt. Douglas pluton); M_c Carboniferous conglomerate; Q surficial deposits

550 Ma. Wardle (1977) found Bouma cycles within the Martinon Formation, and concluded that it formed a slope facies deposit correlative to the shelf facies rocks of the Green Head Group. He interpreted spectacular marble breccia-conglomerate in the western part of the formation (Fig. 2) as proximal turbidite deposits, and the rest of the formation as more distal deposits. This model represented a great advance in understanding the Martinon Formation, but did not explain the sharp contact between Martinon Formation and Green Head Group on the south and east sides, and an apparently very rapid transition between shallow and deep water facies.

Figure 2. Giant block of marble (white) in the Martinon Formation, Highway 7. The truck gives a scale for the block.

Figure 3. Detail of the edge of the marble block shown in Figure 1. Note the number of smaller fragments. Height of field about 8 m.

Currie (1988) pointed out that linkage of the Martinon Formation to the Green Head Group was a historical relic with no obvious geological basis except the lack of a better niche for the Martinon Formation, and that the Martinon Formation resembled siltstone-dominated sequences fringing the Mount Douglas pluton to the northwest. These latter sequences, which consistently contain a significant volcanogenic component, were previously thought to be Silurian in age, but detailed mapping demonstrated that some of them were intruded by plutons dated by the U-Pb method on zircon at about 550 Ma (Currie, 1987).

NEW DATA

Improvements of New Brunswick Highway 7 (Fig. 1) have opened up very large new roadcuts through the Martinon Formation revealing a number of new facts about the formation. On the northwest side of the formation, a road cut intersected a very large (15×20 m), angular inclusion of Green Head marble (Fig. 2). The inclusion is visibly broken up around the edges (Fig. 3), and appears to be the source of well known marble breccia-conglomerate (Fig.4) exposed in an older road cut to the southwest. The inclusion is well within the siltstones of the Martinon Formation, and has no connection with autochthonous Green Head Group to the west. A discontinuous chain of similar inclusions, each with its cloud of breccia-conglomerate on the southwest side, extends east of the highway for at least 1.5 km. Such inclusions do not occur on the southeast side of the formation (completely exposed by road cut). The road cuts also exhibit a somewhat greater lithological variation in the Martinon Formation than previously known, ranging from pale-coloured calcareous quartzite through fine grained feldspathic wackes to finely laminated black chloritic siltstones. The basaltic material noted by Currie (1987) is well represented in the cuts. A reddened, fragmented zone of siltstone near the eastern end of the cuts marks the base of a basaltic flow, while 300 m to the northwest basaltic dykes intersect earlier sills in another new road cut (Fig. 5).

CORRELATION

Fine grained, dark siltstone generally lithologically similar to the Martinon Formation occurs in the Saint John region in (a) the Silurian Jones Creek Formation north of the Long Reach, (b) the Cambro-Ordovician Saint John Group (unit Cs of Fig. 1), (c) undated, probably late Precambrian, strata southwest of the Mount Douglas pluton (unit Hc), (d) the ?Middle Proterozoic Green Head Group (hachured). Correlation with the Jones Creek Formation or Saint John Group is excluded by the minimum age of the Martinon Formation, fixed as late Precambrian by the age of plutons intruding it. The Green Head Group consists of marble, a well-sorted, relatively coarse quartzite, and minor pelitic siltstone. None of these lithologies resemble the lithic wackes of the Martinon. The Green Head Group was sufficiently lithified to shed enormous blocks into the Martinon during sedimentation.

Figure 4. Marble breccia-conglomerate. Road cut through the Martinon Formation on Highway 7 about 15 m south of Figures 1 and 2. Note the metasomatic rinds on some fragments. The diameter of the lens cap is approximately 5 cm.

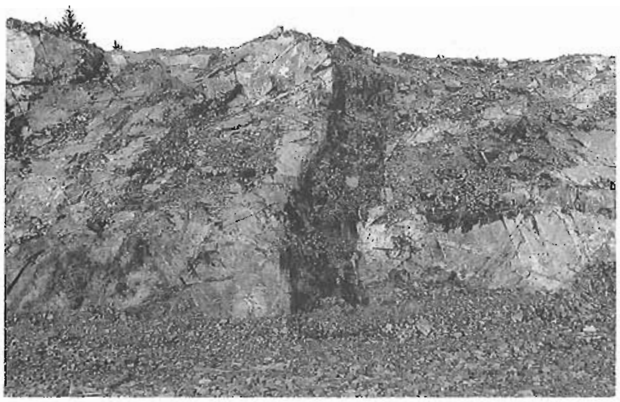

Figure 5. Younger mafic dyke intersecting Martinon Formation siltstone and a horizontal sill approximately coeval with siltstone. Truck inspection station on Highway 7. The height of the face is about 5 m.

These relations show that the marble and quartzite of the Green Head Group and the lithic wackes of the Martinon Formation belong to quite different sedimentation regimes, and suggest a significant unconformity between them, although this unconformity, if present, is now masked by deformation, and age relations between the two units are unknown.

The Coldbrook Group, formed within a continent-floored volcanic arc (Currie and Eby, 1990), consists of acid and mafic volcanic rocks, mainly fragmental, and distinctive thinly laminated, black and white cherty siltstones. Identical siltstones occur in a roadcut 1 km south of Blagdon on Highway 7 associated with acid and mafic tuffs, black siltstone, and pale grey muscovite quartzite. The latter unit is of some interest as one of the few possible sources for detrital muscovite in the lower Cambrian Ratcliffe Brook Formation (Tanoli and Pickerill, 1990). This sequence can be traced southwest on new logging roads into the black siltstone-dominated sequences fringing the Mount Douglas pluton. I conclude on the basis of general lithological and age similarity, and the presence of the distinctive laminated cherty rocks, that these strata are correlative to the Coldbrook Group. However parts of this sequence strongly resemble the Martinon Formation. This resemblance is considerably enhanced by presence of volcanics within the Martinon Formation. A wacke-basalt sequence such as the Martinon Formation falls more naturally into an active volcanic arc setting such as the Coldbrook Group than into a stable shelf setting such as the Green Head Group. Presence of huge blocks of Green Head marble within the Martinon Formation could be ascribed to collapse of lithified Green Head Group into a rifted basin.

Correlation of the Martinon Formation with lithologically similar rocks to the northwest, and in general terms with the Coldbrook Group, would appear reasonable. However Barr and White (1989) proposed a "Brookville terrane" of which the Green Head Group and Martinon Formation form an integral part. According to this proposal the Coldbrook Group formed part of a Caledonia terrane in late Precambrian time, and hence, by definition, could not be correlative with strata in the Brookville terrane. On this view, the Martinon Formation would have no correlatives in the Saint John region. Barr and White (personal communication) consider the belt north of the Long Reach, which contains fossiliferous Saint John Group, to be a displaced fragment of the Caledonia terrane, possibly imported from a considerable distance. Hence there would be no objection to correlating the strata south of Blagdon with the Coldbrook Group in general terms. However a demonstrated correlation of these strata with the Martinon Formation would be fatal to the concept of a Brookville terrane, and support the ideas of Currie et al. (1981) according to which the Brookville "terrane" exposes deeper levels of a single Avalonian terrane. The correct interpretation of the Martinon Formation is critical to resolving this controversy. The volcanic rocks of the Martinon Formation offer potential for both radiometric age dating and chemical fingerprinting for comparison with the Coldbrook Group. These possibilities are being pursued.

ACKNOWLEDGMENTS

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The Carmanville Mélange and Dunnage-Gander relationships in northeast Newfoundland[1]

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Williams, H., Piasecki, M.A.J., and Johnston, D., The Carmanville Mélange and Dunnage-Gander relationships in northeast Newfoundland; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 15-23, 1991.

Abstract

Fieldwork in 1990 is part of a continuing study of mélanges and Dunnage-Gander relationships in Newfoundland. The Carmanville area (2E/8) was also investigated in anticipation of a 1:50 000 mapping project.

The Carmanville Mélange occurs at the periphery of two discrete slices of volcanic rocks. It contains outsize sedimentary, volcanic and plutonic olistoliths in a black shaly matrix. The mélange has both olistostromal and tectonic features. Structural and metamorphic contrasts indicate recycling of deeply buried blocks and matrix. Relationships to Dunnage-Gander interaction are unknown.

The Dunnage-Gander boundary in northeast Newfoundland is tectonic with mainly sinistral transcurrent movements. Structural contrasts between the Davidsville Group (Dunnage) and the Gander Group (Gander) imply a time stratigraphic gap. A sedimentary transition at the top of the Gander Group from psammites to thin-laminated semipelites is not a Gander-Davidsville transition. Relationships support a model of Ordovician ophiolite obduction onto a continental margin.

Résumé

Les travaux de terrain effectués en 1990 s'inscrivent dans le cadre d'une étude suivie des mélanges et des relations entre les zones de Dunnage et Gander, à Terre-Neuve. On a aussi étudié la région de Carmanville (2E/8), pour la préparation d'un projet de cartographie à l'échelle de 1/50 000.

Le mélange de Carmanville se trouve à la périphérie de deux nappes distinctes de roches volcaniques. Il contient des olistolites sédimentaires, volcaniques et plutoniques de grande taille à l'intérieur d'une matrice de shale noir. Ce mélange présente à la fois des caractéristiques propres aux olistostromes et à la tectonique. Les contrastes structuraux et métamorphiques indiquent un recyclage des blocs et de la matrice profondément enfouis. On ne sait quel rapport pourrait exister avec l'interaction des zones de Dunnage et de Gander.

Dans le nord-est de Terre-Neuve, la limite entre les zones de Dunnage et de Gander est de nature tectonique, et principalement caractérisée par des décrochements senestres. Le contraste structural entre le Groupe de Davidsville (Dunnage) et le Groupe de Gander (Gander) implique l'existence d'une lacune chronostratigraphique. La transition sédimentaire marquée par le passage des psammites à des semi-pélites finement laminées que l'on observe au sommet du Groupe de Gander, ne correspond pas à une transition entre les zones de Gander et de Davidsville. Les relations établies confirment un modèle d'obduction d'ophiolites sur une marge continentale à l'Ordivicien.

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INTRODUCTION

Mélanges occur in the central part of the Carmanville area (Fig. 1, 2), within the Dunnage Zone (Exploits Subzone of Williams et al. 1988). The chaotic rocks are black to dark grey pebbly to bouldery mudstones and homogenized shales with outsize blocks of sandstone, pillow lava, massive to fragmental mafic volcanic rocks, gabbro, trondhjemite, and rare limestone. Sedimentary rocks of the Lower to Middle Ordovician Davidsville Group separate the mélanges from the Dunnage-Gander boundary that is more than a kilometre to the southeast.

The presence of the mélanges in the Carmanville area, their age, possible relationships to the local Dunnage-Gander boundary, and comparisons with Dunnage-Gander interaction elsewhere are all reasons for revisiting and revitalizing the Carmanville mélanges.

Geological field studies in central and southern Newfoundland (Colman-Sadd and Swinden, 1984; Williams et al. 1988; Piasecki, 1988; Williams et al. 1989; Piasecki et al. 1990; Williams and Piasecki, 1990), and the results of Lithoprobe experiments (Keen et al. 1986; Marillier et al. 1989) support the idea of an allochthonous Dunnage Zone above the Gander Zone. Reports of stratigraphic continuity between rocks of the Dunnage Zone and Gander Zone in the Carmanville area (Currie and Pajari, 1977; Pajari et al. 1979; Currie et al. 1979; 1980a, b) therefore take on new significance and require further study. Studies of other parts of the Dunnage-Gander boundary between Gander Lake and the Carmanville area are in progress by our colleagues P.F. Williams and Laurel Goodwin.

CARMANVILLE MÉLANGE

Definition and Nomenclature

The most obvious and spectacular mélanges in the Carmanville area occur mainly at the periphery of two large discrete areas of mafic volcanic rocks between Gander Bay and Aspen Cove (Fig. 2). We suggest the

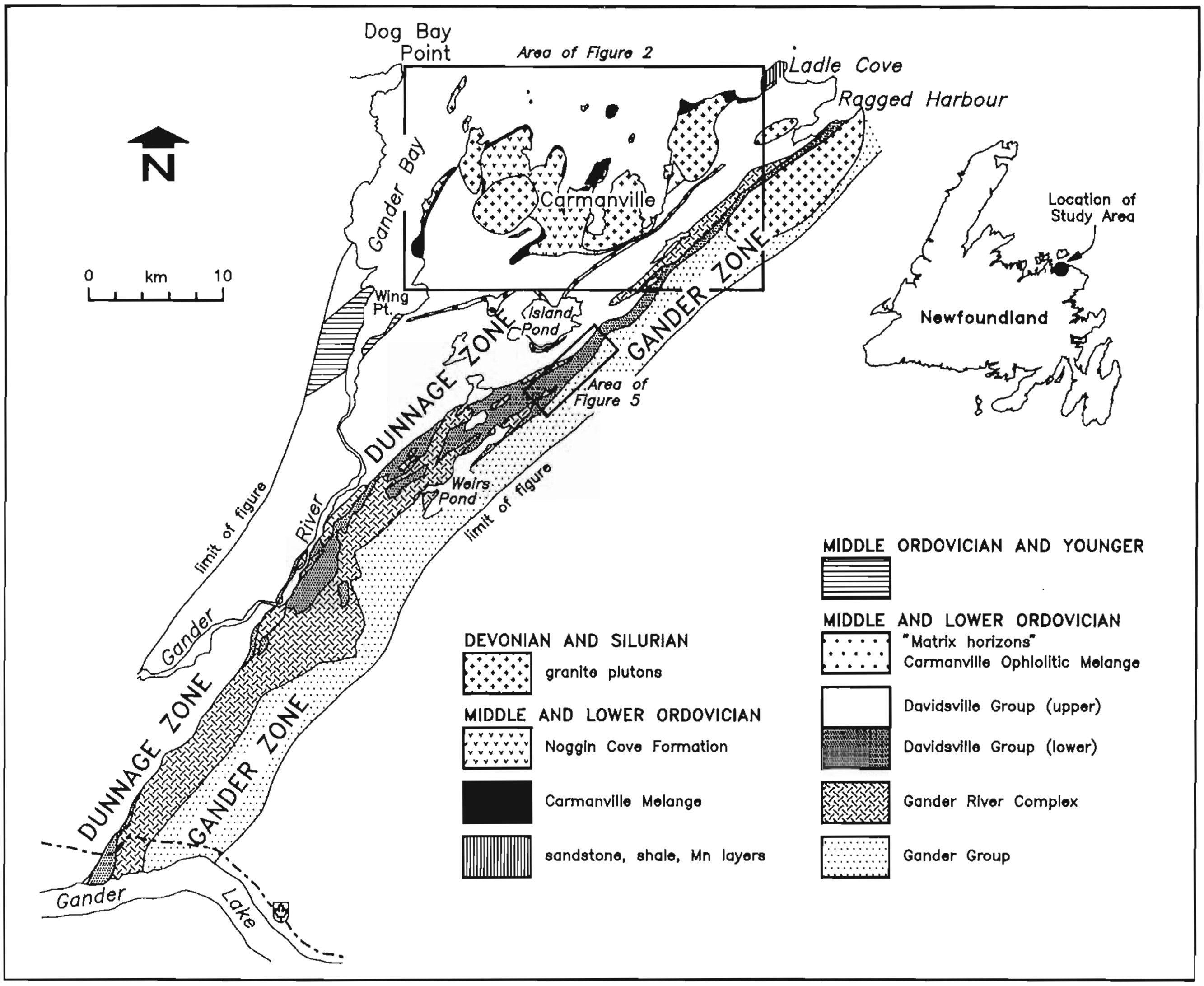

Figure 1. General geology along the Dunnage-Gander zone boundary from Gander Lake to the northeast coast of Newfoundland.

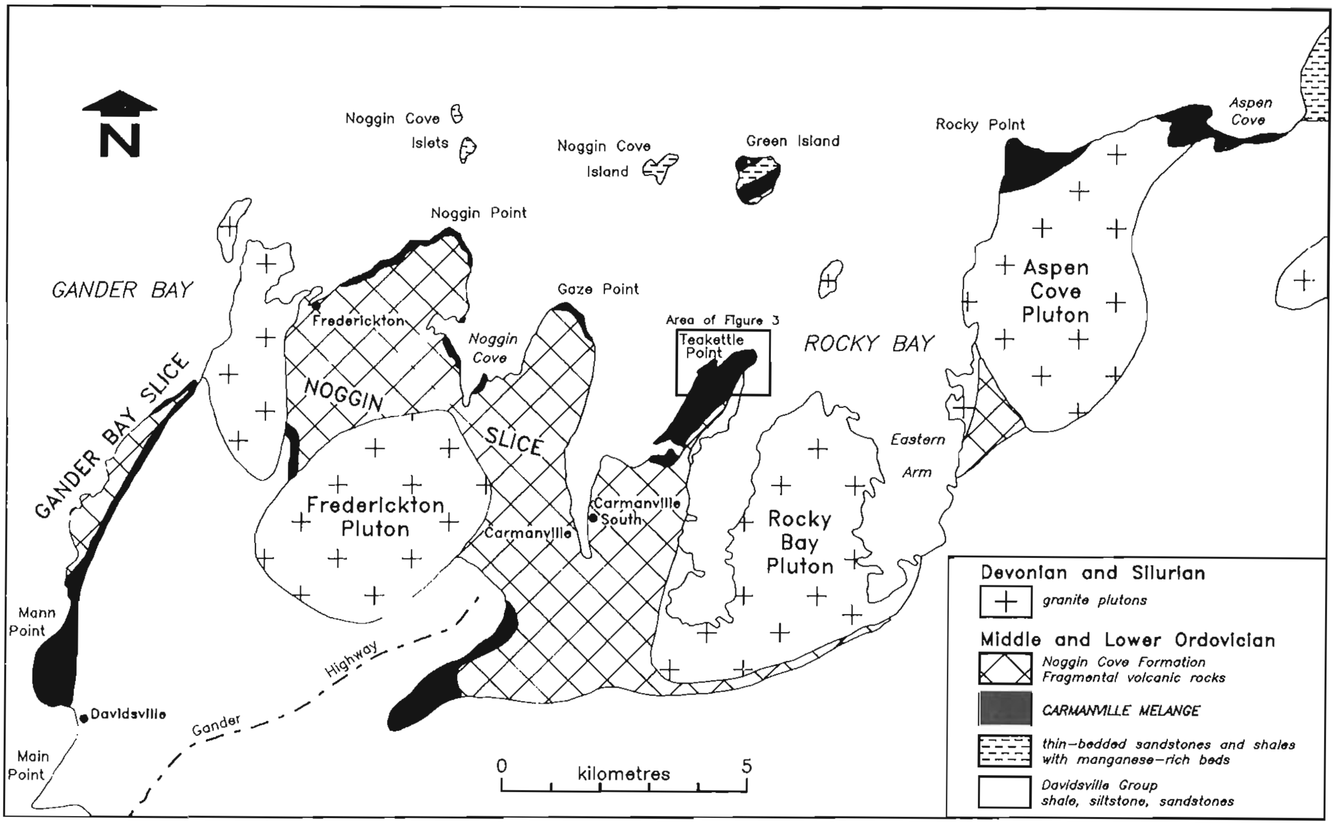

Figure 2. Distribution of the Carmanville Mélange and relationships in the Carmanville area (see Fig. 1 for location).

name Carmanville Mélange for these occurrences. A former name, Carmanville Ophiolitic Mélange (Pajari et al. 1979), was used to designate these mélanges but it included the two large discrete areas of mafic volcanic rocks and broad upper parts of the Davidsville Group. Boundaries of the Carmanville Ophiolitic Mélange were defined by "the geographical extent of matrix horizons" (Pajari et al. 1979). Defined in this manner, the Carmanville Ophiolitic Mélange was a major composite allochthon, made up of intact structural slices separated by narrow zones of mélange matrix. The two large discrete occurrences of volcanic rocks (Gander Bay and Noggin slices of Fig. 2) are omitted from the Carmanville Mélange. The volcanic rocks are a separate entity of formational status and assigned to the Noggin Cove Formation. Intact parts of the Davidsville Group are also omitted. If "matrix horizons" within the Davidsville Group are of formational importance (Fig. 1, Currie et al. 1980b), additional names are recommended.

A conceptual link between the Carmanville Mélange and the Gander River Complex (Pajari et al. 1979) is ignored, because the Carmanville Mélange is a local feature whereas the Gander River Complex (O'Neill and Blackwood, 1989) extends an additional 150 km southwestward. Furthermore, while mafic pillow lava and gabbro blocks of the Carmanville Mélange may be of ophiolitic affinity, ultramafic blocks are rare or absent. For these reasons the name Carmanville Ophiolitic Mélange is dropped.

Relationships to Nearby Groups

The Carmanville Mélange is in tectonic contact with the Noggin Cove Formation, locally with mafic volcanic blocks of the formation present in the mélange. Several lines of reasoning suggest that the Noggin Cove Formation occurs structurally above the mélange as follow: (1) the dimensions and physical continuity of the Noggin Cove Formation imply a structural slice(s) rather than an equidimensional block(s), (2) mélange exposures at Gander Bay and especially between Frederickton and Carmanville suggest that much of the present shoreline is controlled by erosion of the Carmanville Mélange at its boundary with the more resistive volcanic rocks that are topographically, and therefore structurally above the mélange, (3) indications of piercement of the volcanic rocks (above) by mélange matrix (below), (4) the volcanic rocks of the Noggin Cove Formation are distinctive and everywhere the same, dominated by fragmental varieties with minor pillow lavas, (5) there are no stratigraphic links or lithic correlatives of the Noggin Cove Formation in the Davidsville Group, and (6) the direction of stratigraphic tops in the Noggin Cove Formation are incompatible with those of surrounding sedimentary rocks (Pickerill et al. 1978). Separate occurrences of the Noggin Cove Formation may be remnants of a single slice, folded and partly omitted by erosion and intrusion.

Basal relationships of the Carmanville Mélange and the nature of its contacts with sedimentary rocks of the

Davidsville Group to the southeast are unknown and await the results of studies in progress. Other sedimentary rocks north of the mélange at Noggin Cove Islets, Noggin Cove Island, Green Island (Fig. 2) and Ladle Cove (Fig. 1) are thin-bedded shales and sandstones with thin manganese-rich layers, unlike the Davidsville Group to the south. They interact with the Carmanville Mélange at Green Island where olistostromes alternate with bedded sections, and at Aspen Cove (Fig. 2), where equivalents of manganese-rich beds form coticule in metamorphosed mélange. We hesitate to include these sedimentary rocks, the Carmanville Mélange, and the Noggin Cove Formation in the Davidsville Group.

The Carmanville Mélange and its surrounding rocks are affected by a regional steep northeast-trending cleavage. Between Gander Bay and Aspen Cove, the folded rocks are truncated by the Frederickton, Rocky Bay and Aspen Cove plutons (Fig. 2). These are mainly trondhjemitic plutons that contrast with potassic plutons of the Gander Zone (K.L. Currie pers. comm. 1990). Regional metamorphism affects the mélange and surrounding rocks east of Rocky Point.

Description of Carmanville Mélange

Contrasts in structural and metamorphic styles are the most obvious and problematic feature of the Carmanville Mélange. These contrasts are best observed at Teakettle Point (Fig. 3), and are also present at Gaze Point and westward. They occur on a scale of several hundred metres to single deformed blocks in less deformed matrix. Where deformation and metamorphism are mild, the mélange consists of outsize, rounded to lenticular blocks of sandstone, pillow lava, fragmental volcanic rocks, gabbro, trondhjemite, and limestone in a matrix of homogenized black shale or pebbly to cobbly black shale. Where deformation and metamorphism are intense, the mélange consists of banded mafic schist, greenschist, attenuated pillow lava, psammitic schist, and a black pelitic to semipelitic matrix. Areas of intensely deformed and metamorphosed mélange are juxtaposed or surrounded by less deformed and metamorphosed mélange. Contacts are sharp. Areas of mildly deformed and metamorphosed mélange account for more than 80 per cent of exposure with blocks and matrix in roughly equal proportions. Deformed and metamorphosed mélange has less matrix, about 10 per cent, and in places blocks are juxtaposed without intervening matrix.

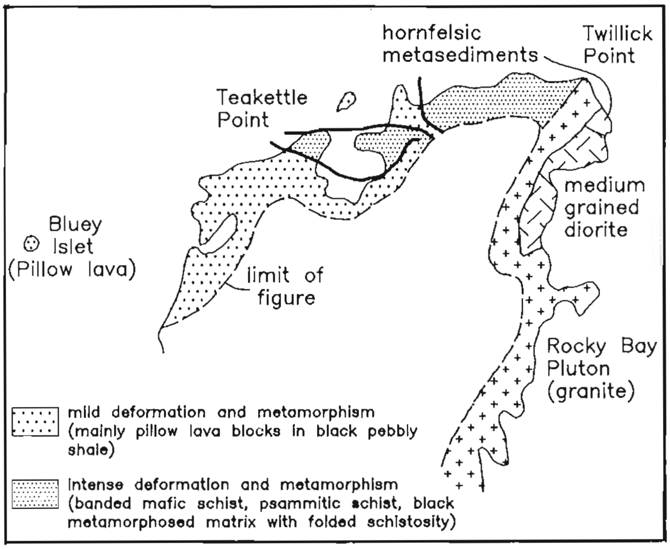

Figure 3. Local distribution of intensely deformed and metamorphosed mélange and mildly deformed and metamorphosed mélange at Teakettle Point (see Fig. 2 for location).

In the mildly deformed and metamorphosed mélanges, most small sandstone clasts and some larger sandstone and volcanic clasts are aligned parallel to the regional cleavage, and some clasts are elongate and plunge steeply northeast. Other large resistive volcanic blocks are equidimensional and one example with a limestone interbed at Carmanville South has internal bedding perpendicular to regional cleavage. The direction of the main fabric and banding in deformed pillow lavas and psammites at Teakettle Point is east-west or northwest-southeast, at high angles to the later northeast regional cleavage. Commonly, the dark pelitic matrix of deformed mélange is strewn with broken quartz veinlets.

Some components of the intensely deformed and metamorphosed mélange are the same as those in the less deformed and metamorphosed varieties, although proportions are different. Attenuated pillow lavas and banded mafic schists of the deformed mélange have associated marble and therefore resemble pillow lavas with associated limestone in the less deformed mélange. Psammitic schists in the deformed mélange have no counterparts in undeformed examples. They resemble the thin-bedded sandstones and shales that alternate with olistostrome units at Green Island (see also Pajari and Currie, 1978; Pajari et al. 1979). Thinly-laminated, black pelitic and semipelitic matrix may be equivalent to thinly-laminated black shale and sandstone at Noggin Cove Island.

Age

Early Ordovician (Arenigian) trace fossils from sedimentary rocks at Green Island indicate an early Ordovician age for olistostromes interlayered with the sedimentary rocks (Pajari and Currie, 1978: Pickerill et al. 1978; Pajari et al. 1979). Pillow lavas and mafic volcanic breccias with associated limestone, like those in the mélange, are common in the Exploits Subzone of the Dunnage Zone, especially at New World Island. There, the limestones contain a Llandeilo fauna (Fåhraeus and Hunter, 1981). A presumed Caradocian or younger age for the mélange was an outcome of including the upper Davidsville Group in the Carmanville Ophiolitic Mélange (Pajari et al. 1979).

Interpretation

The Carmanville Mélange has been interpreted as an olistostrome or major surficial slump (Kennedy and McGonigal, 1972; Pajari et al. 1979). Pre-entrainment deformation in some components was first recognized by Kennedy and McGonigal (1972). Their idea that the deformed components were derived from the Gander Group was always doubtful because of the abundance of deformed pillow lavas in the mélange and their absence in the Gander Group, and because the concept of Ganderian Orogeny (Kennedy, 1975) was not supported by later workers (Currie and Pajari, 1977; Pajari and Currie, 1978; Pickerill et al. 1978; Blackwood, 1978; 1982; Currie et al. 1979; Pajari et al. 1979). Soft rock deformation was suggested for the structural complexities in psammites at Teakettle Point (Pajari and Currie, 1978) and dynamothermal metamorphism, related to ophiolite obduction, was also mentioned (Pajari et al. 1979).

The Carmanville Mélange combines features of olistostromes and tectonic mélanges. The presence of interbedded coherent sedimentary rocks and mélange at Green Island indicates sedimentary processes. Pebbly mudstone matrices and local bedded sections elsewhere also imply surficial processes of slump and mass wastage. In contrast, pervasively deformed and metamorphosed mélanges affected by extensive cataclasism, such as those at Teakettle Point, indicate deeper dynamothermal conditions followed by brittle deformation. The alternation of intensely deformed and metamorphosed mélange with mildly deformed and metamorphosed mélange, intricate outcrop patterns between the two, and the presence of discrete intensely deformed and metamorphosed blocks in a pebbly mudstone matrix all indicate that an earlier generation of mélange was deeply buried, deformed, metamorphosed, affected by cataclasism, and then exhumed or recycled where juxtaposed or embedded in pebbly mudstone. Any model for the Carmanville Mélange must therefore involve dynamic processes whereby surficial olistostromes are subjected to dynamothermal conditions and then returned to the surface.

A spatial association of the Carmanville Mélange and overlying volcanic rocks of the Noggin Cove Formation suggests that emplacement of a major volcanic structural slice may have controlled mélange formation.

The Noggin Cove Formation consists mainly of resedimented fragmental volcanic rocks with unsorted vesicular clasts (Pickerill et al. 1981). Associated pillow lavas, which are not vesicular, and local black shale interbeds suggest deposition in a deep marine basin. A disproportionate number of pillow lava and gabbro blocks in the Carmanville Mélange compared to fragmental volcanic rocks of the Noggin Cove Formation suggests that the Noggin Cove Formation is the remaining upper part of a thicker transported sequence that included pillow lavas and gabbros at its base. It may have been a volcanic edifice above an ophiolite suite. Age of the transported rocks, time of transport, direction of transport, and tectonic controls are all first order concerns.

Mélanges occur all the way across the northeast Exploits Subzone from Carmanville to New World Island (Fig. 4). The oldest is the Dunnage Mélange (Hibbard and Williams, 1979). The Carmanville Mélange may be coeval or slightly younger. Another mélange of late Ordovician or early Silurian age occurs at Dog Bay Point (Pajari et al. 1979; Currie et al. 1980b), and the Silurian Joeys Cove Mélange (Reusch, 1987) occurs farther northwest. Incorporating these mélanges into a prolonged, dynamic model of the Exploits Subzone is a future challenge.

DUNNAGE-GANDER RELATIONSHIPS

Descriptions and conclusions that follow are based on detailed mapping along widely-separated segments of the Dunnage-Gander boundary between Gander Lake and Ragged Harbour (Fig. 1). More systematic descriptions of the area between Gander Lake and Wiers Pond are given by Blackwood (1982) and O'Neill (1987).

The Gander Group consists of impure metasandstones and metasiltstones, quartzites, pelites, and minor amphibolite. Towards its western boundary, silty and pelitic bands are abundant with the sequence passing into grey pelite ribbed with thin bands of semipsammite. Metamorphic facies is mid greenschist, but at Ragged Harbour it rises to amphibolite in metasediments injected by sheets of syntectonic granite.

The most distinctive characteristic of semipsammites along the western flank of the Gander Group, both at Gander Lake and in the Carmanville area, is a layer-parallel pressure solution cleavage, of which there are two generations. The first (S_1), is axial planar to outcrop-size, subrecumbent, north-trending isoclinal folds (F_1). This cleavage, marked by discontinuous wisps of mica and chlorite has a spacing of 5 to 10 mm in massive, coarse psammites, but it is more closely-spaced in metasiltstones. The S_1 cleavage is folded by numerous, small, northeast-trending, isoclinal second folds (F_2). The small, asymmetric F_2 folds transpose the first cleavage into a more thinly-spaced pressure solution cleavage which dips to the northwest.

Third generation folds are open, near upright crenulation folds, coaxial with F_2 folds. They control the dip of the S_2 cleavage. They were followed by locally developed folds and by brittle kinkbands.

The Gander River Complex (O'Neill and Blackwood, 1989) lies at or near the Davidsville-Gander boundary from Gander Lake to the northeast coast of Newfoundland. It consists of serpentinite, pyroxenite, gabbro, trondhjemite and mafic volcanic rocks. The complex is generally viewed as ophiolitic (Kennedy, 1975; 1976; Pajari et al. 1979) but there is no complete sequence preserved. Rather, the distribution of its rock types appears to be that of random structural sheets. An internal brecciation in its rocks is characteristic. This is especially well displayed in the trondhjemite body exposed in a quarry on the Gander Bay Highway.

In the Carmanville area, the Gander River Complex is represented by a large body of sheared, altered and recrystallized serpentinite and pyroxenite, and by several small, separated bodies of serpentinite (Fig. 1, 5).

The Davidsville Group in the Carmanville area comprises dark grey shales with thin bands of grey metasiltstone and metasandstone, and variegated purple shales

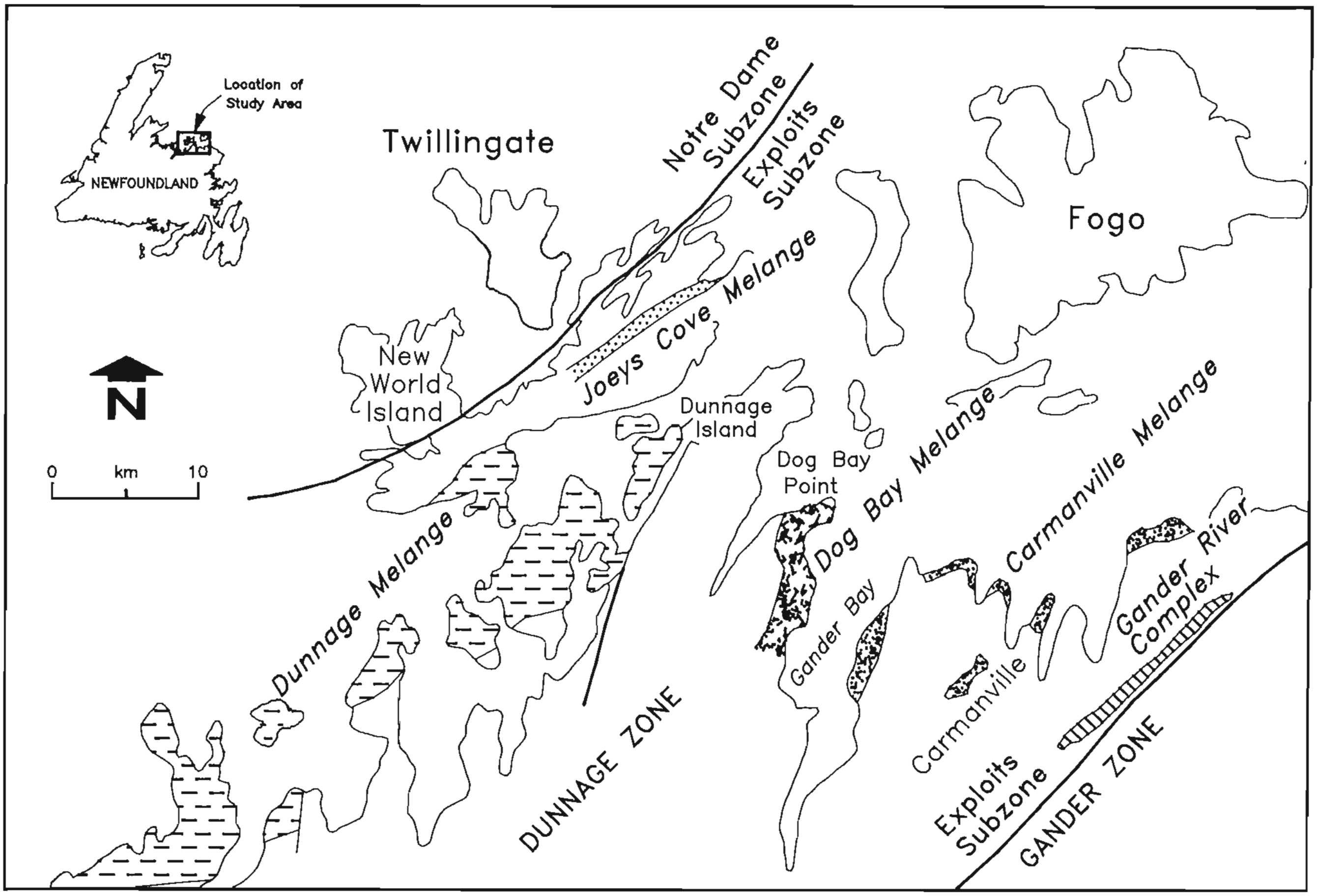

Figure 4. Distribution of mélanges across the Exploits Subzone of the Dunnage Zone in northeast Newfoundland.

with green metasiltstones. The rocks have a pressure solution cleavage and they are metamorphosed to incipient greenschist facies. A Llanvirnian-Llandeilian conodont fauna like that at Wiers Pond (Stouge, 1980) occurs in limestone at Cuff Pond (Fig. 5, K.L. Currie, pers. comm., 1990). Rocks of the Davidsville Group are affected by only one generation of cleavage. This cleavage (S_1) is axial planar to outcrop-size, upright to inclined, gently-plunging folds of bedding which trend northeast. The bedding and intersecting cleavage are modified by at least two subsequent phases of crenulation and kink folding.

Gander-Davidsville Boundary

At Gander Lake (Fig. 1), the northwestern flank of the Gander Group is generally marked by a zone of layer-parallel, ductile shearing separating it from altered ultramafic rocks of the adjacent Gander River Complex. Where the Gander River Complex is absent in the Carmanville area, a zone of shearing some 2 km wide marks the Gander-Davidsville boundary (Fig. 5). In this shear zone, blastomylonite, mylonite and locally ultramylonite anastomose between elliptical "low strain augen" of protomylonites.

Small, asymmetric, isoclinal F_2 folds, which fold the S_1 cleavage and bedding, are common in psammitic and semipsammitic lithologies of the Gander Group toward its northwest margin. These apparently syn-shearing folds with axial trend parallel to a regional stretching lineation, transpose the original wide-spaced, discontinuous S_1 cleavage into a northwest-dipping S_2 pressure solution cleavage. The S_2 cleavage is more intense toward the shear zone and it is progressively closer-spaced and more continuous; producing a paper-thin mylonite C-foliation (Berthe et al. 1979). Where the S_2 cleavage is intense, layers of metasandstone and metasiltstone form boudins within sheared slaty shale (phyllonite). A Gander protolith is indicated, even in intensely mylonitic zones, by small separated semipsammitic lenticles, which represent rootless F_2 hinges that fold an earlier cleavage.

Davidsville rocks at Cuff Pond exhibit megascopic, open F_1 folds of bedding that are more isoclinal toward the shear zone, and their axial cleavage (S_1) is more intense and merges with the main C-foliation of the mylonite belt. Beds of metasiltstone are attenuated but rarely necked into boudins.

Locally within this mylonite belt, the Gander-Davidsville boundary is contained within some 50 m of intensely mylonitic dark grey phyllonites in which neither protolith can be recognized.

The Gander River Complex adjacent to the Gander Group is also commonly mylonitic. On the north shore of Gander Lake and in exposures along the Trans Canada Highway, a narrow belt of green mylonite-ultramylonite with carbonate lenticles is in contact with Gander Group phyllonites. The southeastern flank of the large ultramafic body in the Carmanville area (Figure 1) is marked by a zone of gabbro mylonite and ultramylonite with small augen-shaped relicts of gabbro and fine grained diabase, and local lenses of carbonate. It is separated by an exposure gap of 200 m from Gander Group mylonites. A similar mafic mylonite occurs southeast of Ragged Harbour.

Throughout the shear zone at the Davidsville-Gander boundary in the Carmanville area, S-C fabrics and shear bands in syntectonic quartz veins and mylonitic rocks derived from both the Gander and Davidsville groups indicate layer-parallel transcurrent sense of movements which are both left-lateral and right-lateral (Fig. 5). Of these, the left-lateral fabrics are more common and more penetrative. Furthermore, in a body of mylonitic serpentinite 300 m north of Cuff Pond, a right-lateral shear band cleavage overprints the main mylonitic foliation. This indicates that the dominant fabrics of left-lateral movements are earlier, and are related to the main phase of ductile shearing. In Ragged Harbour, a syntectonic granite records the transcurrent movements and a later phase of ductile thrusting with northwest over southeast, down dip transport.

East of Ragged Harbour (Fig. 1), coarse, semipelitic biotite-sillimanite mylonitic schists with swarms of concordant quartz veins, and psammites of the Gander Group are permeated and migmatized by a white-weathering biotite granite (Ragged Harbour Pluton of Currie et al. 1980b). In numerous places the migmatized metasediments show mobilization with a left-lateral sense of movement, indicating that the granite was emplaced at the time of the regional, left-lateral shearing movements.

Discussion

The tectonic contact between the Gander Group and the Gander River Complex at Gander Lake is a ductile shear zone developed in the thinly-banded pelitic Gander Group and greenschist derived from the Gander River Complex. The mylonitic foliation in this tectonic zone is the second generation fabric in the Gander Group. Nearby Davidsville conglomerates, unconformable on the Gander River Complex, have a single fabric. Where the Gander River Complex is missing in the Carmanville area, the mylonitic fabric at the Gander-Davidsville boundary is the second generation fabric in the Gander Group but the first generation fabric in the Davidsville Group. These relationships imply a time stratigraphic as well as a tectonic break between the Gander and Davidsville groups.

At Cuff Pond (Fig. 5), isolated, mylonitic and non-mylonitic bodies of the Gander River Complex occur within the Davidsville Group in a zone extending approximately a kilometre from its tectonic boundary with the Gander Group. This suggests that the Gander River Complex is basement to the Davidsville Group and that it is structurally emplaced within its cover. The tectonic boundaries between the Davidsville Group and the Gander River Complex are therefore less significant than those between the Davidsville Group and the Gander Group, or between the Gander River Complex and the Gander Group (see also O'Neill, 1987).

The kinematic sequence of orogen-parallel, transcurrent left-lateral movements followed by right-lateral translation recorded from the Gander-Carmanville area matches the kinematics at Dunnage-Gander boundaries

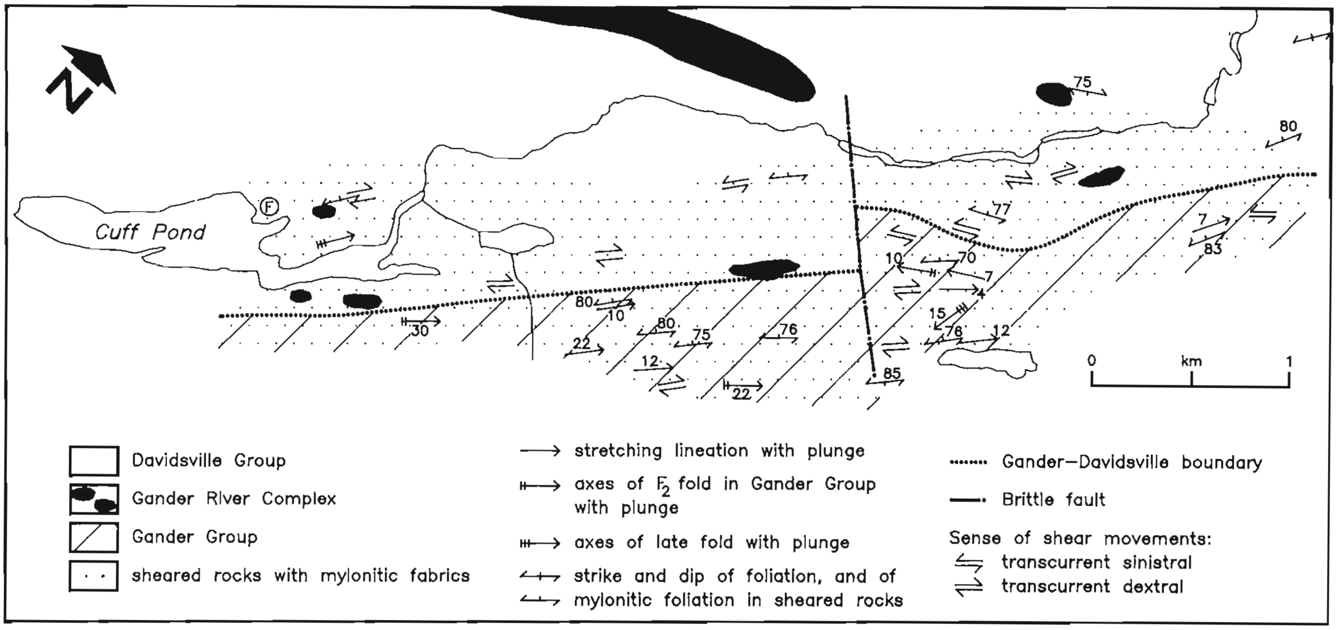

Figure 5. Relationships between the Davidsville Group (Dunnage Zone) and Gander Group (Gander Zone) in the Carmanville area at Cuff Pond.

at Bay d'Espoir, Great Burnt Lake and Cold Spring Pond in central Newfoundland (Piasecki, 1988; Piasecki et al. 1990). Thus, the structural styles of an allochthonous Dunnage Zone above the Gander Zone are extended into the northeastern Gander Lake Subzone of Williams et al. (1988).

We agree with a structural and stratigraphic break between the Gander and Davidsville groups as originally proposed by Kennedy and McGonigal, 1972. We also recognize a sedimentary transition between psammitic Gander (southeast) and semipelitic Gander (northwest), which could be misinterpreted as a Gander-Davidsville transition in deformed rocks of the Carmanville area and elsewhere (Pickerill et al. 1978; Pajari et al. 1979; Currie et al. 1979, 1980a, b; Blackwood, 1978, 1982). If this transition represents a facies change controlled by subsidence, it is easily explained as a result of loading in a model of ophiolite obduction. Deformation in the Gander Group that preceded Davidsville deposition is also expected in such a model.

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Lithostratigraphy of the Upper Gaspé Limestones Group (Early Devonian) west of Murdochville, Gaspésie, Quebec

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Abstract

A stratigraphic and sedimentological study of the Upper Gaspé Limestones Group (Early Devonian) was carried out in central Gaspésie. The three formations of the group (Forillon, Shiphead and Indian Cove) are still recognized but with some major lithofacies variations compared with the previously studied eastern adjacent area. These variations are likely linked to different tectonic regimes as well as to different sedimentary and volcanic settings. The observed north-south lithofacies variation in a dominant medial to deep outer shelf setting is preliminarily interpreted as resulting from proximity with emergent volcanic centres in south-central Gaspésie. From the observed lithofacies variation, a refined lithostratigraphic framework is proposed.

Résumé

Une étude stratigraphique et sédimentologique des Calcaires supérieurs de Gaspé (Dévonien précoce) a été menée dans la région centrale de la Gaspésie. Les trois formations du groupe (Forillon, Shiphead et Indian Cove) présentes plus à l'est, sont également reconnues à cet endroit, avec cependant d'importantes variations de lithofaciès. Ces variations sont probablement reliées à des environnements géo-tectoniques et volcano-sédimentaires différents. Il s'agit d'une succession sédimentaire typique d'une plate-forme externe plus ou moins profonde, avec des variations nord-sud imprimées par la présence de centres volcaniques émergents dans le centre sud de la Gaspésie. Le schéma lithostratigraphique déjà proposé pour le groupe est amendé pour tenir compte de ces nouvelles observations.

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INTRODUCTION

The stratigraphic and sedimentological study of the Upper Gaspé Limestones Group, initiated in 1989 (Lavoie et al., 1990) was resumed during the 1990 field season. Our efforts were concentrated in the area bordered by Murdochville to the east and by Etang à la Truite to the west (Fig. 1). This area covers more than 1500 km^2 over which 15 stratigraphic sections and 1 drill core were described and sampled.

This contribution deals with the lithostratigraphy of the Upper Gaspé Limestones Group and discusses the westward extension of the lithostratigraphic framework established in the adjacent eastern area (Lespérance, 1980a, b; Lavoie et al., 1990).

GEOLOGICAL SETTING

The Praguian Upper Gaspé Limestones Group is one of the few carbonate intervals in the otherwise siliciclastic-dominated sedimentary history of the post-Taconic Gaspé Basin (Lavoie et al., 1990; Bourque et al., in prep.). The establishment of this carbonate-dominated regime coincided with an episode of major drowning that affected the post-Taconic Gaspé Basin (Bourque et al., in prep.).

The Upper Gaspé Limestones Group outcrops, in the 1990 surveyed area (Fig. 2), in east-northeast-trending open synclines and anticlines of a narrow belt less than 25 km wide. This outcropping area is bordered to the north by pre-taconic units with a discordant Silurian-Devonian cover. To the south, the group laterally passes into the Fortin Group. The exact nature of this transition is unknown, due to poor exposures and faulting.

Few prominent faults cut through the studied unit. The major fault of this area is the east-west trending Shickshock Sud Fault, that runs through the northern pre-taconic and silurian units. This fault is thought to connect with the Bras Nord-Ouest Fault (D. Brisebois, pers. comm. to D. Lavoie), the latter controlling in part the tectonostratigraphic history of the eastern segment of the basin (Lavoie and Tassé, 1990). Some northeast or northwest trending faults were recognized. They are particularly abundant and mineralized (Pb-Zn-Cu) in the Lemieux Dome sector (Fig. 2).

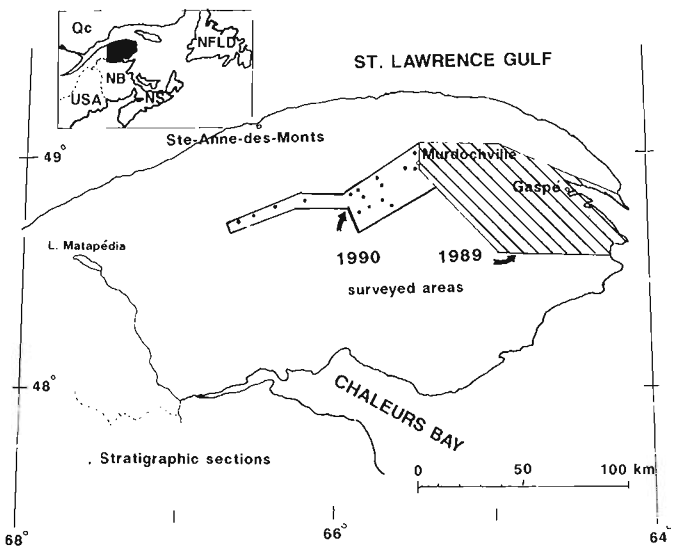

Figure 1. Location map of studied areas in Gaspésie, Québec.

PREVIOUS STUDIES

Lespérance (1980a), following subdivisions recognized by Logan (1845) and Russel (1946) proposed the formal term Upper Gaspé Limestones Group for the lower Devonian carbonate dominated succession in eastern Gaspésie. Details of the stratigraphic and nomenclature evolutions were presented by Lavoie et al. (1990). The sedimentology and paleoenvironments of deposition for this area were discussed by Mason (1971), Lespérance (1980a), Bourque et al. (in prep.), Lavoie (1990) and Lavoie and Tassé (1990).

Geological surveys in the studied area started in the early 20th century, following the discovery of Cu-Pb-Zn mineralization in Lemieux township (Alcock, 1921). Further geological mapping was carried out by Jones (1930, 1931, 1932) and by McGerrigle (1954). At that time, what is now known as the Upper Gaspé Limestones was described as the Cap Bon Ami and Grande Grève formations. More recent geological mapping (Amyot, 1983; Rhéault, 1983), drilling (Duquette et al., 1984), and a study by Lachambre and Brisebois (in prep.) have helped to elucidate the nature of the succession. Stratigraphic studies carried out by Rouillard (1986) recognized the threefold nature of the group. Much work was also done over mineralization within the Upper Gaspé Limestones Group. The Mines Gaspé at Murdochville (Fig. 2) is a major copper deposit hosted in metasomatically altered limestones of the Shiphead and Indian Cove formations (Bellehumeur et Valiquette, 1988; Procyshin and Bernard, 1990).

LITHOSTRATIGRAPHY

Various lithofacies were observed in the Upper Gaspé Limestones Group during the 1990 field season. Carbonate, siliciclastic and volcanic facies were found interlayered (Fig. 3). The threefold lithostratigraphic framework of the group, recognized in eastern Gaspésie (Lespérance, 1980a, b) and based on the recognition of the lithologically heterogeneous Shiphead Formation, still holds west of Murdochville. The nature of the lithofacies present in one unit of the group is however dependant on its actual tectonostratigraphic setting with regard to the various folds in the area, defining three geographic and stratigraphic belts of exposures. The Shiphead/ Indian Cove formations contact is easily recognizable and is used as datum line for correlative purposes.

The Northern Exposures

Eight sections were described in the northern part of the studied area (Fig. 2, 3). They are located either on the northern limb of the northernmost synclines (sections 2 to 4 and 6 to 8), north of the Bras Nord-Ouest Fault (section 1), or in the much faulted Lemieux Dome area (section 5).

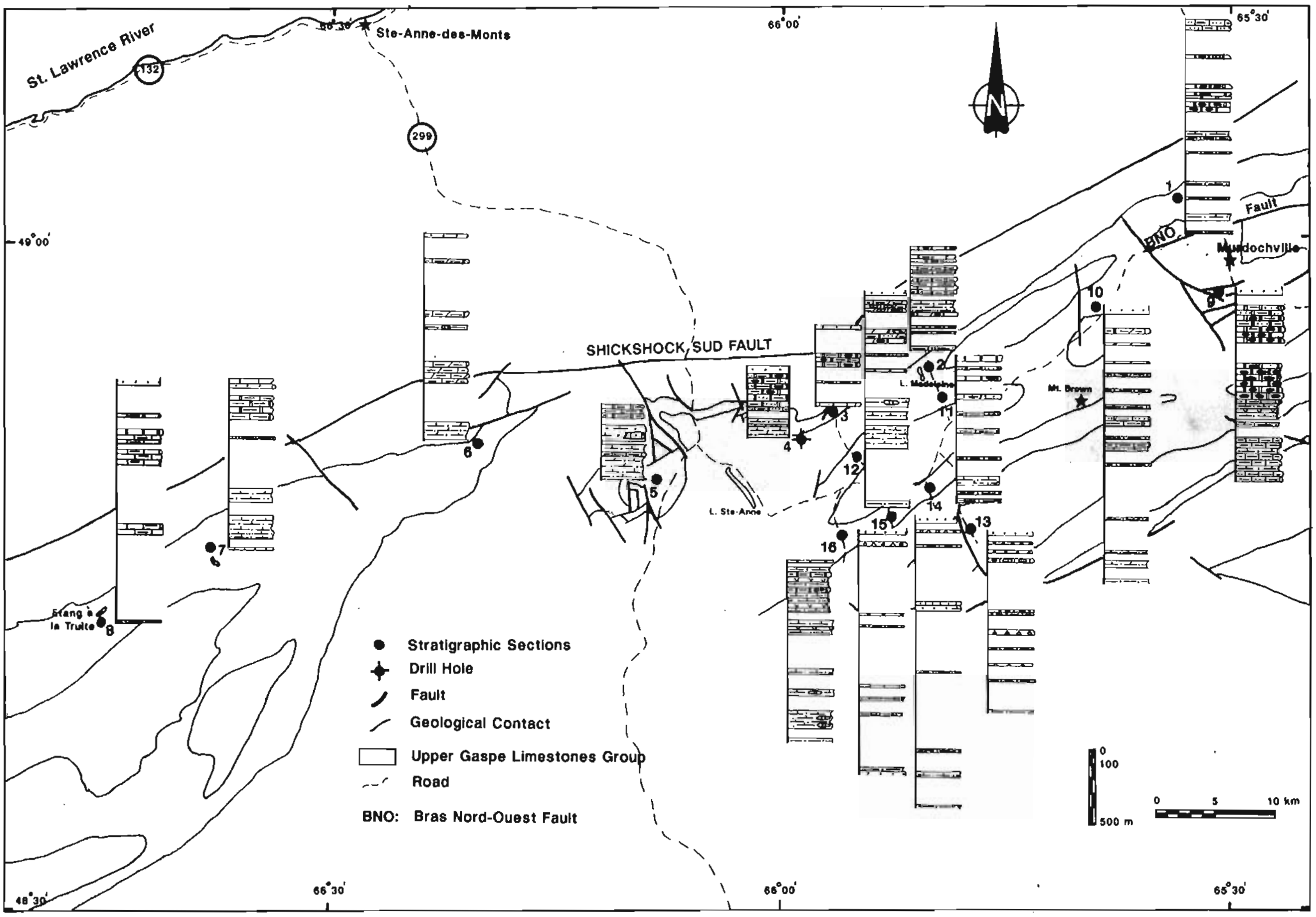

Figure 2. Simplified geological map and general stratigraphy of the studied horizon in central Gaspésie (see Fig. 3 for lithology). Sections; 1: Alcide Brook, 2: Lake Madeleine road (north), 3: Lesseps road (north), 4: Mont-Vallières-de-Saint-Réal drill hole, 5: Lemieux dome, 6: Cascapédia River area, 7: Lake Simoneau road, 8: Etang à la Truite, 9: Needle mountain, 10: Burntwood brook, 11: Lake Madeleine road (south), 12: Lesseps road (central), 13: Little east Cascapédia River, 14: Lac Long Brook, 15: Lesseps road (south), 16: Burma road.

Forillon Formation

The Forillon Formation is at the base of the group. Its thickness varies from 80 to 660 m. It conformably overlies the uppermost unit of the Chaleurs Group (Saint-Léon or Indian Point formations; Bourque, 1975). This contact was never seen exposed in the field but interlayering of Saint-Léon type (greyish siltstones and mudstones) and Forillon type lithologies occurs over a few metres at Lemieux Dome and Lake Simoneau road (sections 5 and 7, Fig. 2, 3) and suggests a gradual contact at least locally. However, for the Mont-Vallières-de-Saint-Réal (MVSR) drill core (section 4, Fig. 2, 3), this contact is sharp and set at the last, non calcareous mudstone bed assigned to the Indian Point Formation.

Lavoie et al. (1990) divided the Forillon Formation in eastern Gaspésie into two members: the mudstones of the Mont Saint-Alban Member (now the Chemin du Roy Member) and the calcilutites of the Cap Gaspé Member. This division is still present at section 1 (Alcide Brook) and a few kilometres west of it. At Lake Madeleine (section 2) poorly exposed cherty calcilutite represents the easternmost reach of the Cap Gaspé Member and the entire Forillon Formation west of Lake Madeleine, is represented by dark, more or less calcareous mudstones to shaly calcilutites with a mudstone texture. In this succession, less than 15% of diagenetic carbonate concretions (averaging 10 cm × 5 cm × 10 cm) are seen. The dark mudstones and calcilutites (Fig. 4) are in beds of a few centimetres to 35 cm in thickness with few parallel laminations and rather uncommon brachiopods. Next to these beds, few horizons, up to 30 cm thick, of allochems-rich packstone calcarenites (Fig. 5) were locally found at the Chemin Lesseps section and the MVSR drill core (sections 3 and 4, Fig. 2, 3).

Rouillard (1986) introduced the term "Lesseps facies" for the dark calcareous mudstone/shaly calcilutite succession. This succession is typical of the Forillon Formation for the 1990 study area. We propose to rank them as a formal member. The Lesseps Member type section is located in the Lesseps Brook area (section 12, Fig. 2, 9) in the central outcrop belt. The lower and upper limits of this member correspond to the limits of the Forillon Formation in this area.

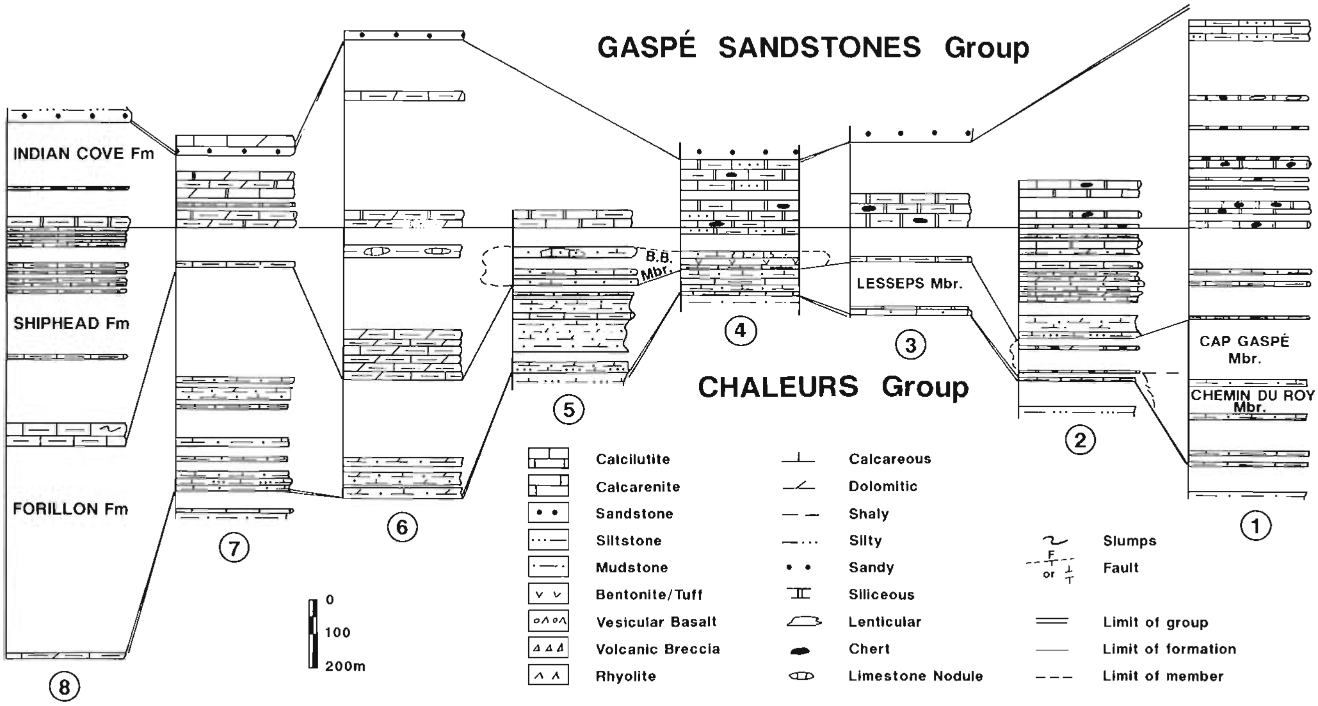

Figure 3. Lithostratigraphic correlations for the northern exposures. B.B. Mbr.: Brandy Brook Member. Location of sections in Figure 2.

Figure 4. Dark calcareous mudstones of the Lesseps Member (Forillon Formation). Section 5.

Figure 5. Scouring and faintly graded crinoid-rich packstone calcarenite with siliciclastic mudstone clasts of the Lesseps Member (Forillon Formation). Section 3.

Shiphead Formation

The Shiphead Formation reaches a thickness of 100 to 640 metres (Fig. 2, 3). It conformably overlies the Forillon Formation. The contact seen only in the MVSR drill core (section 4), is sharp and set at the top of the uppermost dark calcareous mudstone of the Lesseps Member.

The Shiphead Formation succession is made up of various limestone, siliciclastite and volcaniclastite horizons. The limestones constitute around 20 per cent of the unit but are more abundant in the western end of the studied area. They are mostly represented by shaly calcilutite with a mudstone texture, in beds ranging from 5 to 40 cm thick (average 15 cm). The calcilutite commonly shows parallel and crossbedded laminations. Silicification with local chert characterizes the uppermost calcilutites of the unit. Some beds, up to 70 cm thick, of allochems-rich packstone calcarenites are found scattered in the unit. They represent only a small fraction (2 per cent) of the carbonate lithofacies.

Siliciclastic lithofacies make up almost 80 per cent of the unit. They show a wide variety of lithologies from mudstone to sandstone, both calcareous and noncalcareous. The fine grained lithofacies dominates volumetrically. The mudstones are commonly greyish and bedding is generally masked by fissility. The sandstones are best exposed in the Lake Madeleine area (section 2) and at Lemieux Dome (section 5). They are quartz-rich and fine to medium grained. These sandstones occur in beds up to 1 m thick (average 20 cm). They offer a wide spectrum of sedimentary structures from rare scouring and grading to frequent parallel and cross-bedded laminations (Fig. 6). Silica or calcite is the most common cement of this lithofacies. Both the mudstones and sandstones locally bear abundant allochems. This charateristic is best developed at the Lemieux Dome (section 5, Fig. 2, 3), where the Shiphead succession is mostly made up of medium to thickly bedded bioclastic sandstones and subordinate bioclastic mudstones. Crinoids, brachiopods, corals and bryozoans, which total a maximum of 60% of the rock, make up the bioclastic fraction. These bioclasts occur in structured (graded and laminated) lenses of variable shapes and sizes, generally less than 60 cm thick, in the siliciclastic beds (Fig. 7). Locally these pockets are shaly to sandy packstone calcirudite. These allochems-rich beds are known as the Brandy Brook Member (Gentile, 1969; Rouillard, 1986). We suggest that that designation be retained, as it is in common use and represents an important lithological variation from the usual fine grained Shiphead Formation succession.

All the sedimentary lithofacies are commonly bioturbated, mostly by *Scalarituba* sp. with subordinate *Zoophycos* sp. trail, the latter being generally restricted to the carbonate lithofacies.

The volcaniclastic lithofacies is represented by bentonite beds that were only observed in the MVSR drill core (section 4, Fig. 2, 3). These beds have a maximum thickness of 20 cm.

Indian Cove Formation

The Indian Cove Formation, the uppermost unit of the group (Fig. 2, 3), reaches a thickness of 200 to 650 m. Its contact with the Shiphead, as seen in the MVSR drill

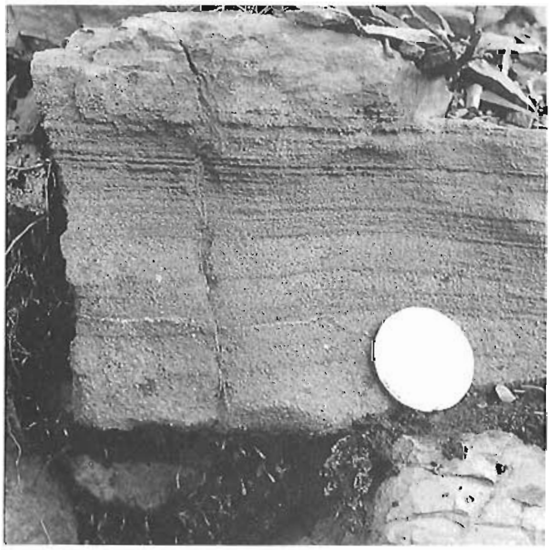

Figure 6. Structured sandstone interlayered with calcareous mudstones. Shiphead Formation. Section 2.

Figure 7. Lens of sandy packstone calcirudite in calcareous sandstone of the Brandy Brook Member (Shiphead Formation). Section 5.

core (section 4) is sharp and set at the top of the uppermost sandstone horizon assigned to the Shiphead Formation. Its contact with the overlying Gaspé Sandstones Group is put at the first non calcareous sandstone or siltstone bed assigned to either the York Lake or York River formations (Lespérance, 1980a).

The Indian Cove Formation is made up of a rather homogeneous succession of limestones. These are represented by shaly and siliceous calcilutites with a mudstone texture. The beds are slightly wavy bedded and generally less than 35 cm thick and showing parallel laminations. Chert is locally present being particularly abundant at the Alcide brook section (section 1), where it represents up to 45 per cent of the rock. Elsewhere, it is less abundant and frequently absent. Chert occurs in nodules of variable sizes and shapes and in thin irregular ribbons. The upper third of the formation is characterized by an higher siliciclastic content that is expressed by presence of silty laminations (parallel and cross-bedded) and lenses of silty calcilutites reaching a maximum of 20 cm in thickness (Fig. 8).

The limestone lithofacies is intensely bioturbated by *Scalarituba* sp. and *Zoophycos* sp. Few bioclasts were noted.

The Central Exposures

Sections 9 to 12 (Fig. 2, 9) are described from the central segment of the study area. They are either located on the southern limb of the northernmost syncline (sections 10 to 12) or south of the Bras Nord-Ouest fault (section 9). Since many field characteristics are common with the already described northern sections, only the notable differences or additions to the studied units will be described.

Forillon Formation

The Forillon Formation reaches a thickness of 280 to 850 m. The Forillon Formation is represented by the dark calcareous mudstones/shaly calcilutites of the Lesseps Member. The Lesseps Brook section (no. 12) was chosen as type section for this member.

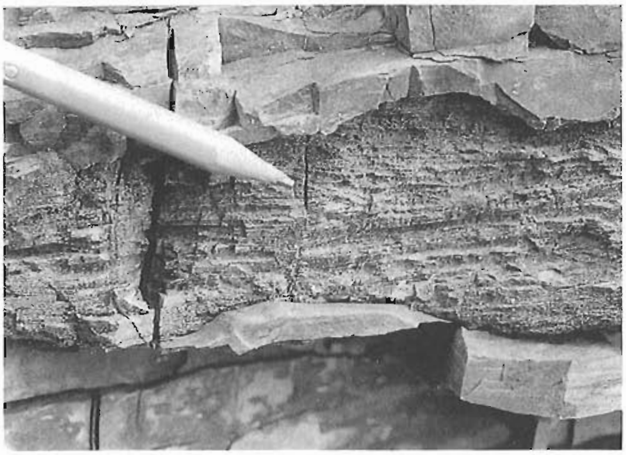

Figure 8. Structured (parallel and crossbedded laminations) lens of silty calcilutite in a partly silicified calcilutite. Indian Cove Formation. Section 3.

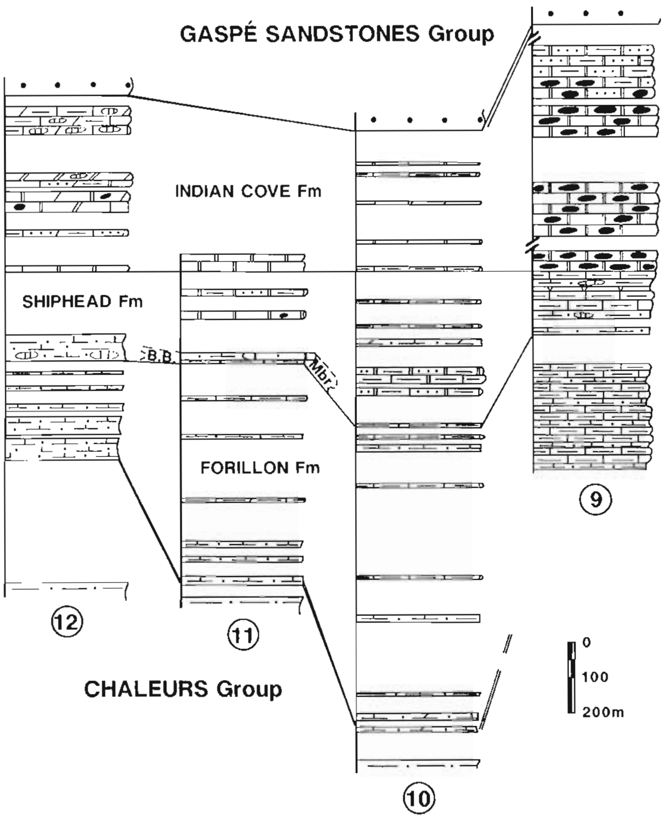

Figure 9. Lithostratigraphic correlations for the central exposures. B.B. Mbr.: Brandy Brook Member. Location of sections in Figure 2 and lithology in Figure 3.

Shiphead Formation

The Shiphead Formation has a thickness ranging from 180 to 425 m. Carbonate lithofacies are similar to the ones of the northern sections but no calcarenite beds were seen. The ratio of siliciclastic versus carbonate is still highly favourable to the former but sandstones are less abundant in that area. Mudstone, both calcareous and non calcareous, is the most characteristic lithology of the Shiphead Formation in that sector. Allochems-rich mudstones assigned to the Brandy Brook Member, are well exposed near Lake Madeleine (section 11, Fig. 2, 9). The unsorted allochems are concentrated in lenses in the mudstone and may constitute up to 90% of the rock locally (Fig. 10). Crinoids are the most abundant allochem with subordinate micrite clasts, corals, bryozoans and brachiopods. Some of the bryozoans are encrusted by algal-like material. Siliciclastic mudstone fragments were also observed. Laterally, the amount of fragments rapidly decrease to give way to a poorly fossiliferous mudstone suggesting a lenticular nature for the allochems-rich lithofacies.

Bentonite beds are present at the Needle Mountain section (section 9, Fig. 2, 9). At this same locality, beds up to 15 cm thick of tuffaceous mudstones were also seen.

Indian Cove Formation

The Indian Cove Formation thickness ranges from 400 to 700 m. In the central outcrop belt, it is almost entirely represented by shaly, siliceous calcilutites with a mudstone texture, the upper metres of the formation being more siliciclastic-rich. Chert is locally present but in small amounts with the exception of the Needle Mountain exposures (section 9, Fig. 2, 9) where it can make up to 50 per cent of the rock, occurring in centimetre-sized ribbons.

A few metres below the contact with the Gaspé Sandstones Group, a concretionnary horizon is locally present (section 12 and outcrops along the Lake Sainte-Anne road, Fig. 2). Micrite concretions (averaging 5 cm × 5 cm × 2 cm) are either highly irregular to nicely rounded in shape. They can locally make up to 50 per cent of the

Figure 10. Crinoids and micrite clasts rich mudstone of the Brandy Brook Member (Shiphead Formation). Section 11.

rock volume and are embodied in a shaly calcilutite having a mudstone to packstone texture. Bioclasts are abundant, with *in situ* well preserved brachiopods, trilobites and cephalopods in the matrix of the rock but also showing various allochtonous large fragments of corals (Fig. 11), bryozoans (Fig. 12) being occasionnally encrusted by algal-like material, subordinate stromatoporoids, and possible fragments of algal micrite (Fig. 13). The fragmented allochems, largely unsorted and forming up to 25% of the rock at section 12, are seen to locally act as nuclei for carbonate precipitation in the process of concretion formation (Fig. 12).

Mafic tuff beds ranging from 10 to 50 cm thick, are seen near Lake Sainte-Anne at the top of the Indian Cove Formation.

The Southern Exposures

Sections 13 to 16 (Fig. 2, 14) are located in the southern segment of the study area. They are either located on the north limb of the southern syncline (sections 14 to 16) or on the southern limb of that same syncline (section 13). This area is particularly poor in outcrops and thickness of the sections could be tectonically altered.

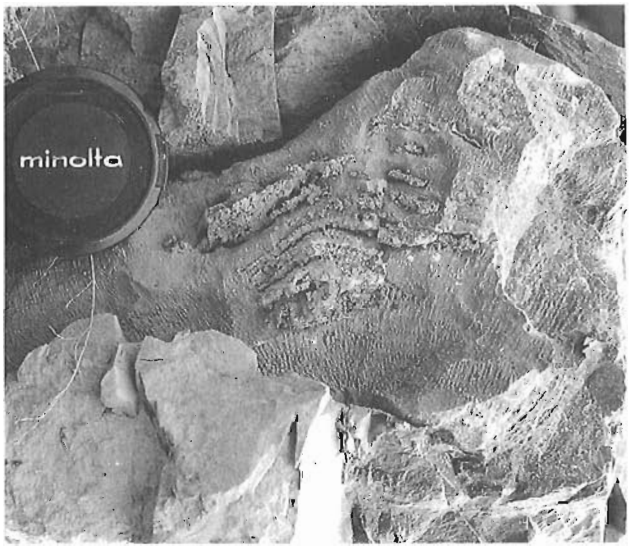

Figure 11. Large (20 cm × 10 cm × 10 cm) tabulate coral fragment set in impure calcilutite. Indian Cove Formation. Section 12.

Figure 12. Bryozoan fragment acting as nuclei for a carbonate concretion. Indian Cove Formation. Section 12.

Forillon Formation

The Forillon Formation is represented by the dark calcareous mudstones/shaly calcilutites of the Lesseps Member.

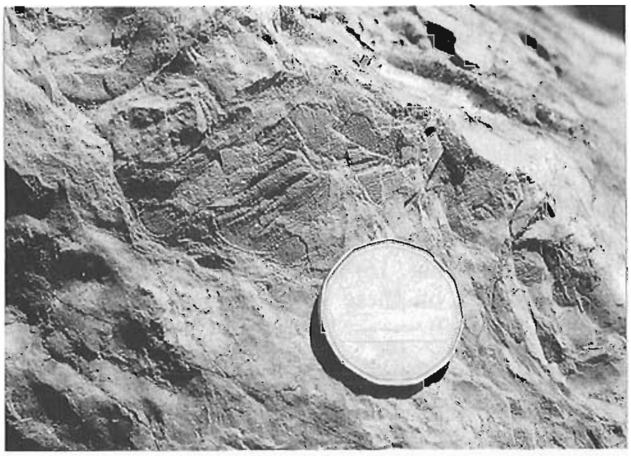

Figure 13. Angular fragment of likely algal micrite set in impure calcilutite. Indian Cove Formation. Section 12.

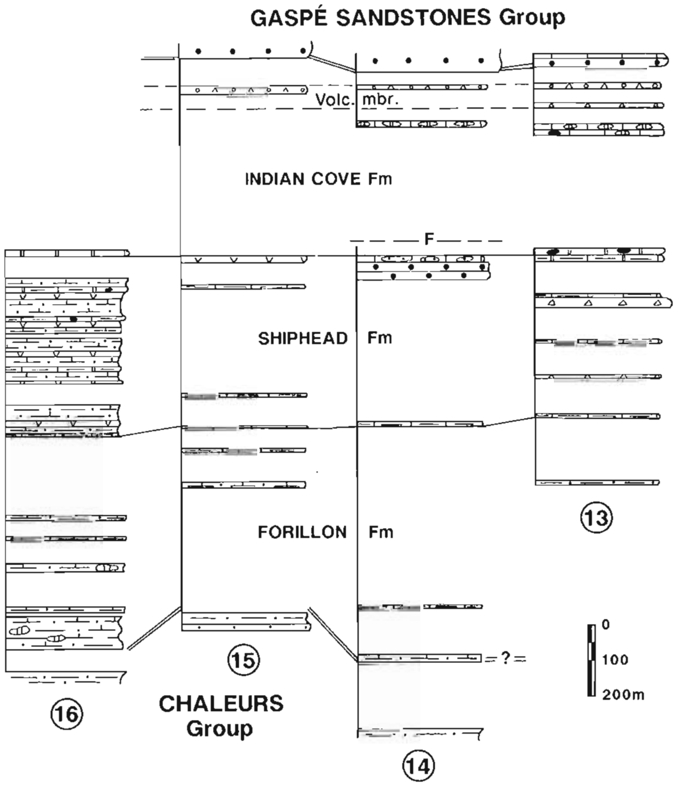

Figure 14. Lithostratigraphic correlation for the southern exposures. Volc. mbr.: Volcanic member. Location of sections in Figure 2 and lithology in Figure 3.

Shiphead Formation

The best exposure of the Shiphead Formation is at the Burma road (section 16, Fig. 2, 14) where beds of calcareous to non calcareous greyish mudstones are interlayered with calcilutites having a mudstone texture. This section also shows numerous beds, averaging 15 cm in thickness, of mafic and felsic, fine- to medium-grained tuffs with locally abundant and thick bentonite layers. In the Lac Long Brook area (section 13), a relatively thick succession (50 m) of bioturbated, quartz-rich calcareous sandstones, up to 15 cm thick, with abundant parallel and trough crossbedded laminations (Fig. 15) was seen. These are in turn overlain by calcilutites having a mudstone texture with interbeds and lenses of less than 15 cm thick of reddish calcareous sandstones. Southward (section 13), volcanic lithofacies interlayer with siliceous calcilutites and are assigned to the Shiphead Formation. Numerous thick (up to 6 m wide) rhyolite horizons and thickly bedded (up to 3 m), faintly graded, coarse felsic pyroclastic polymictic (rhyolite, porphyry, basalt, tuff, gabbro) breccias (Fig. 16) are found. Petrological and geochemical data on these volcanics can be found in Doyon et al. (1990).

Indian Cove Formation

The Indian Cove Formation is poorly exposed in this sector. Few beds of siliceous to seldom cherty calcilutite with a mudstone texture were seen. Near the top of the unit, the previously described concretion-bearing horizon was locally seen (sections 13 and 14, Fig. 2, 14) but is devoid of large metazoan clasts. This unit is in turn overlain by a felsic pyroclastic horizon bearing bioclast (corals and stromatoporoids) and micrite fragments (Fig. 17). The carbonate fragments make up around 10 per cent of the breccia. At the top of the Indian Cove Formation, immediately underlying the Gaspé Sandstones Group, few metres of vesicular basalts were seen. The basalts contain some large clasts of micrite (Fig. 18) and few bioclasts fragments (corals). The basalt vesicules, mostly plugged by calcite, are sometimes filled by micrite (Fig. 19), some with geotropic textures. These vesicular basalts were found westward (section 15, Fig. 2, 14), whereas eastward (Brown Mountain area, Fig. 2), field investigations carried by Brisebois (pers. comm. to D. Lavoie) seem to show that this horizon is discordant with the stratigraphy.

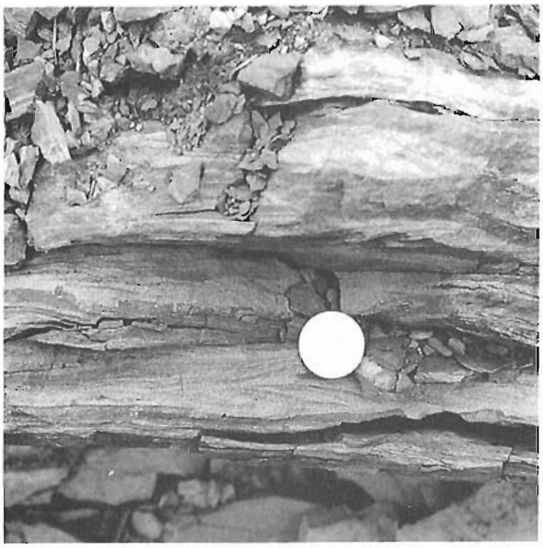

Figure 15. Well-sorted sandstone with trough cross-bedded muddy laminations. Shiphead Formation. Section 14.

Figure 16. Thickly bedded, coarse- to fine-grained, felsic pyroclastic polymictic breccia. Note the very angular shape of the fragments. Shiphead Formation. Section 13.

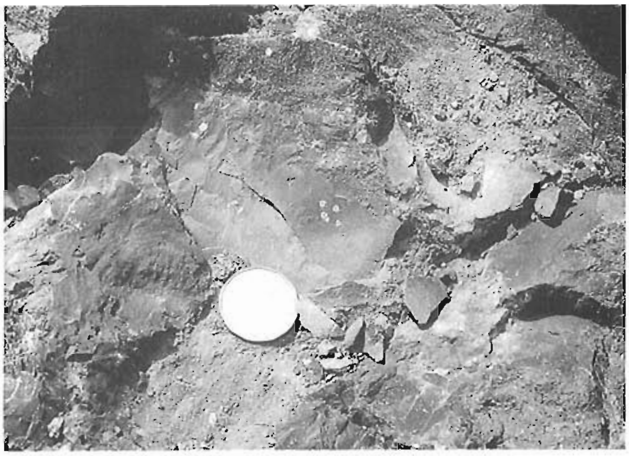

Figure 17. Angular fragment of micrite set in felsic pyroclastic breccia. Volcanic member (Indian Cove Formation). Section 13.

The summital volcanic horizon represented by tuffs (Lake Sainte-Anne road), pyroclastic breccias (section 13) and vesicular basalts (sections 13, 14, 15) is a mappable unit. The preliminary informal designation of volcanic member is thus introduced for this 70 m thick unit. This member is overlying, in the southern and central outcrop belts, the concretion-bearing horizon of the Indian Cove Formation. It is in turn overlain by beds of the Gaspé Sandstones Group.

DISCUSSION AND CONCLUSIONS

Figure 20 sketches the regional lithofacies distribution for the Upper Gaspé Limestones Group in the area covered by this report. Figure 21 summarizes the lithostratigraphy of the group for both the 1989 and 1990 surveyed sectors. From these, the following stratigraphic considerations and preliminary paleoenvironmental interpretations are drawn:

— The Forillon Formation, west and south of the Acadian strike-slip Bras Nord-Ouest (BNO) Fault, is a rather uniform succession of dark calcareous mudstones and shaly calcilutites (Lesseps Member). North and east of the Bras Nord-Ouest Fault, the Forillon Formation succession offers the bimodal lithological suite that characterized the eastern adjacent area (Lavoie et al., 1990). As suggested for the eastern part of Gaspésie (Lavoie et al., 1990; Lavoie and Tassé, 1990), the Bras Nord-Ouest Fault was also likely synsedimentary active, at least in Forillon time, in the eastern end of the area covered in 1990.

— The Shiphead Formation is an heterogeneous unit with siliciclastic and subordinate volcaniclastic and limestone facies. The Brandy Brook Member, at the base of the unit, is at the regional and outcrop scales, a lenticular body composed of bioclast-rich sandstones and mudstones. Allochems are typical of shallow water environments. They are clearly allochtonous to their likely deeper water depositionnal environment and could represent a mixture of various event beds and debris flow units. A few paleocurrent indicators

Figure 18. Large (1.5m × 75cm), irregularly shaped, micrite clast embodied in vesicular basalt. Volcanic member (Indian Cove Formation). Section 13.

Figure 19. Micrite geotropic infilling of voids in vesicular basalt. Volcanic member (Indian Cove Formation). Section 13.

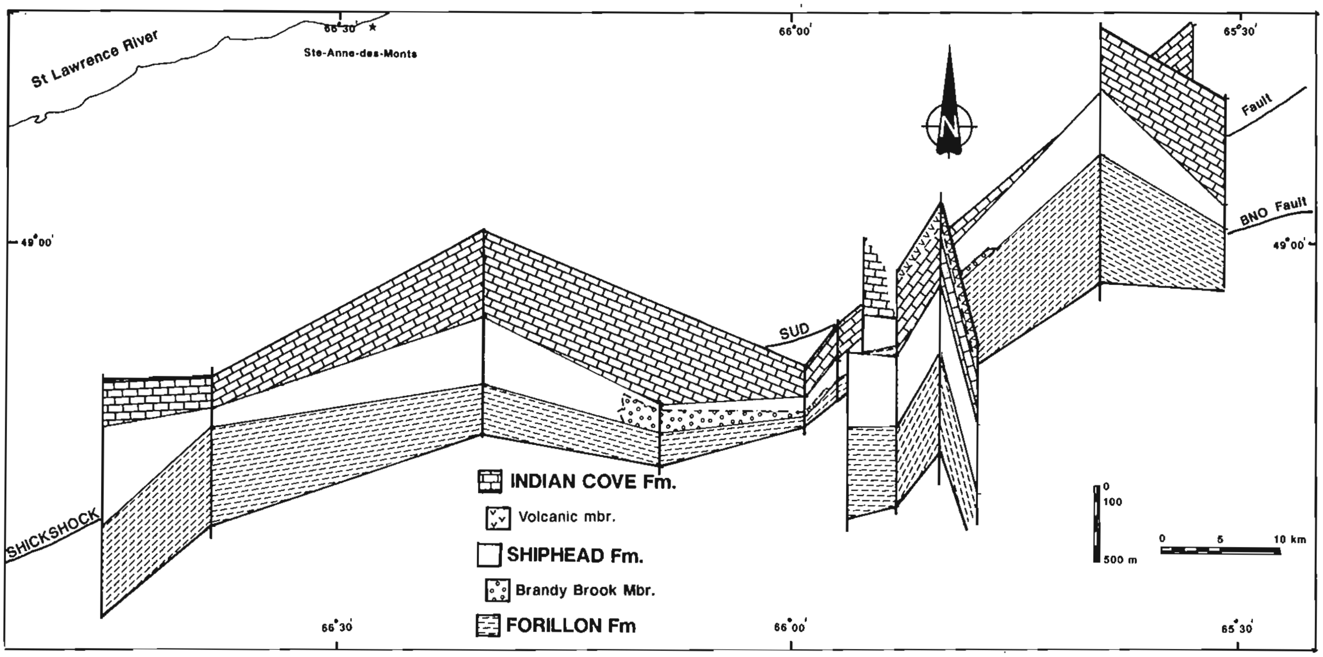

Figure 20. Early Devonian megafacies distribution in central Gaspé Peninsula.

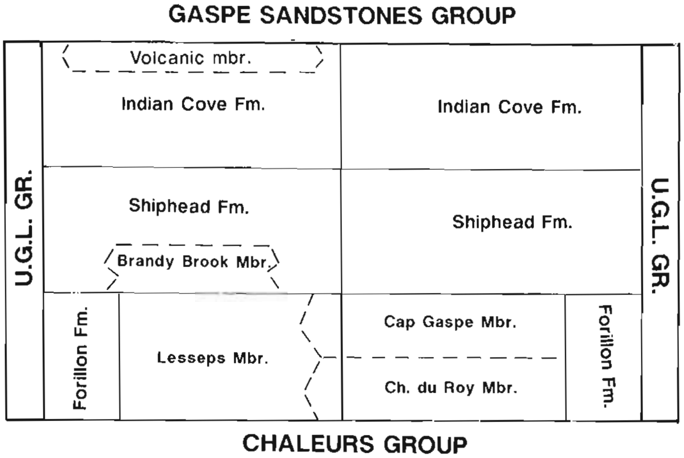

Figure 21. Lithostratigraphic framework and internal correlations for the Upper Gaspé Limestones Group (U.G.L.GR.) in central and eastern Gaspésie. The left side refers to the 1990 study area and the right side to the 1989 area.

(channels, cross-bedded laminations) point for a south-southwest source area.

— Felsic volcanism in the Early Devonian Gaspé Basin culminates in Shiphead time, going from bentonite beds and few tuffaceous mudstones in the eastern adjacent area (Lavoie et al., 1990) and northern segment of the 1990 surveyed area, to a succession of bentonite — mafic and felsic tuff — rhyolite and pyroclastic polymictic breccia in the south. The volcanic centre(s) were likely located not far from the southernmost outcrop belt of the group based on the north-south lateral lithofacies variations of the volcanic units.

— The Indian Cove Formation is composed of a rather homogeneous succession of siliceous to cherty calcilutites. The uppermost metres of the formation have a higher siliciclastic content and are locally characterized by the presence of a concretion-bearing lithofacies with small to large-sized clasts of typical shallow water environments set in a lithofacies typical of outer shelf setting. This is in turn overlain by a volcanic unit at the top of the formation, displaying evidence of eruption in a shallow marine environment as shown by the allochems-bearing pyroclastic breccias and vesicular basalts. Moreover, the basalt vesicules are locally plugged by geotropic infilling and likely early shallow marine water cements.

The evidence of synsedimentary volcanism in the 1990 study area is most convincing. This volcanic evidence was restricted, in the eastern adjacent area (Lavoie et al., 1990), to bentonite horizons within the Shiphead Formation. Volcanism was irregularly active from Shiphead to Indian Cove time. Bimodality of this volcanism was observed and lithofacies are, in many cases, suggestive of subaerial eruptive events and shallow marine depositionnal environments (see also Doyon et al., 1990). The volcanic centre(s) was probably surrounded by a narrow (?) shallow marine water belt (atoll-like ?), as suggested by the presence of shallow marine water lithoclasts and bioclasts embodied in volcanic rocks and by evidence of shallow marine sedimentological and diagenetic events in the vesicular basalts. Northward of this likely shallow water belt, lithofacies are regionally indicative of medial to deep shelf setting. To the south, relations are obscured by intense faulting. Lower Devonian rocks are represented by the siliciclastic Fortin Group.

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Critical review by D. Brisebois (MERQ) has resulted in a much improved manuscript.

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Examples of geotechnical investigation using ground probing radar surveys

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Abstract

Ground penetrating radar (GPR) can provide two dimensional stratigraphic and structural information for geotechnical investigation. With new digital GPR systems, survey parameters can be tailored to the individual investigation. Results presented here were obtained in diverse environments: frost evolution beneath road beds, ice wedges detection and distribution in the Arctic, thin sandstone bed detection in red schists deposits, and deep structural and stratigraphic investigation in arid climate. GPR proved to be a suitable investigation technique in all these cases.

Résumé

Le géoradar peut fournir des informations stratigraphiques et structurales bidimentionnelles pour des études géotechniques. Avec les nouveaux systèmes numériques, les paramètres de levé peuvent être ajustés selon les cas étudiés. Les résultats présentés ici ont été obtenus dans différents environnements: l'évolution du gel sous des routes, la détection et la répartition des coins de glaces dans l'Arctique, la détection de lits de grès minces dans des shistes rouges et une étude stratigraphique en climat aride. Le géoradar s'est avéré un outil efficace dans tous ces cas.

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INTRODUCTION

Ground probing radar (GPR) was developed as a means of rapidly and economically mapping near subsurface features. It operates on the same principles as conventional radar. The main difference is that the radar beam is directed into the ground. A short pulse of electromagnetic energy is emitted by a transmitter antenna, reflects off a distant electrical boundary, and the reflection is picked up by a receiver antenna. The time is measured for the pulse to travel from the transmitter to the receiver antenna via the reflector (Davis and Annan, 1989). The depth to a reflector can be calculated, when the propagation speed of the pulse in the material is known. The propagation velocity can be measured in situ by conducting a Common Mid Point (CMP) or a Wide Angle Reflection and Refraction (WARR) survey. In air, the pulse travels at the speed of light, 0.3 m/ns (metres per nanosecond). In the subsurface, the pulse travels at a velocity that is dependent upon the electrical properties of the material traversed. This velocity will be some appreciable fraction of the speed of light. Current operating frequencies vary from 10 MHz to 1 GHz and modern digital instruments incorporate signal stacking and digital data processing (LaFleche et al., 1987) . By moving the instrument along a survey line, one acquires a radar cross section that can be related to known earth material properties.

The overall depth to which a radar pulse will penetrate effectively is dependent upon the electromagnetic absorbtion of the subsurface. Reflectors detected by a GPR system are caused by dielectric contrasts in the subsurface materials. Common causes of subsurface reflections are material interfaces (e.g., overburden-rock, sand-clay). A large dielectric contrast exists between water and most geological materials. As a result, the presence or absence of water controls, to a large degree, the subsurface propagation characteristics of the radar pulse. Thus, the ability of a material to retain water within its pore space is an important factor in the determination of the bulk electrical properties.

Examples

Frost in roadbeds

The first case history presented here is a study of frost evolution in road beds undertaken to further the understanding of its long term effects on highway maintenance and their lifespan. This study was conducted with the help of the Ministère des Transport du Québec (MTQ). The objective was to look at frost distribution and evolution on a seasonal basis. Surveys were conducted at a series of test sites north of Montréal in the Laurentians. Each of the sites was instrumented in collaboration with MTQ with thermistor cables and fast neutron probe access wells. In addition, the drilling for these installations furnished a clear granulometric profile at each site (Tremblay et al., 1990).

The following example is from Sainte-Agathe-des-Monts, 90 km north of Montréal at the end of Highway 15. This is one of five sites studied. It consists of a paved highway section and a gravel highway section, located about 30 m apart. Drilling below the paved road indicates bedrock at a depth of 2.3 m, under a thin clay silt layer overlain by the compacted sand. The stratigraphy under the gravel road is 2 m of sand and gravels with clay traces between 2 and 2.5 m. Bedrock was not reached in this case.

The MTQ study indicates that, from 1979 to 1989, the average frost penetration beneath the highway is 150 cm, with a maximum thickness of 172 cm in 1982 and a minimum of 131 cm in 1983. Figures 1A, B, C show a sequence a GPR data across the paved road, beginning with maximum frost penetration, partial thaw, and complete thaw. Figures 2A, B, C show a similar sequence for the gravel road at the same dates. As can be seen, the freezing cycle can easily be followed.

Massive sandstone beds in schists

The second example of a geotechnical problem to which the GPR was applied is the detection of massive sandstone beds in schists. The brick industry uses clayey schists as a prime ingredient for the manufacture of bricks. In the Québec City area, the Sillery Group consists mostly of this type of schist. However, in this formation, there are often associated sandstones. When the sandstone beds are thicker than 30 cm and numerous, this renders the material unsuitable for quarrying and brick manufacturing. The following case history is from the Brique Citadelle Inc. quarry, 25 km west of Québec City in Saint-Appolinaire. Two surveys were carried out in the field, one on the bedrock surface (Fig. 3) over an area where the overburden had already been removed, and the other on the overburden surface (Fig. 4). The geological structure is monoclinal, with all strata dipping at about 20° to the southeast. The survey set up was facilitated by conducting an experimental survey in an active quarry area. Through it, we determined that best results were obtained at a frequency of 100 MHz with a transmitter-receiver separation of 1 m, 0.25 m steps, and a time window of 320 ns. The results show that the problematic sandstone beds, which in this case are those over 30 cm thick, were successfully detected on all survey lines, even those conducted on the overburden. A subsequent trench, on the later survey line, indicated that the overburden thickness was on the average 1.5 m, and as can be seen in Figure 4. The zone interpreted as containing low dipping sandstone beds beneath the overburden actually contains beds of 30 to 40 cm thick with a 20° to 25° south dip.

Ice wedges

The third example presented here is a study of the detection of ice wedges on an alluvial terrace near Salluit, northern Quebec. This work was undertaken as part of a doctoral research program at Université Laval's Centre d'études nordiques. Two GPR profiles across polygonal nets show that detection of ice wedges and other stratigraphic details are possible. Profile SALD1 (Fig. 5) shows a 100 MHz survey across a damp area of low-centre polygons. As yet, no topographic corrections have been applied to these data. The topographic variation is slight, usually less than 0.6 m.

The maximum extent of thaw of the active layer, which rarely exceeds 50 cm, can be seen. Water at the surface slows signal penetration appreciably, thus appearing to make the ground wave drop. Water filled troughs or water-filled centres of polygons are thus readily observable. The deposit at this site is 3 m of peat containing many fine layers of sand or fine gravel from overbank river flooding. This unit is then underlain by fluvial sands and gravels. The thin bands of sands and gravels within the peat are detectable as secondary reflective layers.

Parabolic reflectors indicate the presence of ice wedges. This characteristic shape of the reflectors is due to multiple refraction surfaces within the ice body. The relatively flat tops of wedges act as large point source reflectors. The intensity of the reflection is due to the dielectric difference between the body of pure ice and the surrounding frozen mass. Some of these reflectors can be seen at 22.2 m, 30.5 m, and 41 m (Fig. 5).

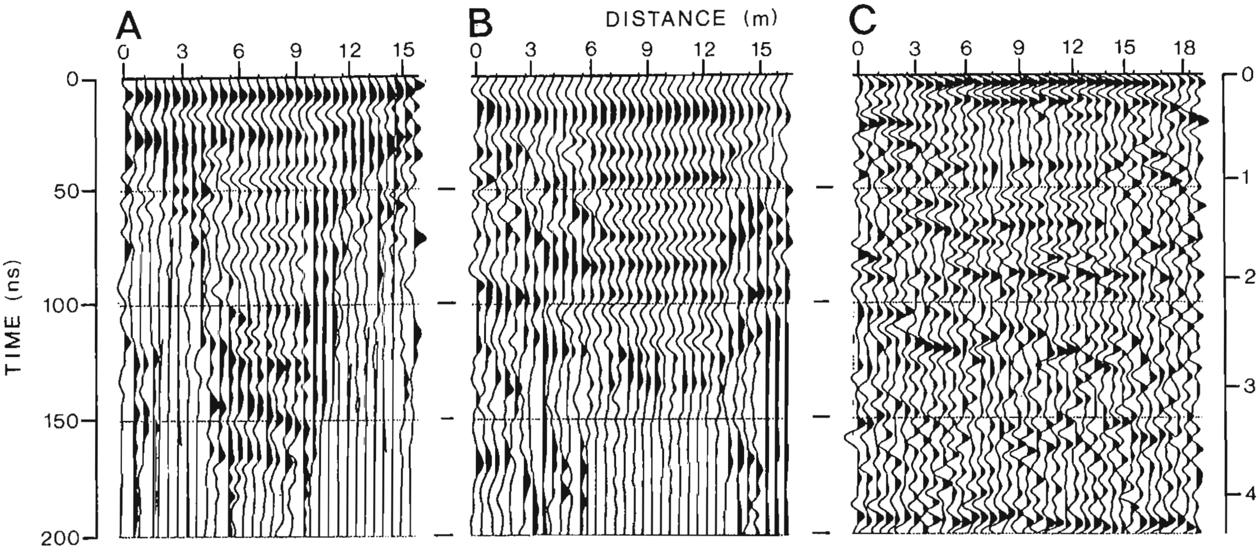

Figure 1. Radar profiles across the north-bound lane at the northern end of Highway 15, north of Sainte-Agathe-des-Monts with 50 cm intervals.

A. Survey for 20 February 1990, 100 MHz; frost extends down to approximately 65 ns (1.6 m).

B. Survey for 23 March 1990, 100 MHz; frost extends down to approximately 90 ns (1.95 m).

C. Survey for 4 May 1989, 200 MHz; no frost is observed; however, the subsurface stratigraphy is clearly seen.

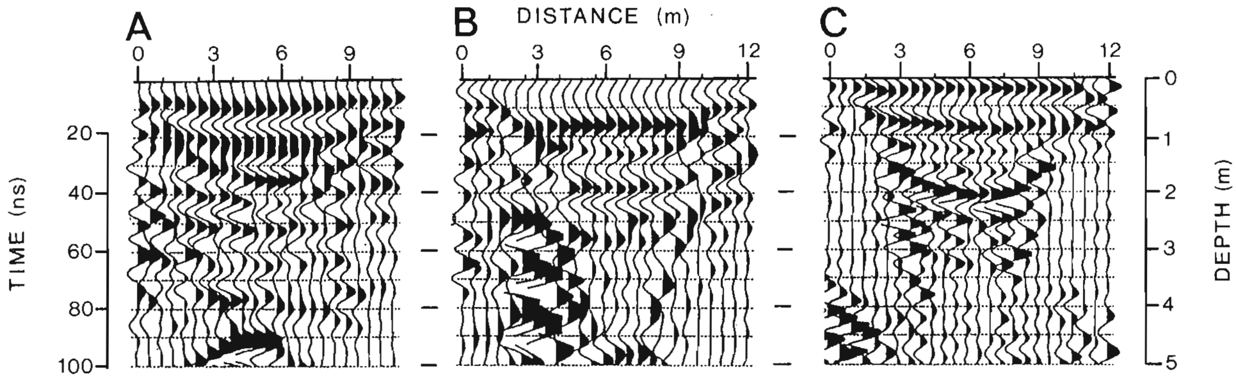

Figure 2. Radar profiles across the gravel road, north of Saint-Agathe-des-Monts with 50 cm intervals.

A. Survey for 20 February 1990, 100 MHz; frost extends, as shown by the very strong return, to 35 ns (1.75 m).

B. Survey for 23 March 1990; 100 MHz; frost extends, as shown by the strong return, to 40 ns (2 m).

C. Survey for 4 May 1989; 200 MHz; the residual frost lens beneath the road bed extends between 40 and 50 ns (2-2.5 m). Noise below 50 ns, on Stations 1 to 10 is caused by nearby power line interferences; this is also visible in Figures 2B and 2C.

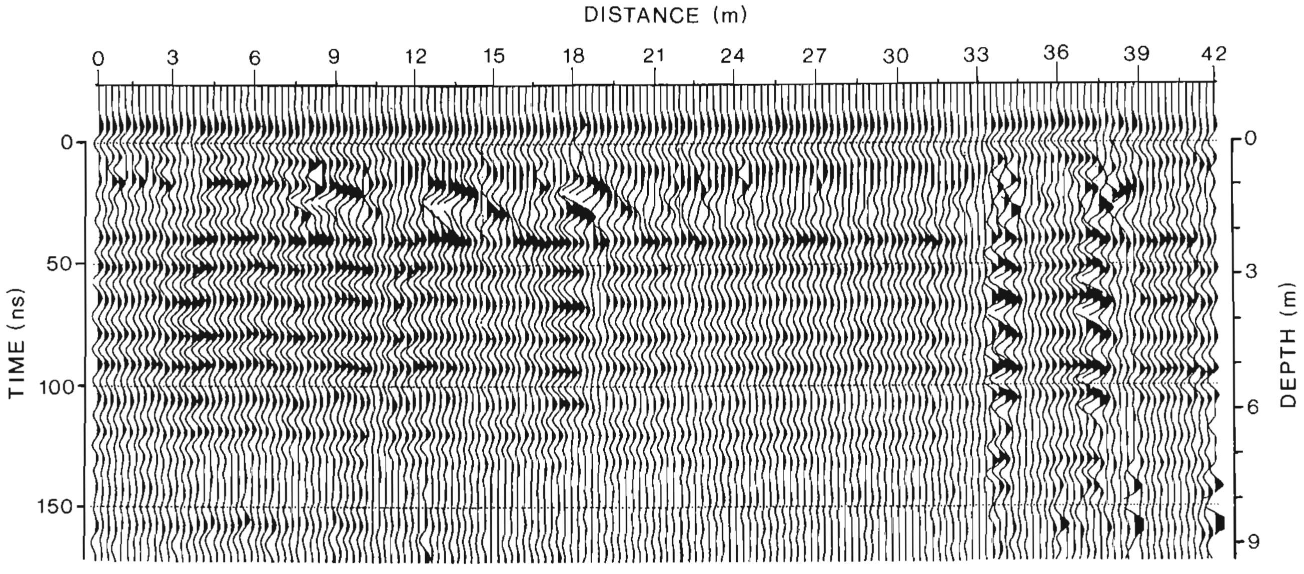

Figure 3. Radar profile in the Brique Citadelle Inc. quarry, in an area which has been stripped of overburden. Radar survey frequency: 100 MHz with 25 cm intervals. Known sandstone beds are located at 9, 12.5, 17, 34 and 37 m along the profile.

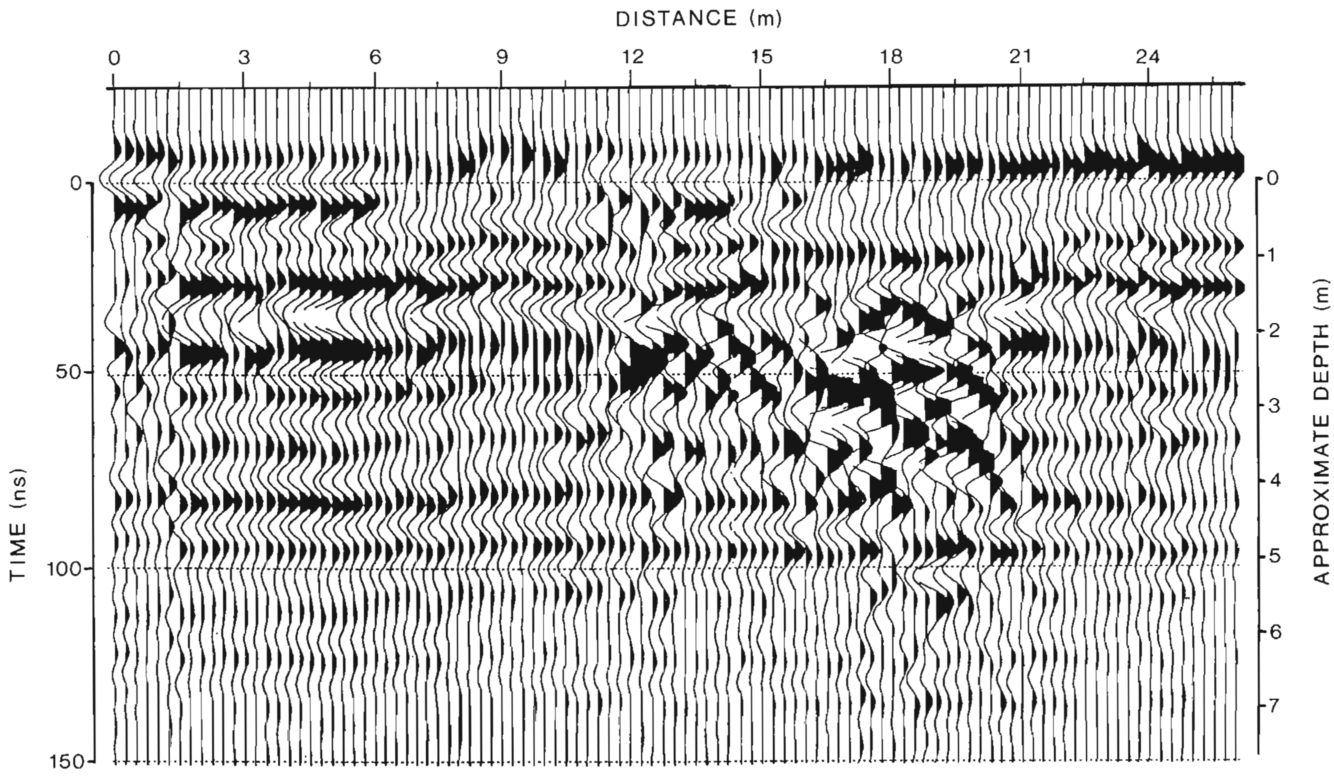

Figure 4. Radar profile in a projected quarry area through the overburden, average thickness is 1.5 m. Survey frequency 100 MHz with 25 cm intervals. Bedrock surface is visible as a flat echo at 35 ns (1.5 m). Strong echos at 50 ns beginning at 13 and 17.5 m indicate the presence of two thick massive sandstone beds (>30 cm).

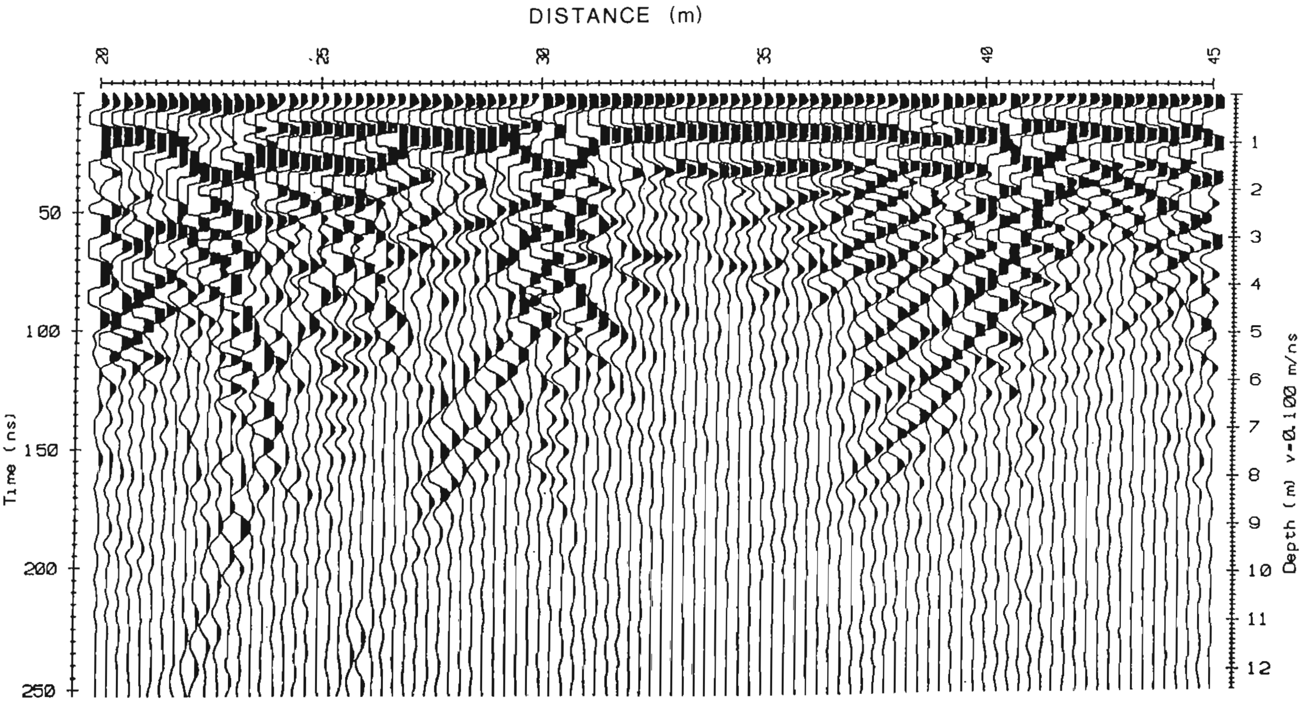

Figure 5. A 100 MHz radar survey across a damp area of low-centre polygons.

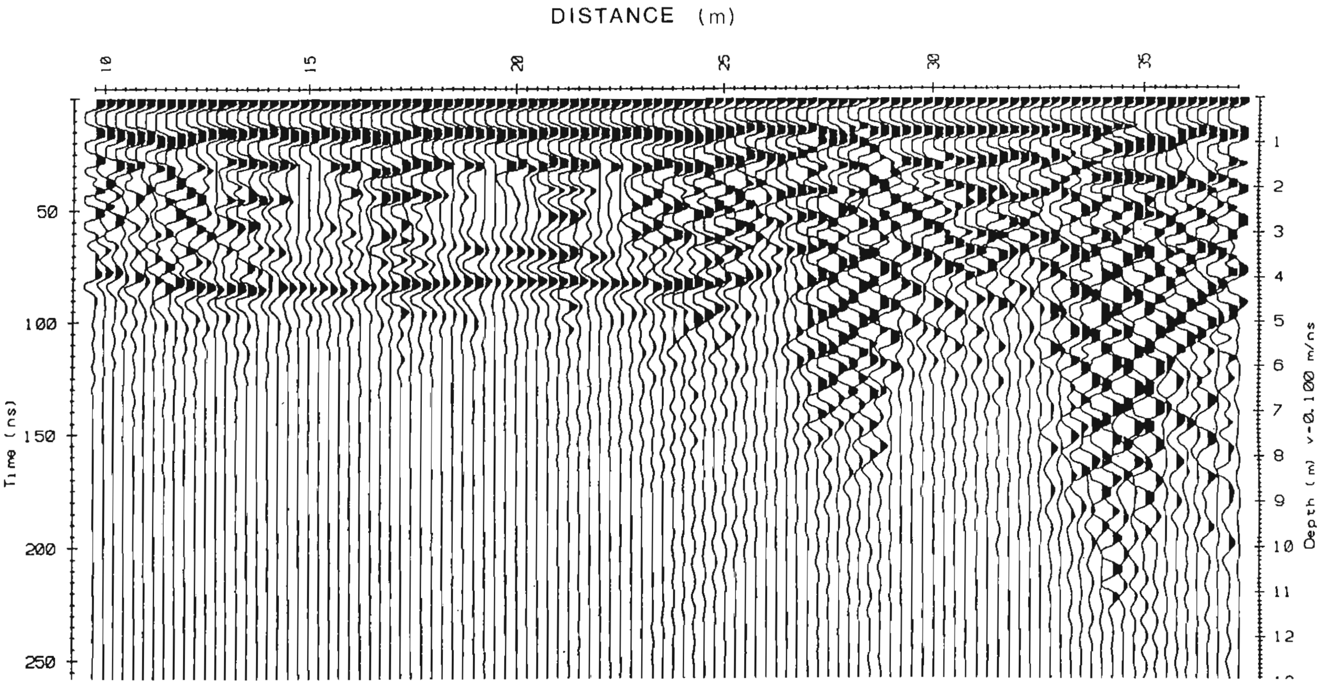

Figure 6. Transition from an area with little polygonal development to an area with a greater number of massive ice wedges.

The ridges outlining the wedge-filled furrows of low-centre polygons contain material which is commonly severely contorted by pressure exerted by growing wedges. The GPR can detect the patterns of some of these contortions. When conditions within these ridges are conducive to growth of smaller, secondary wedges, a complex reflective pattern will result (38.25 m, Fig. 5).

The second profile, SALD8 (Fig. 6) shows a transition from an area with little polygonal development to an area with a greater number of ice wedges. In the first section (10-24 m, Fig. 6) the absence of deeper-reaching ice bodies and greater dryness allow the transition from peat to the underlying sands and gravels to be easily detected. At 11.25 m, a small, shallow ice body, most likely a recently formed fissure, gives off a parabolic reflector. Here proximity to a 3 m escarpment allows for greater subsurface drainage, thus inhibiting the growth of larger wedges. The second section is farther from the embankment, and less well drained.

Comparison with the reflectors in the second half (24-40 m) shows the smaller size of this ice body and the typical parabolic reflectors point to the location of several ice wedges (Fig. 6).

One of the goals of this particular research project was to establish whether GPR can be used to reflect accurately the size and depth of ice wedges. Wedges which were scanned with GPR were excavated to determine their actual size and shape. It is hoped that GPR will prove to be as valuable a tool for this, as it is for simple detection.

Deep radar soundings

The last case illustrated is the detection of the water table in an arid environment. In the fall of 1987 a reconnaissance GPR survey was conducted at Meteor Crater, Arizona (Pilon et al., 1990). The survey was designed to acquire information on the subsurface nature of the crater and the surrounding ejecta blanket. Meteor Crater is located in the Canyon Diablo region of the southern part of the Colorado Plateau. The arid climate and the thin and sparse soil contribute to extensive exposures of bedrock on an essentially flat, low relief erosional surface. The target material intersected by the crater consists of a series of sandstones and dolomites. These lithologies offer particularly low attenuation coefficients, ranging around 0.01 mS/m, for electromagnetic radiation at frequencies around 100 MHz (Davis and Annan, 1989), which ensure good GPR signal penetration. These characteristics, combined with the deep water table approximately 230 m below the Colorado plateau surface and some 60-64 m below the bottom of the crater, depending on the season (E. Shoemaker, personal communication, 1987), make for an ideal test site.

Figures 7 and 8 present the processed GPR data and their interpretation along transect 4 across the floor of the crater. The most notable feature is the almost continuous echo obtained around 750 ns (Fig. 7), which represents the surface of the water table. The strength of this return is due to the large dielectric contrast between the unsaturated and the saturated breccia, which is mostly made up of blocks of sandstone. The dielectric contrast

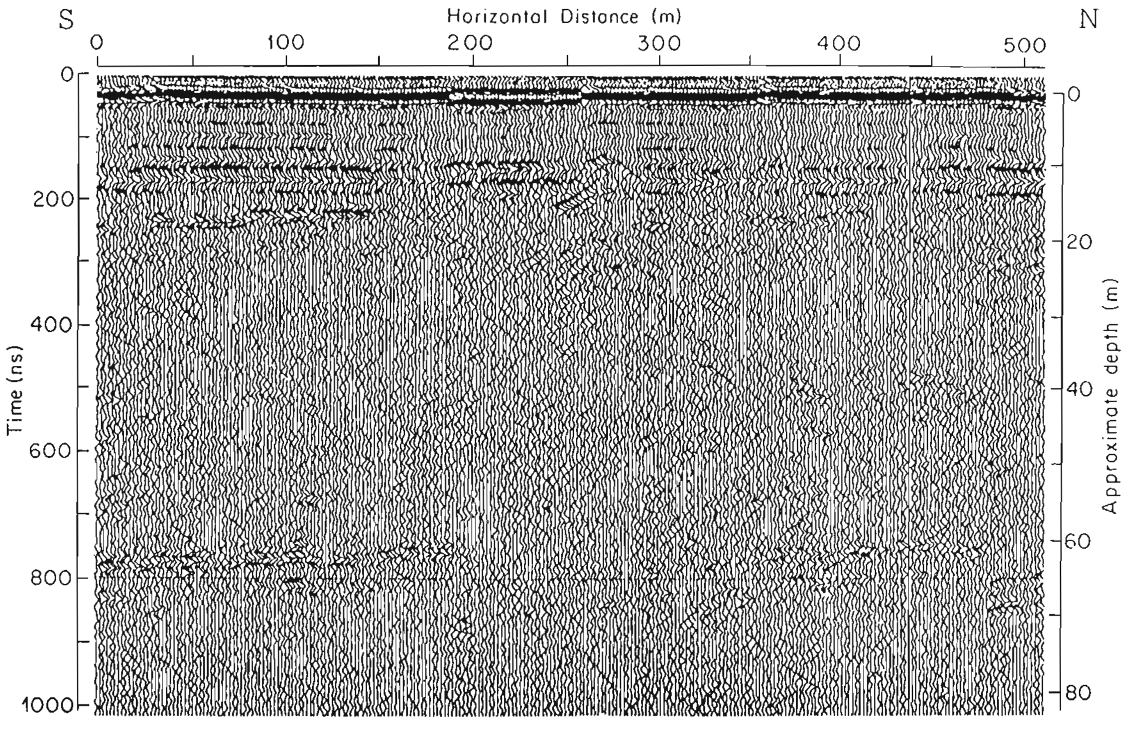

Figure 7. Radar profile across the centre of Meteor Crater. The almost continuous echo obtained around 750 ns (65 m) is the water table. Parabolic reflectors above are caused by imbedded blocks in the fallback fill.

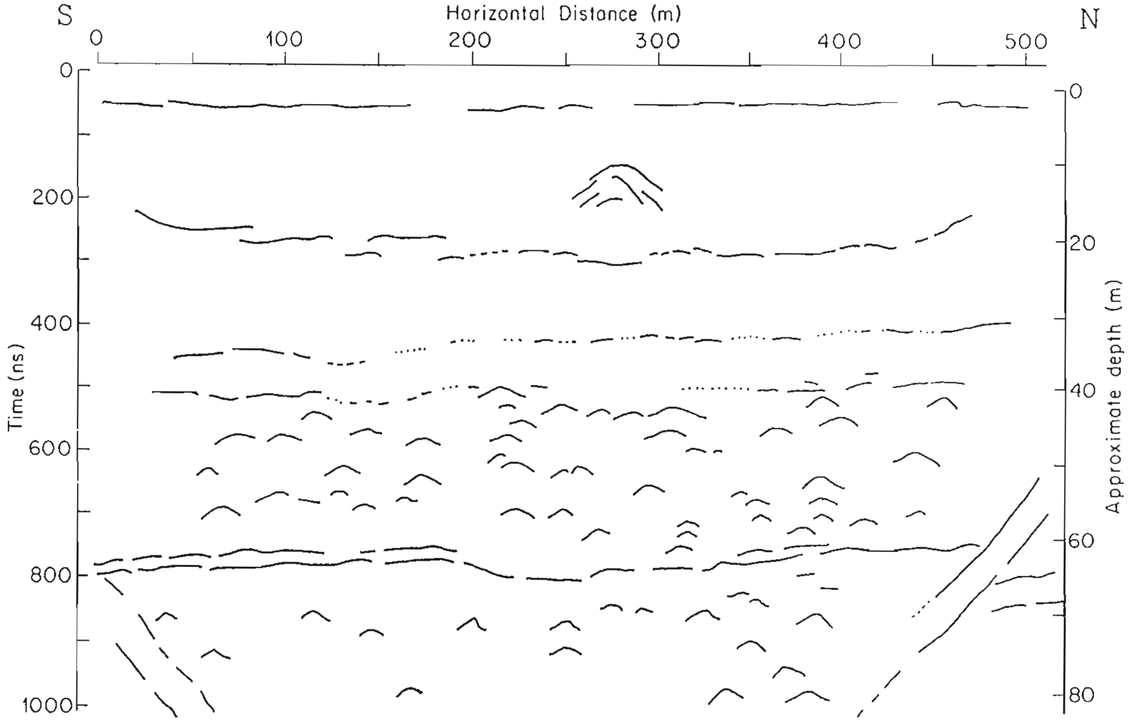

Figure 8. Interpretation of the radar profile shown in Figure 7, indicating the stratigraphy and structure.

of this material should be similar to that found between dry sand (k = 3-5) and freshwater saturated sand (k = 20-30) (Davis and Annan, 1989). Since the dielectric contrast is constant at this boundary, the variations of signal intensity observed for this boundary on the profiles are due mainly to variations in the antenna/ground coupling resulting from the surface roughness, due to boulders and small gullies on the crater floor. This is responsible for significant variations in the amount of electromagnetic energy transferred in the ground and thus of the intensity of the returns.

In addition, some details of the stratigraphy of the crater fill and the target rocks are available in this transect (Fig. 8). The distortions and discontinuity in the echoes, at what must be the horizontal water table level, are due to different speeds of propagation for the radar signal in the local target rocks on one side and the disturbed rock breccia fragments infilling the crater on the other side.

The successful interpretation of the Meteor Crater radar data in terms of crater structure resulted, in a large part, from the availability of pre-existing data on the subsurface structure. It is important to note the capability of state of the art, high powered, digital GPR systems to detect features such as the water table in arid conditions, even if they are at substantial depth from the surface (65 m). This is a capability GPR systems had not demonstrated before and which may have considerable utility in locating wells and planning irrigation schemes in semi-desert environments.

CONCLUSION

It has been illustrated by the preceding case histories that ground penetrating radar is a geophysical system allowing the fast and inexpensive collection of significant information on subsurface characteristics, such as localizing the water table, massive ice bodies, large blocks, in addition to local stratigraphic variations. Numerical data acquisition allows state of the art data processing, leading to easier interpretation. Current research and development is leading to increased penetration, still better signal to noise ratio and yet easier interpretation.

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Aeromagnetic survey program of the Geological Survey of Canada, 1990-91

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Teskey, D.J., Tod, J., Stone, P.E., Ready, E.E., Knappers, W.A., Kiss, F., Dostaler, F., and Gibb, R.A., Aeromagnetic survey program of the Geological Survey of Canada, 1990-91; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 45-48, 1991.

Abstract

During 1990-91, the GSC collected 131 950 line kilometres of aeromagnetic data. Of these, 35 550 line kilometres were flown in the northern Yukon as part of the GSC's ongoing aeromagnetic survey program, 86 400 line kilometres in Alberta during the first phase of a three-year cost-sharing government/industry survey, and 10 000 line kilometres in the Lincoln Sea area as the second phase of a joint Institute of Aerospace Research/Defence Research Establishment Pacific/Geological Survey of Canada project. The Lincoln Sea survey is now 95 percent completed. Approximately 57 000 line kilometres of aeromagnetic data collected in 1988-90 in the Yukon (24 840 line kilometres), in B.C. (21 636), and in the Lincoln Sea (10 742) were processed during the year. The GSC was commissioned by the Canadian International Development Agency (CIDA) to monitor the scientific and technical aspects of a contract survey in Zimbabwe (145 000 line kilometres). This contract represents the third and final phase of the CIDA-sponsored aeromagnetic survey of Zimbabwe.

Résumé

En 1990-1991, la CGC a recueilli des données aéromagnétiques sur 131 950 kilomètres linéaires. De ce total, 35 550 kilomètres ont été survolés dans le nord du Yukon dans le cadre du programme continu de levés aéromagnétiques de la CGC, 86 400 kilomètres linéaires ont été survolés en Alberta au cours de la première étape d'un levé de trois ans financé à parts égales par le gouvernement et le secteur privé et 10 000 kilomètres dans la région de la mer de Lincoln dans le cadre de la deuxième étape d'un projet conjoint auquel participaient l'Institut de recherche aérospatiale, le Centre de recherche pour la Défense du Pacifique et la Commission géologique du Canada. Le levé de la mer Lincoln est maintenant terminé à 95 %. Au cours de 1988-1990, on a recueilli des données aéromagnétiques sur quelque 57 000 kilomètres linéaires au Yukon (24 840 kilomètres linéaires), en Colombie-Britannique (21 636) et dans la mer de Lincoln (10 742), données que l'on a traitées au cours de l'année. La CGC a été chargée par l'Agence canadienne de développement international (ACDI) de surveiller les aspects scientifiques et techniques d'un levé réalisé sous contrat au Zimbabwe (145 000 kilomètres linéaires). Ce contrat représente la troisième et dernière étape du levé aéromagnétique du Zimbabwe parrainé par l'ACDI.

INTRODUCTION

The aeromagnetic survey program of the Geological Survey of Canada is now in its forty-fourth year of operation. Since the early sixties the program has operated primarily as a contract program in which contractors bid for and are awarded contracts to fly, compile, and deliver the final printed products and digital data (Hood et al., 1985). From the mid-seventies to the mid-eighties, some experimental high-sensitivity and gradiometer surveys were carried out by the GSC's Queenair aircraft during the development of that system, the technology of which was transferred to the private sector. In 1988 the Queenair aircraft system was sold to industry. From 1984 to 1987, the Queenair was used in surveys of the Great Lakes as a contribution to "Lithoprobe" studies and to the North American Magnetic Anomaly Map. Contracting of these latter surveys would have been complicated by the necessity to acquire data across the international boundary.

Since the early sixties, the National Aeronautical Establishment (now Institute for Aerospace Research (IAR)) has collected data in the Arctic to examine particular problems in co-operation with the GSC. In 1989, IAR commenced a survey in the Lincoln Sea in co-operation with the Defence Research Establishment Pacific (DREP) and the GSC. The IAR/DREP/GSC survey in the Lincoln Sea was extended in 1990. In 1990, other aeromagnetic survey activity included completion of the flying phase of the northern Yukon survey and the first phase of a joint GSC/industry cost-sharing survey of Southern Alberta. In addition an agreement was signed in September with the Canadian International Development Agency (CIDA) under which the GSC will act as technical and scientific authority for a 145 000 line kilometre survey in Zimbabwe, the third part of a program to complete the coverage of that country. Flying commenced in September with completion expected by March 1991.

Survey activity for 1990-91 is summarized in Figure 1 and in Table 1.

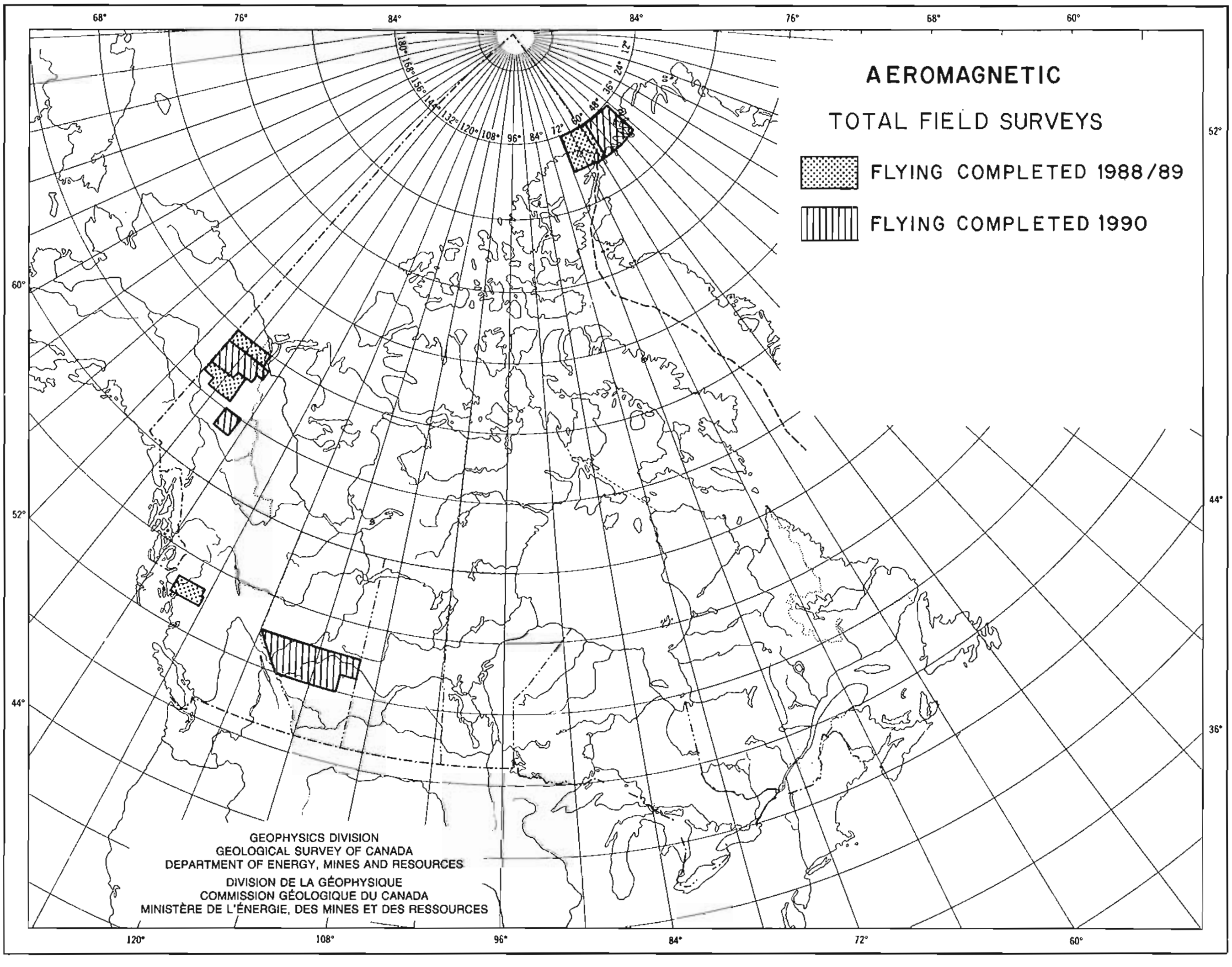

Figure 1. Locations of aeromagnetic total field surveys for 1990-91

Table 1. Aeromagnetic surveys

Survey	Type	Line Km	Line Spacing	Altitude (m)
Yukon (1988-90)	Aeromagnetic Total Field	24 840	2 km (southern portion) 3 km (northern portion)	2 745 Baro. 2 135 Baro.
Yukon (1990-91)	Aeromagnetic Total Field	35 550	2 km (southern portion) 3 km (northern portion)	2 135 Baro. and 300 Radar
British Columbia (1988-90)	Aeromagnetic Total Field	21 636	2 km	2 900 Baro.
Lincoln Sea (1989-90)	Aeromagnetic Total Field	10 742	4 km	305 Radar
Lincoln Sea (1990-91)	Aeromagnetic Total Field	10 000	4 km	300 Radar
Alberta (1990-91)	Aeromagnetic Total Field	86 400	1.6 km	Various Baro. Altitudes
Zimbabwe (1990-91)	Aeromagnetic Total Field	145 000	2 km	300 Radar

NORTHERN YUKON

The Northern Yukon survey (Fig. 1) was initiated in 1988 (Teskey et al., 1989). The initial contract was terminated after completion of 25 000 km of the planned 57 000 km in 1989 due to problems experienced by the contractor with weather and navigation. The survey blocks completed under that contract have now been compiled. A second contract was let in 1990 to complete the area. Although weather and diurnal conditions again made operations difficult, the availability of the Global Positioning System (GPS) network has greatly improved data production. Flying has now been completed and compilation has commenced.

SOUTHERN ALBERTA

The Southern Alberta survey was suggested by G. Ross of the GSC's Institute of Sedimentary and Petroleum Geology (ISPG) in order to complement seismic studies of the structure beneath the western Canadian basin being carried out under the "Lithoprobe" program. Because of the interest by oil companies in the structure and its effect on location of oil deposits, a number of companies were approached as possible participants. As a result of this initiative, a memorandum of agreement was signed between six oil companies whereby the western basin would be flown in a three-year program, with the GSC and the companies (collectively) paying approximately 50 percent of the cost each. Two mining companies subsequently joined the consortium. The first phase of the program (Fig. 1) was flown in the summer of 1990 and initial products and data were delivered to the companies. According to the agreement the data will be made available to the general public five years after delivery to the participants. Accurate flight path recovery, using combined GPS and inertial guidance and tight specifications, has resulted in data that should be capable of detecting intrasedimentary anomalies and magnetic kimberlites.

LINCOLN SEA

The Lincoln Sea survey, jointly funded by IAR, DREP, and GSC continued in 1990 and is now approximately 95 percent complete. Preliminary analysis of the available data indicates that it will be very valuable for defining the oceanic-continental boundary (Nelson et al., 1991). Flying is expected to be completed in May 1991 as part of a program that also includes the commencement of a new survey in the area northwest of Ellef Ringnes Island.

ZIMBABWE

The current Zimbabwe survey represents the third and final phase of a program to provide complete aeromagnetic coverage of that country to assist in geological mapping and mineral exploration. The GSC will act as technical and scientific authority for this CIDA-sponsored project which is scheduled to be completed by late 1991.

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National gravity survey program of the Geological Survey of Canada, 1990-91

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Hearty, D.B., and Gibb, R.A., National gravity survey program of the Geological Survey of Canada, 1990-91; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 49-52, 1991.

Abstract

In 1990, six gravity surveys were completed under the national gravity survey program: three were reconnaissance surveys located in the Arctic, in northern British Columbia, and on Lake St Clair and Lake Simcoe; two were local surveys over targets in New Brunswick and northern Quebec; and one was a gravity control station inspection survey in the southern Yukon and northern British Columbia. More than 2000 new gravity stations were added to the National Gravity Database as a result of these surveys. In addition, absolute gravity measurements were made at 14 sites in Ontario, Saskatchewan, British Columbia, and the Yukon Territory in support of the GSC's crustal dynamics program and the Canadian Gravity Standardization Network.

Résumé

En 1990, on a complété six levés gravimétriques dans le cadre du programme national de levés gravimétriques: trois étaient des levés de reconnaissance effectués dans l'rctique, dans le nord de la Colombie-Britannique et dans la région des lacs St. Clair et Simcoe; deux étaient des levés régionaux de cibles localisées au Nouveau-Brunswick et dans le nord du Québec; et un autre était un levé d'inspection d'un point fixe gravimétrique dans le sud du Yukon et dans le nord de la Colombie-Britannique. Plus de 2 000 nouvelles stations gravimétriques ont été ajoutées à la base de données gravimétriques nationale à la suite de ces levés. En outre, des mesures gravimétriques absolues ont été réalisées à 14 emplacements en Ontario, en Saskatchewan, en Colombie-Britannique et au Yukon; l'information recueillie permettra à la CGC d'étendre son programme sur la dynamique de la croûte et viendra s'ajouter aux données du réseau canadien de normalisation gravimétrique.

INTRODUCTION

In 1990-91, the national gravity survey program comprised three regional reconnaissance surveys and two local surveys over geological targets. Significant regional gravity coverage was obtained in the Foxe Basin and in northern British Columbia, two logistically difficult areas that remain to be surveyed as part of the systematic regional coverage of Canada's landmass. Gaps in coverage on Lake St Clair and Lake Simcoe were filled. Local surveys were completed over the Elmtree inlier and the Miramichi Massif in northern New Brunswick and over the Cape Smith volcanic belt in Ungava. In addition, the Geophysics Division participated in two airborne gravity test surveys in response to requests from the private sector and the United States Naval Research Laboratory (NRL). Most surveys were conducted as co-operative efforts with other divisions of the GSC or other government departments.

FOXE BASIN

A gravity and bathymetry survey of the Foxe Basin, NWT, was completed by the Geophysics Division during February and March 1990 in co-operation with the Canadian Hydrographic Service (CHS), Department of Fisheries and Oceans and with logistical support from the Polar Continental Shelf Project (PCSP). Approximately 1350 gravity and bathymetry stations were established on the ice at a 5 km grid spacing using damped LaCoste & Romberg gravimeters and Edo 9040 sounders. Horizontal positioning was established using Syledis for in-shore stations and portable GPS receivers (TANS Pathfinder) in differential mode for stations in excess of 60 km from shore. The Pathfinders were successfully deployed for the first time in a 206B helicopter using an antenna mounted on the vertical stabilizer. Variable ice conditions in the survey area hampered routine collection of gravity data and resulted in irregular coverage. Further survey activity in this area will be deferred until 1992 to coincide with CHS deployment of their Through the Ice Bathymetry System (TIBS).

NORTHERN BRITISH COLUMBIA

A regional gravity survey in northern British Columbia was completed in the summer of 1990 by the Geophysics Division in co-operation with the Pacific Geoscience Centre and the Geodetic Survey of Canada. The survey covered all or part of eight 1:250 000 NTS map sheets and more than 800 gravity stations were established at 10 to 12 km intervals. A two-pass system was used for this survey. Gravity observations using LaCoste & Romberg, Model G gravity meters and reconnaissance information were obtained on the first pass and horizontal and vertical positions were established using the Litton Inertial Survey System(ISS/LASS II) on the second pass by the Geodetic Survey. GPS was deployed occasionally during the gravity collection phase to confirm positioning and gather information on its use and capability in mountainous terrain.

LAKE ST CLAIR AND LAKE SIMCOE

A regional gravity survey of Lake St Clair was conducted on the ice by the Geophysics Division in late January 1990. Twenty-nine gravity stations were established at random intervals due to poor ice conditions and extensive open water caused by unseasonably mild weather. Bathymetry was interpolated from available hydrographic charts and Loran C was used for horizontal positioning. After forced abandonment of Lake St Clair, the regional coverage for Lake Simcoe was completed at 4 km intervals. Approximately 56 stations were established using Loran C for positioning and hydrographic charts for interpolated bathymetry. A Bell 206B helicopter was used for all traversing.

BATHURST, NEW BRUNSWICK

A local gravity survey was conducted by the Geophysics Division over the Elmtree inlier and the Miramichi Massif in the Bathurst area of New Brunswick in response to a request from the Continental Geoscience Division. Three hundred and seventy-six gravity stations were established at intervals of 200 to 700 m along seven profiles having a total length of approximately 160 km. Horizontal and vertical positions were derived from NTS maps and altimeters respectively. Horizontal positioning was supplemented by GPS along roads that were not on available NTS maps.

UNGAVA PENINSULA

In August, a local gravity survey was completed by the Geophysics Division across the Cape Smith belt west of Kangiqsujuaq in response to a request from the Continental Geoscience Division. Approximately 170 gravity stations were established at 1.5 to 3 km intervals along three profiles having a total length of 320 km. Horizontal and vertical positions were determined from NTS maps and altimeters respectively.

AIRBORNE GRAVITY

In response to a request from Intera Kenting Limited for technical assistance and instrumentation, the Geophysics Division participated in an airborne gravity test over Lake Ontario and the Alexandria Test-Range east of Ottawa. Four east-west lines were flown over both test areas using a Piper Navajo aircraft with an accurate barometer, a radar altimeter, two Ashtech MXII GPS receivers, and a LaCoste & Romberg dynamic linear gravimeter (SL1) owned by the GSC. Results are currently being analyzed and an accuracy in the range of 1 to 3 mGal is predicted.

The GSC was also invited by the Naval Research Laboratory (NRL), Washington, to participate in an airborne gravity survey as part of their effort to demonstrate that GPS interferometric navigation is sufficiently accurate for airborne gravity. The dynamic linear gravimeter (SL1) was deployed in parallel with a LaCoste & Romberg Air-Sea gravity meter (Model S), a Bell gravimeter, and three GPS navigation systems in a P-3 (ORION) aircraft for a survey over Virginia. The data processing and analysis results should be available by January 1991.

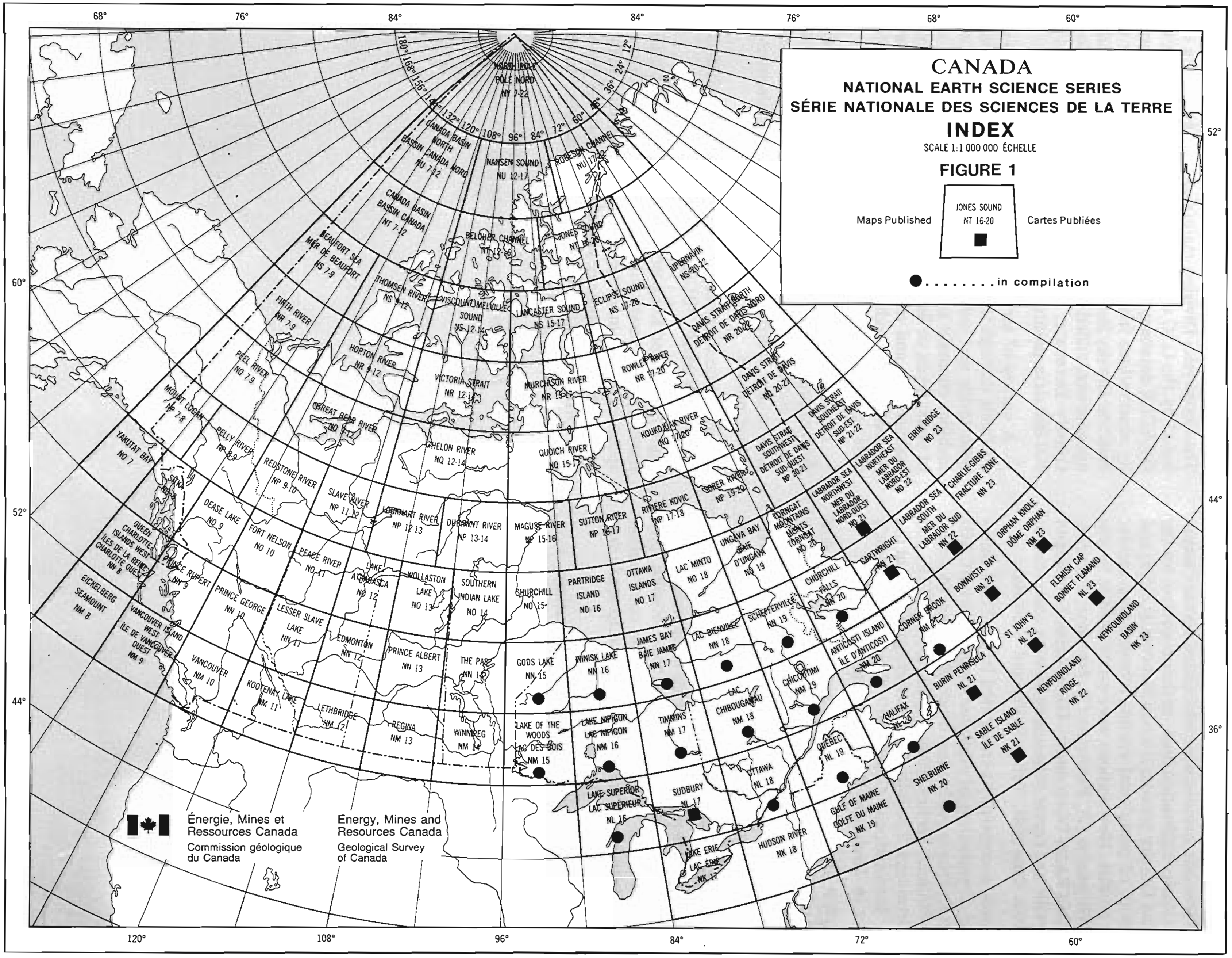

Figure 1. National Earth Science Series Index for Gravity Maps

GRAVITY STANDARDS

Several absolute stations were established across Canada for gravimeter calibration lines, as datum control for the Canadian Gravity Standardization Network (CGSN) and in support of postglacial crustal rebound and secular change studies. New stations were established at Saskatoon, Whitehorse (2), Penticton (2), Nanoose, Ucluelet, Algonquin Radio Observatory, Gananoque, Orillia, and Orangeville and repeat measurements were completed at Penticton, Churchill, Yellowknife, and Victoria.

Canada successfully participated in the Third International Comparison of Absolute Gravimeters (ICAG) in Sèvres, France with nine other countries (Austria, China, Finland, France, Italy, Japan, USA, USSR, and West Germany). The initial results were presented at the annual International Gravimetric Commission (IGC) meetings in Toulouse, France in September 1990.

A GWR superconducting gravimeter was installed in the fall of 1989 at the Canadian Absolute Gravity Site (CAGS) in Cantley, Quebec as a joint project among several Canadian universities (McGill, Memorial, University of New Brunswick, York, University of Western Ontario, Saskatchewan, and Manitoba) and the GSC. The meter has been calibrated against the absolute gravimeter and the data are used to provide a continuous gravity baseline at the national fundamental gravity site. Universities are using the data to refine the contribution of the atmosphere to gravity changes and to study free oscillations of the earth and the core.

Approximately 130 gravity control stations were inspected in the CGSN in the southern Yukon and northern British Columbia. Stations were categorized according to national standards, descriptions were updated, and reference sites re-established where necessary.

GRAVITY DATABASE

The Geophysics Division maintains all gravity data collected by the GSC on a VAX 8700 computer using the ORACLE database package. Approximately 2000 gravity observations from five surveys completed under the National Gravity Survey Program or contributed by other agencies were added to the digital holdings of the National Gravity Database. It currently contains in excess of 600 000 data points.

A variety of services and gravity data products are available through the Geophysical Data Centre to users both inside and outside the GSC. The major products include: anomaly and control station data in the form of plots, tapes, or listings; digital terrain data; open file-, manuscript-, and versatec-type maps; contour overlays; derived maps; gridded gravity values; calibration line data; earth tide values; and computer software.

In excess of 500 requests for gravity information or derived products were processed in the Geophysical Data Centre during the year. More than 300 maps were generated to user specifications on the Versatec electrostatic colour plotter.

GRAVITY MAP PRODUCTION

Eight free air and Bouguer anomaly maps in the National Earth Science Series (NESS) (scale 1:1 000 000) corresponding to four map areas in Canada's eastern offshore area were printed and are available through the publications office. Available maps and those currently in compilation are indicated in Figure 1.

One Open File Bouguer gravity map (GSC 2153) was produced for Peel Sound, District of Franklin (scale 1:1 000 000).

Seismic reflection mapping of bedrock topography and Quaternary seismo-stratigraphy of the middle St. Lawrence Estuary, Ile aux Coudres, Québec

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Todd, B.J., Occhietti, S., and Burns, R.A., Seismic reflection mapping of bedrock topography and Quaternary seismo-stratigraphy of the middle St. Lawrence Estuary, Ile aux Coudres, Québec; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 53-59, 1991.

Abstract

Analysis of high resolution seismic reflection data from the north shore of Ile aux Coudres, Québec, shows 160 m of stratified, basin-fill sediment in a broad, U-shaped valley. The sediment is probably glacial marine and would have been deposited during late Illinoian and early Sangamonian time. This seismic reflection work extends previous geological interpretations, based on regional marine seismic surveys, onto land. The results could be used in planning CCDP drillhole locations in the Middle Estuary of the St. Lawrence.

Résumé

L'analyse des données de sismique-réflexion à haute résolution, obtenues sur la côte nord de l'Île aux Coudres au Québec, montre la présence de 160 m de sédiments stratifiés de remplissage de bassin dans une large vallée en forme de U. Ces sédiments, probablement d'origine glacio-marine, se sont sans doute accumulés au cours d'une période s'étendant de la fin de l'Illinoien au début et du Sangamonien. Ces travaux de sismique-réflexion étendent aux régions terrestres une interprétation géologique jusque-là faite en fonction de levés de sismique marine régionale. On pourrait utiliser les résultats obtenus pour planifier les emplacements des sondages du PFCC dans l'estuaire moyen du Saint-Laurent.

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INTRODUCTION

A recent Canadian Continental Drilling Program (CCDP) drillhole proposal (Syvitski et al., 1990) has focussed attention, among three projects, on the shore of the Ile aux Coudres Channel in the middle St. Lawrence River estuary (Fig. 1). This site has particular significance in the study of Global Climatic Change because it represents a thick sequence of sediments underlying a well known exposure of material deposited between apparently the end of the climatic optimum of the Sangamonian Interglacial Stage (120-115 ka) and the Holocene Epoch (10 ka) (Brodeur and Allard 1985; Occhietti 1989; Occhietti and Clet, in press). A drillhole at the Ile aux Coudres site would provide detailed information on the Illinoian/Sangamonian transition and the beginning of the Sangamonian (135-115 ka). The continental climatic change during this transitional period is poorly documented. From ocean-bottom drill cores, a decrease in the ^{18}O concentration suggests a rapid decay of the world continental ice sheets during this time (Syvitski et al., 1990). The Ile aux Coudres site is attractive because approximately 160 m of material was deposited over 20 ka (135-115 ka). This sedimentation rate provides a greater degree of possible stratigraphic resolution than in any oceanic setting, where sedimentation rates are lower.

The application of the seismic reflection technique to Quaternary studies has become well established in the 1980s and the method has been tested in a variety of geological settings both in Canada and worldwide (Hunter et al., 1984; Jongerius and Helbig, 1988). To provide an accurate depth to bedrock and a reconnaissance investigation of the seismo-stratigraphy at the potential drill site, "optimum offset" seismic reflection survey lines were shot in the vicinity of Pointe de la Prairie, northwest Ile aux Coudres, in June 1990 (Fig. 1). Data from this survey are presented in this report, along with a geological interpretation.

METHOD

Under favourable conditions, the "optimum offset" high-resolution seismic reflection profiling technique (Pullan and Hunter, 1990) can provide a detailed picture of the overburden-bedrock contact and of the overburden stratigraphy. The ability of the system to resolve subsurface features depends on the frequency of seismic energy returned from reflectors to the surface. Experience has shown that the best geological conditions for the technique consist of near-surface fine-grained materials that are water saturated. The tidal mudflats on the northwest shore of Ile aux Coudres provide just such an environment.

The "optimum offset" refers to a source-receiver separation that allows reflectors to be recorded without interference from "noise". Each trace on an optimum offset profile represents ground motion recorded by a geophone at a constant distance from a seismic energy source.

Approximately 2 km of optimum offset data were shot at Ile aux Coudres. Line 1 was shot parallel to the shoreline at Pointe de la Prairie (Fig. 1). Line 5, situated in the middle of the mudflats, was shot 348 m WNW of Line 1. Line 2 intersects these two lines and reaches an elevation of approximately 30 m a.s.l. at its southeastern end. Two short lines, 3 and 4, were shot on top of the island at an elevation of about 80 m a.s.l. Lines 1, 2, and 5 are shown in this report. The source-geophone offset in all cases was 18 m and the geophone spacing was 3 m. The data were filtered in the field by high-frequency geophones (50 Hz) and by a pre-A/D 300 Hz high pass filter on the seismograph. A 12-gauge Buffalo gun was used as the seismic energy source (Pullan and MacAulay, 1987). When the gun could not be pushed into the saturated sediment, shot holes were drilled to a depth of 1.5 m.

The data were recorded on a Scintrex S2-Echo engineering seismograph. The record length was 200 ms of 1024 samples for all lines except Line 5, which was recorded to 300 ms. Preliminary processing and plotting were carried out in the field office. Final processing of the data included: 1) shifting traces in time to align first arrivals (refractions from the top of the water table); 2) digital filtering (300-800 Hz bandpass); and 3) the application of an automatic gain control and time-varying gain tapers.

A velocity-depth function was calculated from an analysis of reversed wide-angle spreads shot at the southwest and northeast ends of Line 1. These spreads consisted of 96 channels with a 3 m geophone spacing. Refracted arrivals on the seismic data indicated overburden velocities of 1650-1725 m/s. Depth scales on the seismic sections are given with respect to the ground surface.

SEISMIC RESULTS

Line 1

Figure 2A shows the 972 m-long optimum offset profile along Line 1. Interpretation lines are superimposed on the section in Figure 2B. A prominent reflector (erosional unconformity) at a depth of 120-160 m (U1) is interpreted as the top of bedrock. This reflector forms a broad U-shaped valley. There are no coherent reflectors beneath this surface. The valley is infilled with up to 160 m of highly stratified material. Coherent, conformable reflectors can be traced the length of the section and these reflectors onlap the basement reflector. At the top of the section (< 10 m depth), the stratified basin-fill exhibits a low-angle unconformity (U2) with an overlying unit.

Line 5

Line 5 (Fig. 3) shows the bedrock unconformity (U1) farther offshore (Fig. 1). The northerly dipping surface of the bedrock valley extends from 180 to 200 m in depth. Similar to Line 1, the material overlying bedrock exhibits many coherent internal reflectors and is separated from an overlying thin unit by a low-angle unconformity (U2).

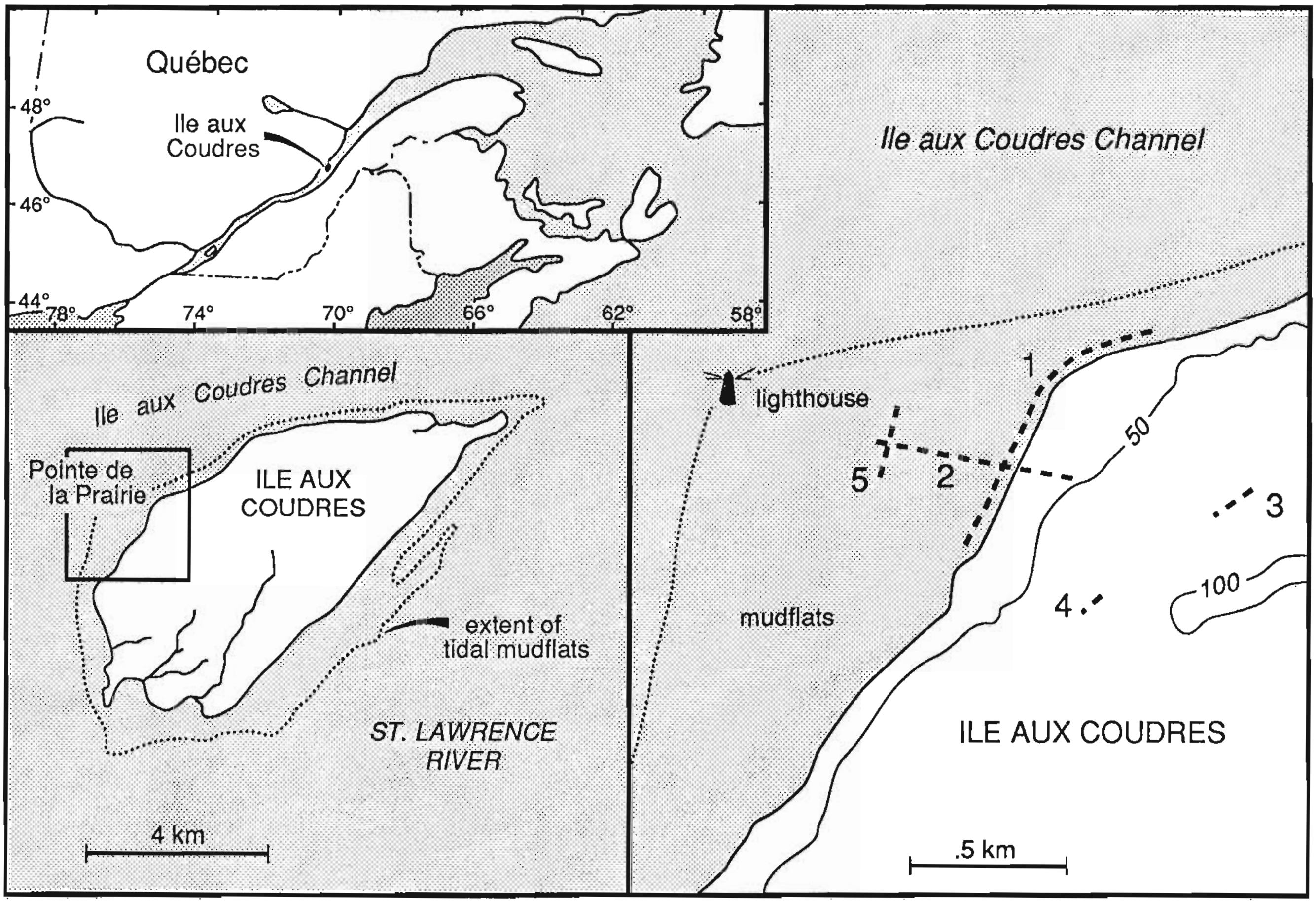

Figure 1. Location map of Ile aux Coudres. Outlined area at Pointe de la Prairie in map at lower left shows extent of detailed map at right where seismic reflection lines 1 to 5 are shown (dashed). The proposed CCDP drillhole site is near the intersection of Lines 1 and 2. Contour interval is 50 m. Modified from Brodeur and Allard (1985).

Line 2

Line 2 (Fig. 4) intersects Lines 1 and 5 (Fig. 1). The apparent dip of the bedrock unconformity (U1) is to the WNW at a depth of about 100 m to > 160 m. Reflectors within the valley-fill material can be correlated from Lines 1, 5, and 2.

INTERPRETATION

The pronounced basement reflector on the seismic records suggests a strong velocity/density contrast with the overlying sediments. Refracted sedimentary arrivals show a velocity of 1650-1725 m/s. No refracted basement arrivals were obtained on Lines 1, 2, or 5. However, refracted basement arrivals recorded on Line 3 on top of Ile aux Coudres (Fig. 1) show a velocity of 3.7 km/s. Outcrops of dipping mudstone occur along the northwest shoreline of the island and the velocity of 3.7 km/s may correlate with this material. The basement seen in the seismic profiles could be the eroded surface of this mudstone.

The thick sequence of stratified basin-fill material is interpreted as glacial marine, fine grained, ice-distal sediment. According to the exposed deposits, the sediment would have been deposited during late Illinoian and early Sangamonian. The sediment has similar seismic reflection characteristics as deposits mapped elsewhere in the Middle Estuary using marine seismic methods (Praeg et al. in press), but the latter are correlated to Late Wisconsinan — early Holocene Goldthwait Sea marine clays and silts. These recent deposits are commonly mapped and drilled along the shores of the St. Lawrence Estuary. The Ile aux Coudres U-shaped valley would represent a trap of older sediments.

The thin unit lying unconformably on the thick, stratified valley fill is interpreted as basinal muds and sandy and/or gravelly lags (Praeg et al., in press). This material is evidence of postglacial sedimentation patterns (Syvitski and Praeg, 1989).

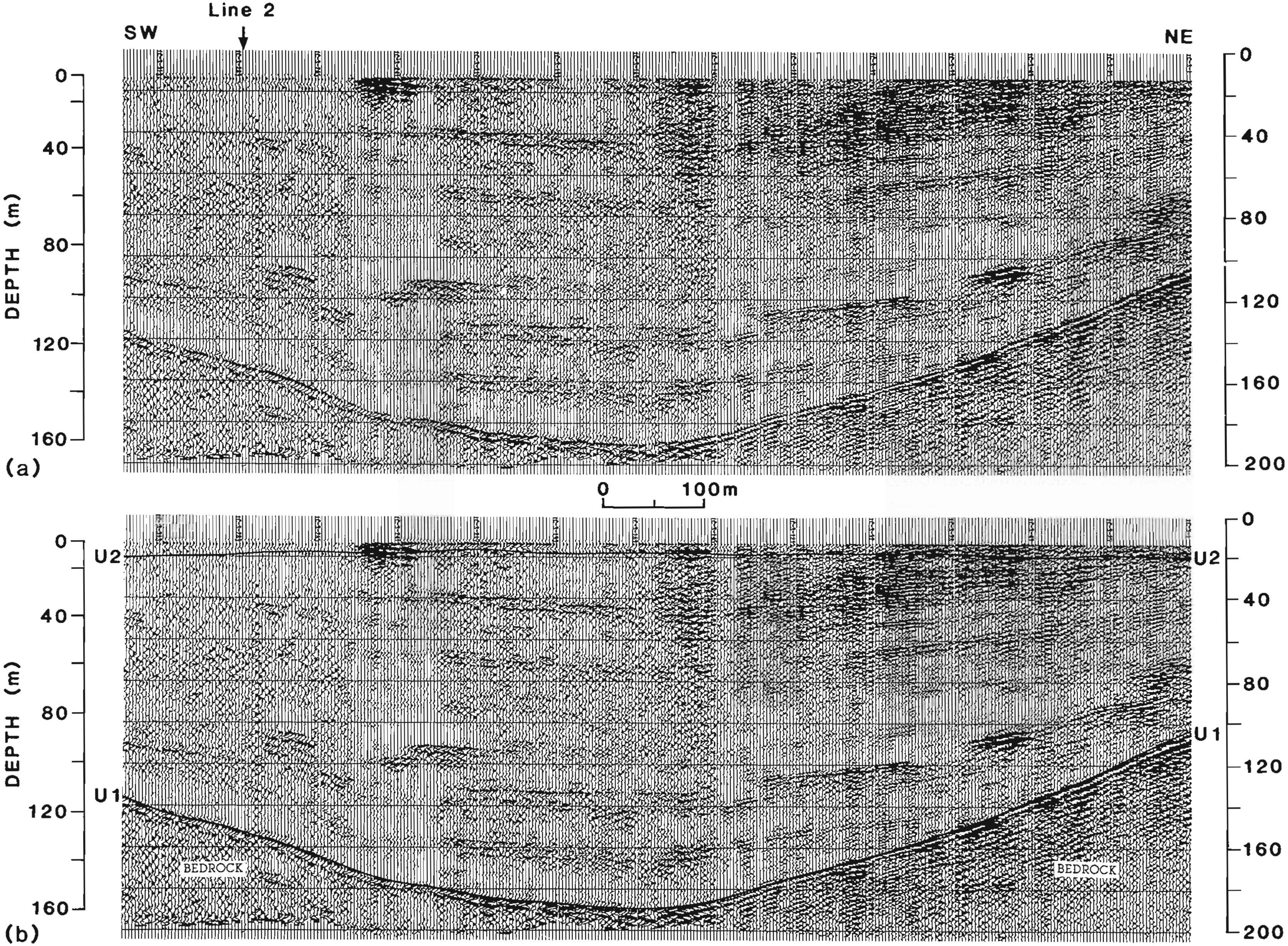

Figure 2. A. Optimum offset seismic reflection profile along Line 1. B. Interpretation of A, showing bedrock topography and seismo-stratigraphic units. This section, and subsequent sections, have a vertical exaggeration of approximately 1:1 for a velocity of 3.7 km/s. See text for discussion.

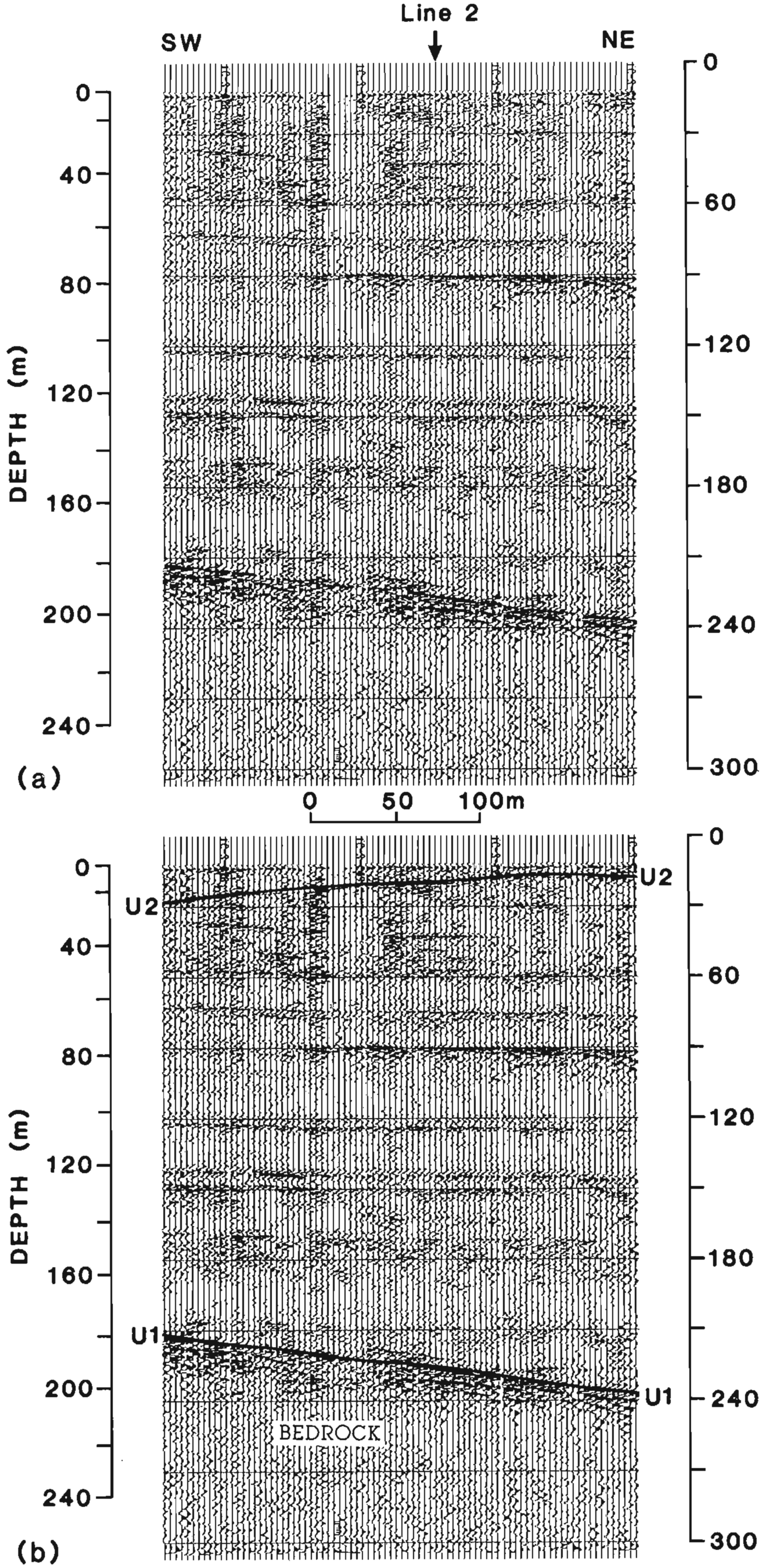

Figure 3. A. Optimum offset seismic reflection profile along Line 5. B. Interpretation of A, showing bedrock topography and seismo-stratigraphic units.

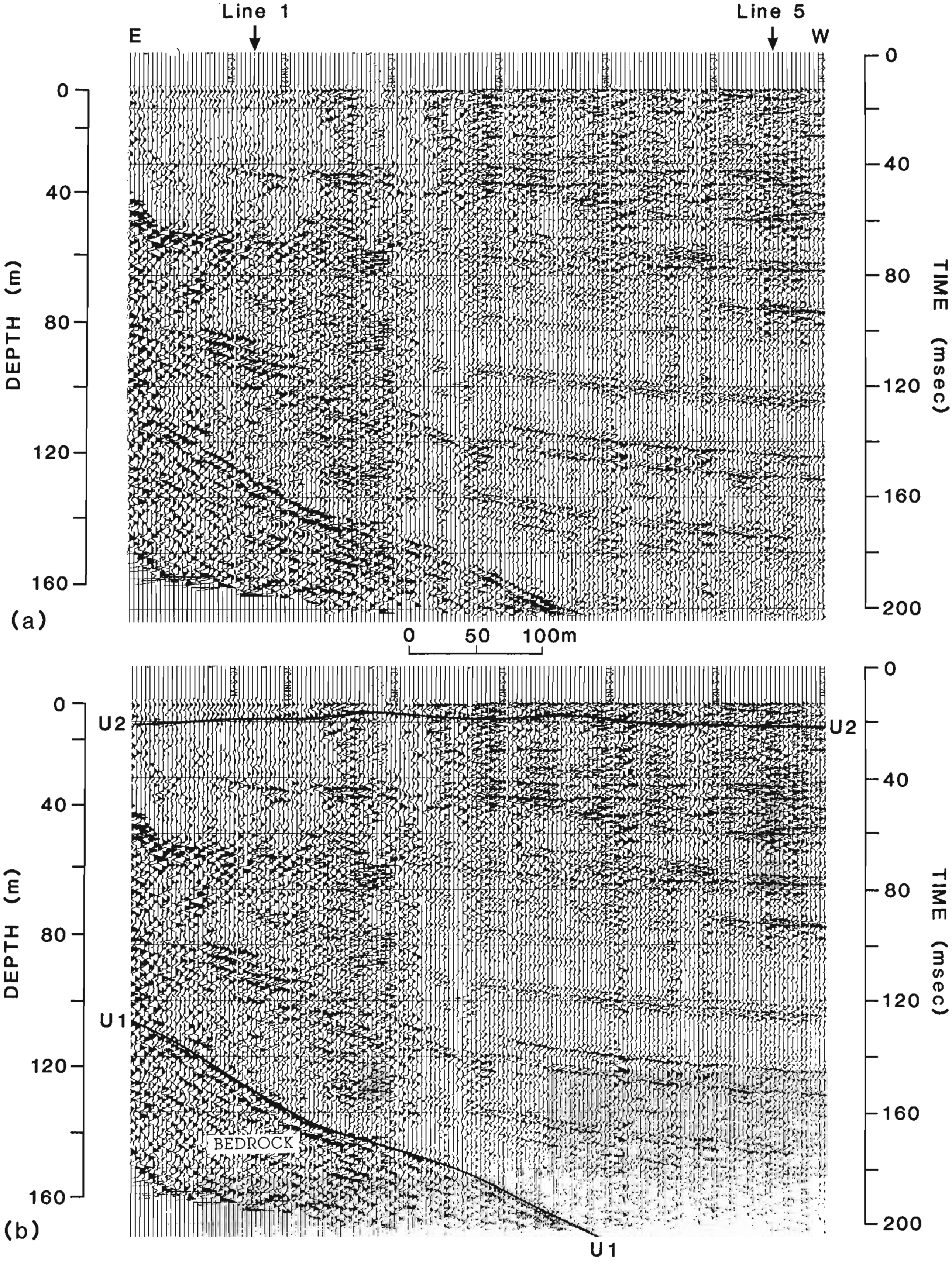

Figure 4. A. Optimum offset seismic reflection profile along Line 2. B. Interpretation of A, showing bedrock topography and seismo-stratigraphic units.

CONCLUSION

The optimum offset seismic reflection method worked well on Ile aux Coudres. Bedrock topography was delineated and the seismo-stratigraphic characteristics of the 160 m thickness of basin-fill sediments were resolved. The technique would be extremely useful for future investigations of potential CCDP drilling sites in the St. Lawrence estuary and for onshore-offshore geological correlations in the area.

ACKNOWLEDGMENTS

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Determination of overburden P-wave velocities with a downhole 12-channel eel

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Hunter, J.A. and Burns, R.A., Determination of overburden P-wave velocities with a downhole 12-channel eel; in Current Research, Part C, Geological Survey of Canada, Paper 91-1C, p. 61-65, 1991.

Abstract

A 12-channel downhole eel has been constructed and tested in small diameter (5 cm), plastic-cased, water-filled boreholes in southern Ontario. A close receiver spacing of 0.5 m allows velocity determinations over vertical distances as small as 1 m. The multichannel capability is necessary to compensate for uncertainties in time zero between successive shots. With a 12-channel eel lowered down the borehole in increments of only 1 or 2 hydrophone spacings, it is possible to record a redundancy of data for each hydrophone location, and use this to reduce the significance of the picking errors. Two ways of analyzing such data sets are discussed. It is suggested that this technique allows the determination of P-wave velocities within water-saturated, unconsolidated sediments to within a few per cent, and that it may therefore have possible applications in lithological studies of overburden materials.

Résumé

On a construit et mis à l'essai, dans le sud de l'Ontario, une flûte de sondage à 12 canaux dans des sondages de petit diamètre (5 cm), garnis de plastique et remplis d'eau. Avec des récepteurs espacés de 0,5 m seulement, il est possible de déterminer les vitesses de propagation des ondes sur des distances verticales aussi petites que 1 m. La capacité de traitement multicanal est une fonction de l'appareil qui s'impose puisqu'il faut compenser les incertitudes relatives aux temps initiaux entre les tirs successifs. Lorsqu'une flûte à 12 canaux est descendue dans le sondage par accroissements de seulement 1 ou 2 espacements des hydrophones, il est possible de noter une surabondance de données à chaque emplacement d'hydrophone, et d'employer celle-ci pour réduire l'importance des erreurs de captage. On étudie deux façons d'analyser ces groupes de données. Cette technique pourrait permettre la détermination des vitesses des ondes P à l'intérieur de sédiments non consolidés et saturés en eau avec une précision de quelques centièmes, et pourrait donc se prêter à certaines applications au niveau des études lithologiques des matériaux de couverture.

INTRODUCTION

During the course of shallow seismic reflection surveying by the Terrain Geophysics Section, Terrain Sciences Division, we had the opportunity to shoot downhole seismic surveys in slim drillholes to obtain better velocity (hence depth) control for our seismic sections. These holes are commonly cased with 5 cm diameter thin-walled plastic casing and are water-filled.

The initial downhole work used a single hydrophone lowered at intervals down the hole, and the 12-gauge "Buffalo" gun (Pullan and MacAulay, 1987) as a seismic source on surface. Since both high density digital recording (0.05 ms sample rate) and high frequency data are necessary to pick first arrival times accurately when downhole intervals of the order of 1 m are used, we found that, between successive shots, errors in the determination of zero time of only a fraction of a millisecond became very significant. These errors are attributed to timing errors inherent in the electronic trigger switch, as well as small positioning errors and varying ground conditions from shot to shot. With a reference surface geophone, we were able to obtain useful average velocity-vs-depth data; however, attempts to obtain accurate interval velocities with this single hydrophone system came to nought because of residual timing errors.

CONSTRUCTION OF DOWNHOLE EEL

To improve the timing uncertainties and to attempt the determination of detailed interval velocities, we designed and built a 12-channel downhole eel with 0.5 m spacing between hydrophones. Early prototype versions of the eel used Benthos AQ-16 hydrophones and preamplifiers encased in a 4 cm diameter oil-filled sleeve. A much more flexible unit could be made, however, from a standard seismic bay cable design, with the hydrophone and amplifier moulded onto the end of a pigtail (Fig. 1). By ensuring that less than 0.5 cm of moulding compound was placed around one side of the hydrophone, no loss of signal strength in comparison to the oil-filled eel configuration could be measured. This design also ensured ease of access to repair or replace components. Thus, with an active section of 12 hydrophones spanning a length of only 5.5 m, it is possible to obtain interval velocities across any number of hydrophones (3-12) which are independent of errors associated with time zero. Also, if the eel is lowered in small increments (1 or 2 hydrophone spacings), it is possible to obtain a redundancy of data, which may be used to reduce the significance of picking errors.

From experience we have found that for boreholes of 50 m depth or less, sufficient high frequency energy is

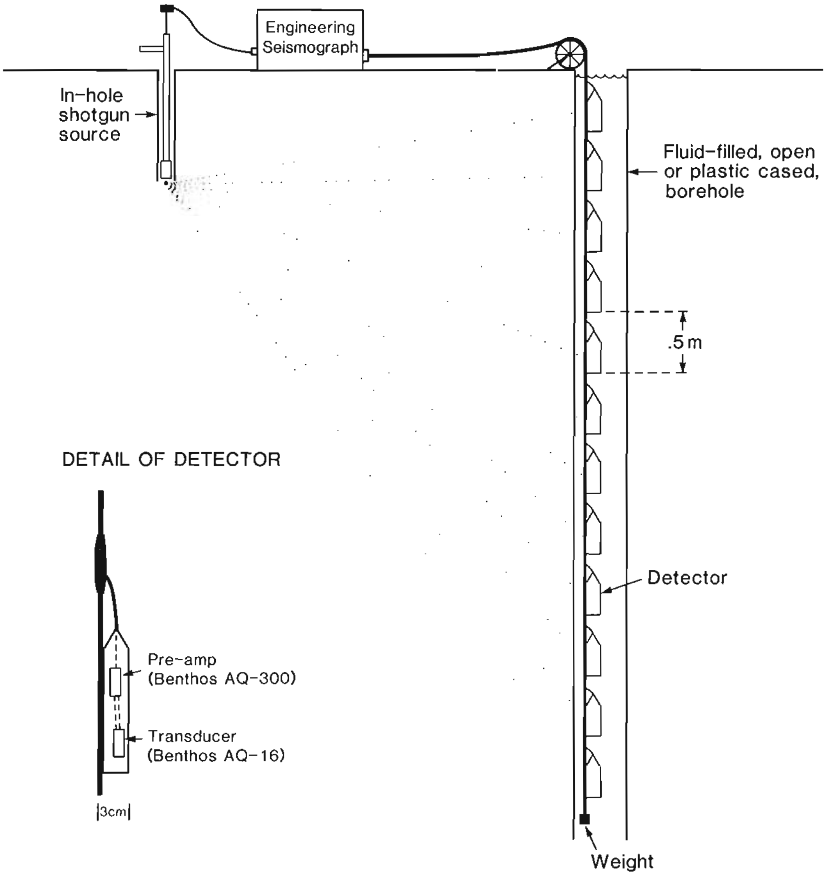

Figure 1. Diagram of 12-channel downhole eel.

transmitted so that the determination of first arrival times may be made to an accuracy of ± 0.05 ms. Figure 2 shows an example of the type of high-quality data that can be obtained. The move-out of first arrival times across the 12 channels (5.5 m) is of the order of 3-5 ms in water-saturated unconsolidated sediments.

DATA ANALYSIS

We have devised two differing interpretation schemes for redundant data sets, where the eel is lowered 0.5 or 1 m between successive shots, resulting in several overlapping hydrophone positions. The two interpretation schemes are outlined in Figure 3.

The "merge" method first corrects for errors in time zero by shifting all traces from one shot to best match the arrival time picks of overlapping hydrophone positions from the previous shot (with the array at a shallower depth position). This procedure is carried out progressively all the way down the hole, and can be accomplished interactively on a microcomputer. The best fit of two sets of picked first arrival times is defined as the one in which the sums of the squares of the residual times between the two data sets is a minimum. The resultant time depth plot consists of arithmetically averaged travel times from several overlapping arrays at each hydrophone position. From this merged data set, a running least squares velocity fit can be obtained from 3 or more points. Hence interval velocities across as little as 1 m vertical distance can be obtained.

The "independent" method initially analyses each 12-channel travel-time data set separately without reference to time zero. Interval velocities are computed across 3 or more data points and are assigned to the depth location corresponding to the mid-point of the hydrophone positions being used. With overlapping arrays, several independent interval velocities can be computed for the same depth point; for example, if the array is moved down the hole in 0.5 m increments, 10 independent velocity measurements from a 3 point fit can be assigned to the same depth point. The final interval velocity assigned to each depth point is the weighted average of the independent measurements; the weights used are the inverse of the least-squares standard error of the velocity fit.

Figure 4 shows a comparison of "merged" vs "independent" interval velocities for a 5 point fit (2 meter interval) from a borehole in southern Ontario in predominantly coarse-grained overburden. Error bars associated with the velocities are the standard error of the velocity fit in the case of the "merged" data set, and the standard deviation about the mean in the case of the "independent" velocity determination method. The results sug-

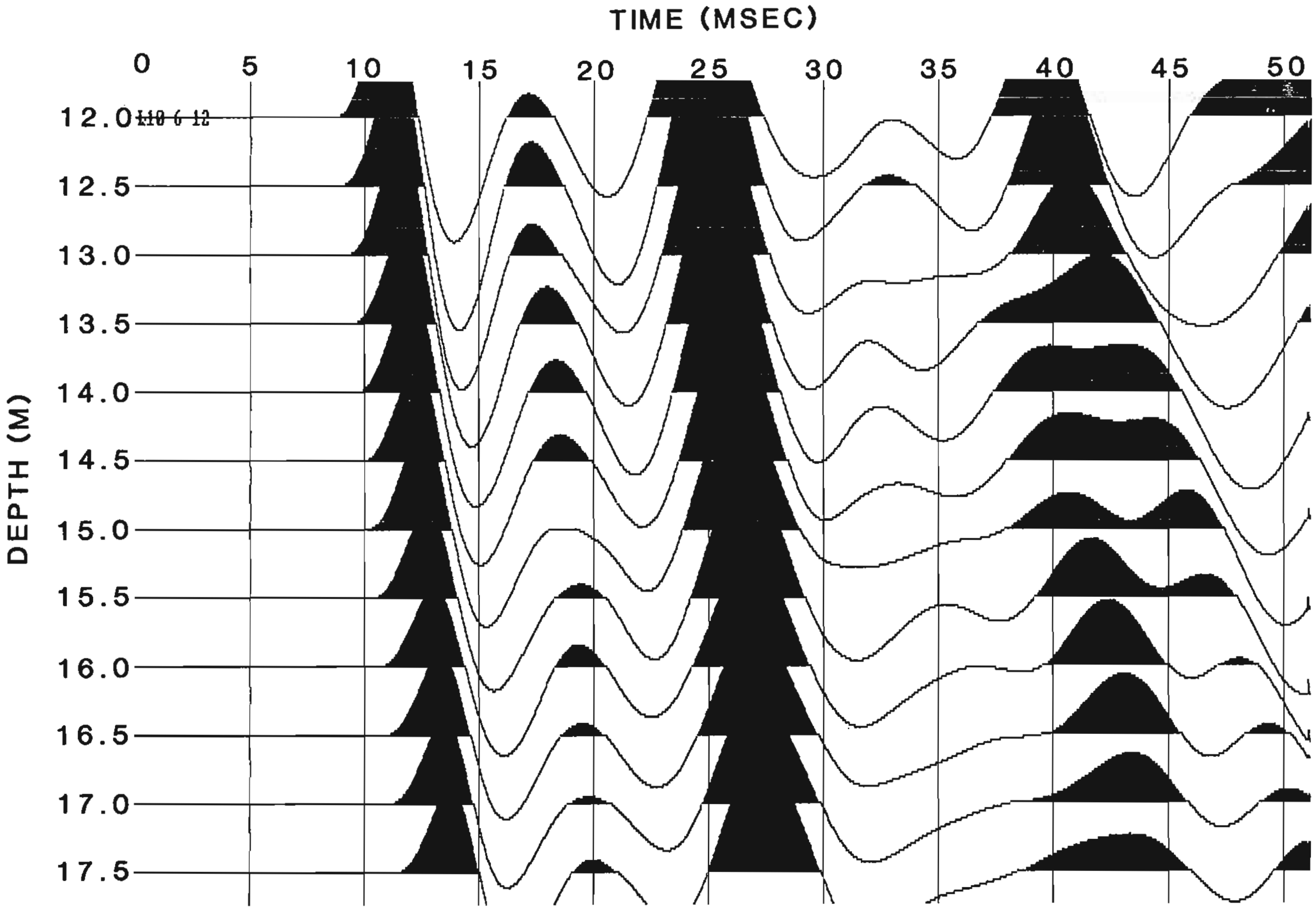

Figure 2. Seismic record obtained with downhole eel.

PICK FIRST ARRIVALS

MERGE METHOD ↙

MERGE TRAVEL-TIME SETS

- OPERATOR INTERACTIVE
- MINIMIZE RESIDUALS
- OUTPUT IS ONE TRAVEL-TIME/DEPTH DATA SET

↓

COMPUTE INTERVAL VELOCITIES

- LEAST SQUARES FITS ON MERGED DATA SET
- 3 PT, 4 PT, 5 PT, ...

INDEPENDENT METHOD ↘

COMPUTE INTERVAL VELOCITIES

- LEAST SQUARES FITS ON INDEPENDENT DATA SETS
- 3 PT, 4 PT, ... 12 PT

↓

AVERAGE INTERVAL VELOCITIES

- WEIGHTED AVERAGE CALCULATED FROM ALL DATA SETS

Figure 3. Flow diagram outlining two different methods of data analysis.

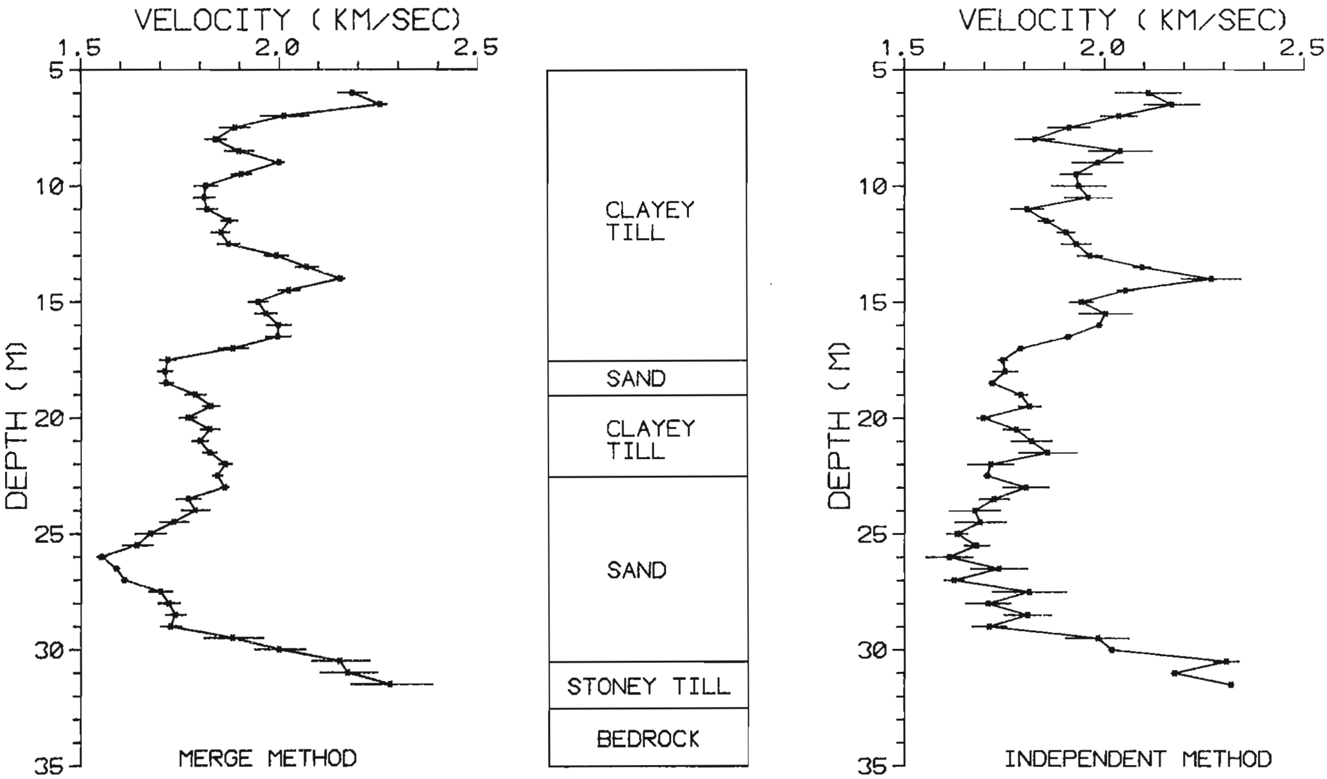

Figure 4. Comparison of the results of the "merged" and "independent" analyses of data from a borehole in southern Ontario.

gest that both methods give similar results, and that the error bars shown are reasonable estimates of the uncertainties in the determination of the interval velocities within the overburden. In this case, the velocities are considered to be accurate to within 6% (± 100 m/s).

CONCLUSIONS

Experience with a 12-channel downhole eel with hydrophones spaced every 0.5 m has shown that it can yield accurate P-wave interval velocities in unconsolidated overburden materials. The multichannel capacity is required to compensate for uncertainties in time zero between successive shots, and allows a redundancy of data to be obtained at each hydrophone location so that the significance of picking errors can be reduced. The ability to obtain interval velocities to within an accuracy of a few percent suggests that this technique may have possible applications in lithologic studies of overburden materials.

ACKNOWLEDGMENTS

The authors would like to thank R.L. Good for his help in constructing the eel and R.M. Gagné for his assistance in the field. Thanks are also due to the Ontario Geological Survey who drilled and cased the boreholes we used, and made the boreholes and borehole logs available for this work.

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Shallow shear wave reflection tests

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Pullan, S.E., Hunter, J.A., and Neave, K.G., Shallow shear wave reflection tests; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 67-70, 1991.

Abstract

Shallow shear wave reflection tests have been carried out at a variety of sites with differing surface conditions and overburden stratigraphy with the aim of obtaining shear wave reflections from the overburden-bedrock interface. Horizontal geophones (8 Hz and 50 Hz) and small "hammer and block of wood" sources have been used for most of these tests. Results have ranged from poor to excellent, with well defined bedrock reflections obtained in areas where groundroll energy is quickly attenuated and does not interfere with reflection events. In such cases, the use of shear wave reflection techniques may be complimentary to P-wave reflection methods or may offer improved resolution.

Résumé

Des essais de réflexion des ondes de cisaillement de faible profondeur ont été réalisés à divers emplacements présentant des conditions de surface et une stratigraphie des morts-terrains différentes; il s'agissait d'obtenir des réflexions des ondes de cisaillement au niveau de l'interface entre les roches de couverture et le socle. On a employé dans la plupart de ces essais des géophones horizontaux (8 Hz et 50 Hz) et de petites sources du type "marteau et bloc en bois". Les résultats ont variés de médiocres à excellents, et des réflexions bien définies ont été obtenues au niveau du socle dans des régions où l'énergie des ondes parasites de surface est rapidement atténuée et n'interfère pas avec les diverses réflexions. Dans ce cas, l'emploi de techniques de réflexion des ondes de cisaillement pourrait servir à compléter les méthodes de réflexion des ondes P, ou pourrait même permettre une meilleure résolution.

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INTRODUCTION

The Terrain Geophysics Section, Terrain Sciences has now conducted many years of experimental work in high resolution P-wave reflection profiling in unconsolidated overburden. We are beginning to recognize some of the limitations of the P-wave techniques, and believe that some may be overcome through the use of shear wave reflection methods.

In general, high frequency P-wave reflections are difficult to generate when the surface conditions consist of dry, coarse grained, and/or loose material. Since shear waves are "framework" waves and are largely unaffected by moisture content, shear wave reflection techniques may succeed in some dry areas where P-wave techniques fail. Even in areas where P-wave reflections can be obtained, shear wave reflection methods may provide better resolution because the low velocity of shear waves in unconsolidated overburden materials may result in shorter wavelengths than can be recorded using P-waves.

Recently, we conducted a series of tests with the aim of obtaining shear wave reflections from the overburden-bedrock boundary beneath a variety of overburden materials. The characteristics of both SV (vertically polarized shear wave) and SH (horizontally polarized shear wave) records are being studied to determine optimum "viewing windows" and field parameters. Taking guidance from the work of Hasbrouck (1986) and Neave and Pullan (1989), we use small "hammer and block of wood" sources, but are also examining an in-hole sideways-firing shotgun source.

Computer modelling of SH, SV, and P-SV converted waves by Norminton (1990) suggests that both SH and SV reflections from the bedrock interface should be observable at small angles of incidence, and as well, that significant energy may exist in the converted P-SV or SV-P wave.

In the early phases of this program our primary target is the bedrock surface; however, with increased experience it may be possible to obtain reflections from infra-overburden structure.

RESULTS

To date, we have found a number of sites in southern Ontario where good shear wave reflections have been obtained from the bedrock surface at depths to 60 m. At other sites, the bedrock reflection is poor or uninterpretable. These "poor" sites yield seismic records contaminated by dispersed groundroll (Rayleigh and Love waves) which mask the bedrock reflection. We believe that these sites are characterized by large shear wave velocity gradients in the near-surface, causing both the dispersion as well as attenuation of the high frequency components of reflection events. It is suggested that this effect is comparable to the situation in shallow P-wave reflection surveys when both source and receiver are well above the watertable.

An example of a record which shows some interference between dispersed groundroll and the bedrock reflection is shown in Figure 1. This record is probably typical of many field situations. At this site 35 m of glacial till and sand overlie bedrock. The surface shear wave velocity is 150 m/s, and increases with depth to 250 m/s. The result is dispersed groundroll which masks much of the viewing window. A weak bedrock reflection can be observed behind the groundroll on the near traces. The reflection event increases in amplitude and is more easily observed at wider angles; unfortunately the best signals are at offsets greater than the depth of the reflector, which suggests that the resolution of bedrock topography, or any overburden structure, will be poor.

In contrast to the record described above, the results shown in Figures 2 and 3 come from a site near Ottawa where excellent shear wave (and compressional wave) reflection records can be obtained. The depth to bedrock is approximately 24 m, confirmed by nearby waterwells. The surface material is clay, but otherwise the overburden stratigraphy is unknown.

Figure 2A is a P-wave reflection record from the site, recorded with vertical geophones and a 7.5 kg hammer and plate source. A 300 Hz low cut analog filter was applied to the data prior to digitization by the seismograph. Even so, the near traces are badly saturated with groundroll. The P-wave reflection from the bedrock interface is clearly observable beyond 12 m offset. The dominant frequency of the reflected pulse is approximately 300 Hz, corresponding to a wavelength of 5 m. This site would be considered to be an excellent one for P-wave reflection surveying, though we would probably conduct such a survey using the buffalo gun source (Pullan and MacAulay, 1987) to reduce airwave interference.

Figure 2B is an expanding spread recorded at the same site using radially oriented 50 Hz horizontal geophones and a small SV source. The source consisted of a 40 cm

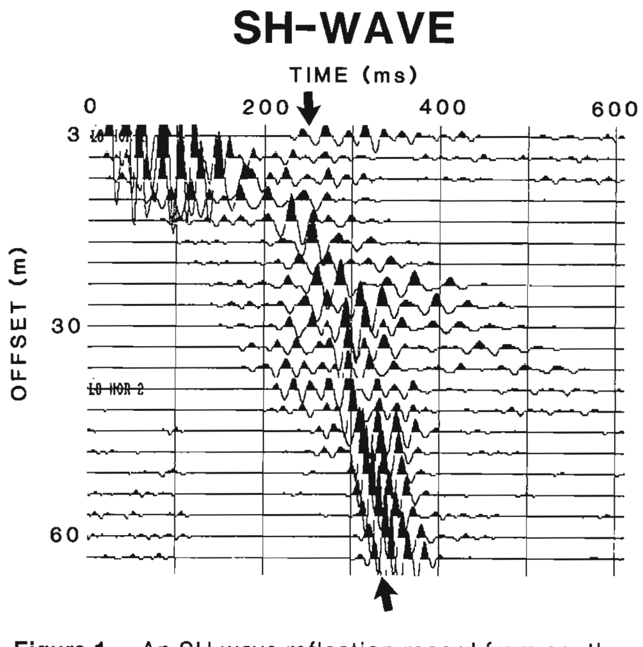

Figure 1. An SH-wave reflection record from southern Ontario, illustrating the interference of dispersed groundroll with the bedrock reflection (indicated by arrows).

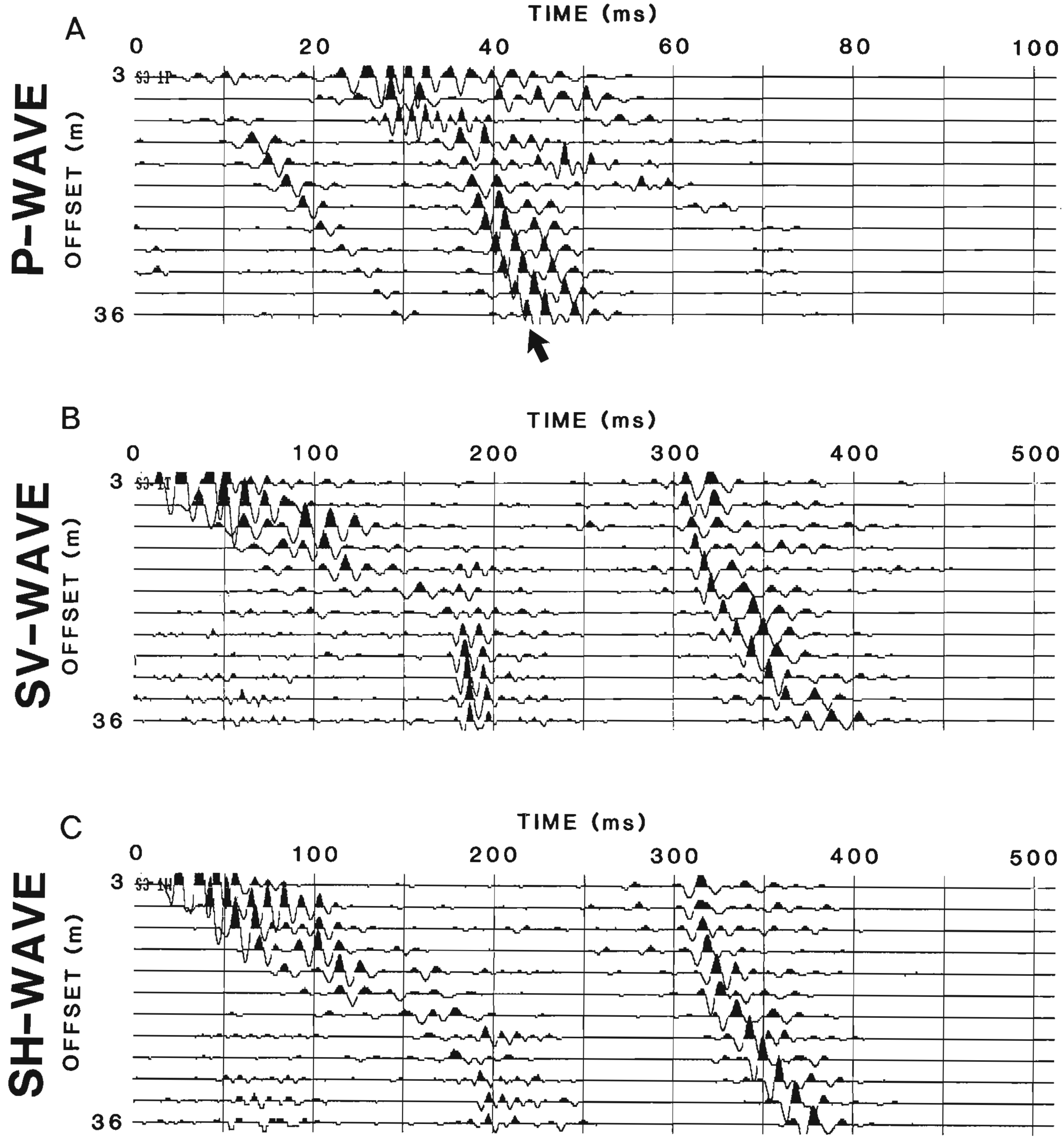

Figure 2. A. P-wave reflection record from near Ottawa. The bedrock reflection can be clearly observed at offsets beyond 12 m and is indicated by the arrow. B. An SV-wave reflection record showing a well defined bedrock reflection between 300 and 400 ms, and significant converted wave energy at approximately 180 ms. C. An SH-wave reflection record from the same site. The bedrock reflection is a well defined event, but the converted wave energy is much reduced.

long triangular piece of wood which was hit with an ordinary carpenter's claw hammer. The block was struck in such a manner as to impart polarized energy radially along the geophone array. A well defined bedrock reflection can be observed at 300 ms, with a dominant frequency in the 75-100 Hz range. Since the average measured shear wave velocity is approximately 170 m/s, the dominant wavelength of the pulse is 1.7-2.3 m. Because the groundroll packet is limited in time and distance from the source, the SV reflection can be clearly observed across the entire array and at source-receiver offsets as small as 3 m. The use of 50 Hz horizontal geophones and a small energy source probably contribute to the rapid attenuation of groundroll observed on this record.

A weaker, but well defined, event with little moveout can be seen in the 170-180 ms range. This event is interpreted to be the converted SV-P wave, with probable contributions from the P-SV conversion as well (since no seismic source is entirely SV).

Figure 2C shows another expanding spread record from the same site, with both source and horizontal geophones oriented transversely to record SH waves. As with the SV record, the bedrock reflection is clearly visible across the entire array, since the groundroll packet is quickly attenuated spatially. No obvious converted wave is evident, and theory does not predict such a conversion. However, the real (anisotropic) earth may produce some limited converted wave energy, and there is a weak indication of this on the far offsets of Figure 2C at approximately 180 ms.

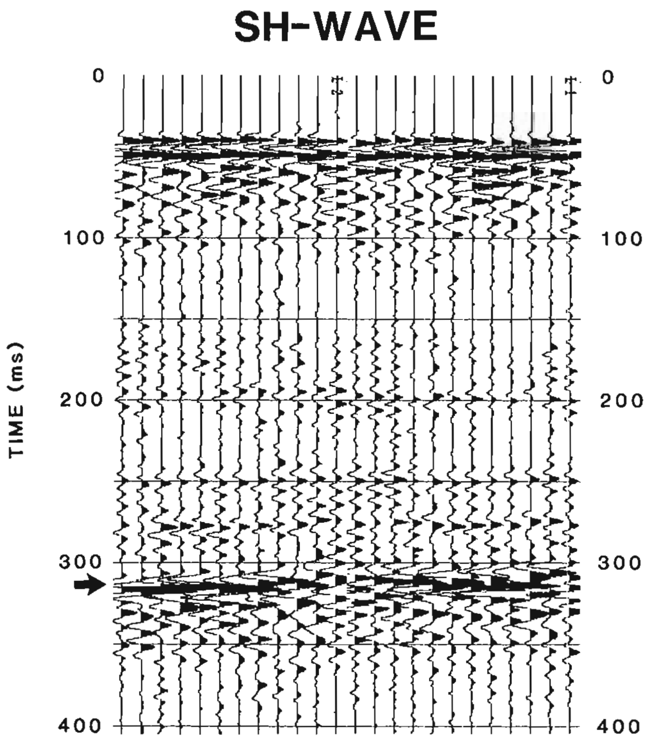

Figure 3. An "optimum offset" section from the site near Ottawa. The bedrock reflection is indicated by the arrow. Several infra-overburden reflections can also be observed.

Both the SV and SH expanding spreads suggest that there is almost an unlimited "viewing window" for the shear wave reflection from the bedrock surface at this site. Figure 3 is a portion of an "optimum offset" section shot in SH mode. The source-receiver offset for each trace is 6 m, and each trace is the result of a single hit with the claw hammer. The small offset was chosen so that first arrivals (direct shear wave energy) could be used for static corrections, and to maximize the vertical resolution at shallow depths. As well as the bedrock reflection, it is also possible to identify infra-overburden reflections on this section.

CONCLUSIONS

Shear wave reflection tests have been conducted in a number of differing geological situations. The aim has been to record reflections from the bedrock surface, and to evaluate field techniques and parameters. Results have ranged from poor or uninterpretable records to excellent ones.

The key to successful shear wave reflection surveying appears to be the nature of the groundroll packet of combined Rayleigh and Love waves. If the packet is dispersive, resulting from strong velocity gradients near the surface, the groundroll will mask much of the viewing window. As well, low shear wave velocities near surface are coupled with high shear wave attenuation, resulting in a lower frequency response.

Our initial tests are promising in areas where the groundroll is quickly attenuated. Future research will be directed towards developing methodologies to tap the potential of shear wave reflection techniques and to overcome the difficulties of working through dispersive groundroll.

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Review of the geochronology of the Cobequid Highlands, Avalon composite terrane, Nova Scotia [1]

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Doig, R., Murphy, J.B., Nance, R.D., and Stokes, T. Review of the Geochronology of the Cobequid Highlands, Avalon composite terrane, Nova Scotia; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 71-78, 1991.

Abstract

The Cobequid Highlands are underlain by Precambrian to Lower Carboniferous rocks. Recent U-Pb and $^{40}Ar/^{39}Ar$ dates have given precise information on the timing of tectonothermal events resulting in significant re-interpretation of the regional tectonic history. The 580-600 Ma U-Pb ages from the Great Village River Gneiss (GVRG) indicate it is part of the Avalonian orogenic cycle. A minimum depositional age for the Gamble Brook Formation, which structurally overlies the GVRG, is given by the 734 Ma U-Pb age of the Economy River Gneiss which intrudes it. U-Pb data also indicate that the volcano-sedimentary Folly River Formation, which unconformably overlies the Gamble Brook Formation, was deposited, deformed and intruded at ca. 610 Ma; supporting its interpretation as the product of a strike-slip environment. A pilot U-Pb study on selected Paleozoic plutons yields ages of ca. 360 Ma, which are more tightly constrained than published Rb-Sr data.

Résumé

Les hautes terres de Cobequid reposent sur des roches du Précambrien au Carbonifère inférieur. Des datations récentes obtenues au moyen des méthodes U-Pb et $^{40}Ar/^{39}Ar$ ont permis d'établir avec précision la chronologie des événements tectonothermaux et de réinterpréter, à partir de ces renseignements, l'histoire tectonique régionale. Les datations U-Pb de 580 à 600 Ma du gneiss de Great Village River (GGVR) indiquent qu'elles font partie du cycle orogénique avalonien. Un âge sédimentaire minimal de la formation de Gamble Brook qui repose structuralement sur le GGVR est donné par la datation U-Pb de 734 Ma obtenue pour le gneiss d'Economy River qui la recoupe. Selon les datations U-Pb, la formation volcano-sédimentaire de Folly River, qui repose en discordance sur la formation de Gamble Brook, s'est accumulée, a été déformée et a été pénétrée d'intrusions vers 610 Ma, appuyant l'interprétation selon laquelle elle serait le produit d'un décrochement. Une étude U-Pb pilote sur des plutons paléozoïques a donné des âges de 360 Ma environ, donc restreints à un intervalle de temps beaucoup mieux défini que celui indiqué par les données Rb-Sr publiées.

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INTRODUCTION AND GEOLOGICAL SETTING

The Cobequid Highlands of northern mainland Nova Scotia consist of Precambrian to Lower Carboniferous rocks overlain to the north and east by Upper Namurian and Westphalian clastic sediments and bound to the south by the Cobequid Fault (Donohoe and Wallace, 1982, 1985, Fig. 1). The Highlands form part of the Avalon composite terrane (Keppie 1985) which occupies much of the southeastern flank of the Appalachian orogenic belt. This terrane is characterized by ca. 630-570 Ma volcano-sedimentary rocks and co-genetic plutons unconformably or conformably overlain by lithostratigraphically correlative, subaerial to shallow marine lower Paleozoic successions containing Acado-Baltic fauna (Williams, 1979, O'Brien et al. 1983, Keppie 1985). The similarity of these lower Paleozoic rocks contrasts with the diversity of the Upper Precambrian successions and suggests that amalgamation of the composite terrane was completed by late Precambrian time (Keppie 1985). Recent mapping in the Cobequid Highlands (Murphy et al. 1988, 1989, Nance and Murphy, 1988, 1990, Pe-Piper and Piper 1987, 1989, Pe-Piper and Murphy, 1989) provides a framework by which the geological history of the Cobequid Highlands may be re-interpreted. Detailed models, however, have been hindered by poor age constraints which generally consisted of Rb-Sr and K-Ar dates. Recently, U-Pb (Doig et al., 1989, 1990, and manuscripts in preparation), and $^{40}Ar/^{39}Ar$ geochronology (Keppie et al., 1990) provide, for the first time, precise data on the age of magmatic and structural events in the Cobequid Highlands. The purpose of this paper is to review available geochronological data briefly to evaluate models for the evolution of the Cobequid Highlands.

PREVIOUS WORK

Previous work in the Cobequid Highlands was discussed in detail in Murphy et al. (1988). Preliminary maps of the Cobequid Highlands were prepared by Kelley (1966). Donohoe and Wallace (1982) published maps of the Cobequid Highlands at a scale of 1:50 000. These authors also produced reports and field guides (Donohoe and Wallace 1985, and references therein) in which most of the units outlined below were described and defined. The earliest published geochronological data were K-Ar dates from gneisses (now called the Great Village River Gneiss) and Late Precambrian plutons (Wanless et al., 1972, 1973, Fig. 2). Most of the plutonic rocks of the Cobequid Highlands have been dated using Rb-Sr geochronology (e.g. Cormier, 1979, 1980; Gaudette et al. 1984; Donohoe et al. 1986, Fig. 2). These dates provided a broad framework for the interpretation of the geological history of the Cobequid Highlands (e.g. Donohoe et al. 1986). However, precise U-Pb data on zircons (Doig et al. 1989, 1990, and manuscripts in preparation) and $^{40}Ar/^{39}Ar$ data on hornblendes (Keppie et al. 1990) provide tighter constraints on the timing of Precambrian and Paleozoic tectonic activity resulting in a re-interpretation of the geological history of the Cobequid Highlands.

PRECAMBRIAN ROCKS

Detailed descriptions of the Precambrian rocks in the Cobequid Highlands may be found in Cullen (1984), Donohoe and Wallace (1985), Murphy et al. (1988) and Pe-Piper and Piper (1987, 1989) and will not be repeated here.

Precambrian rocks in the eastern Cobequid Highlands occur in a narrow fault slice between the Rockland Brook and Cobequid faults. The Mount Thom Complex in the east consists of quartzofeldspathic gneisses and amphibolites. At present, the only available age for the complex is a 934 +/− 82 Ma Rb-Sr whole-rock isochron (1, Table 1) which has been interpreted as the age of the main metamorphic event affecting these rocks (Gaudette et al. 1984). Attempts to find zircons suitable for U-Pb analyses were unsuccessful. The more extensive Bass River Complex (e.g. Donohoe and Wallace, 1985, Fig. 1) consists of (a) orthogneisses and amphibolites of the Great Village River Gneiss structurally overlain by (b) quartzites and pelites of the Gamble Brook Formation which in turn is thought to be unconformably overlain by (c) greenschist facies mafic volcanics and turbidites of the Folly River Formation (Murphy et al. 1988, Table 2). The mafic volcanics of the Folly River Formation are intraplate continental tholeiites thought to have been deposited in an extensional tectonic regime of a volcanic arc (Pe-Piper and Murphy, 1989).

The contact between the Great Village River gneiss and the Gamble Brook Formation is a ductile shear zone that obscures the original relationships (Murphy et al., 1988). Syn- to late-tectonic granitic gneiss is spatially associated with the shear zone. Until recently, the age of the Great Village River Gneiss was poorly constrained (3, 8, 9, Table 1) but, on the basis of lithological comparison, it was correlated with the gneisses of the Mount Thom Complex (Donohoe and Cullen, 1983). Both gneisses were thought to represent mid-Proterozoic continental basement that was metamorphosed prior to the deposition of the Gamble Brook Formation (e.g. Donohoe and Wallace 1985, O'Brien et al. 1983, Murphy et al., 1989). However, U-Pb dates do not support this interpretation (Doig et al. 1989, 1990, and manuscripts in preparation). Except for the very local occurrence of 734 +/− 3 Ma gneisses on Economy River (Doig et al., manuscript in preparation), the 600-580 Ma dates for the Great Village River Gneisses (Doig et al., manuscript in preparation) suggest that they are an integral part of the Avalonian orogenic cycle.

The Gamble Brook Formation has been correlated with the lithologically similar Green Head Group of southern New Brunswick (e.g. O'Brien et al. 1983, Nance 1986) which contains stromatolites assigned to the mid-Riphean (Hoffman, 1974). Yet, until recently, the depositional age of the Gamble Brook Formation was poorly constrained. However, the 734 +/− 3 Ma gneisses on Economy River (2, Table 1, Doig et al. 1989 and manuscript in preparation) intrude the Gamble Brook Formation thereby providing a minimum depositional age for the formation and confirming its antiquity relative to the

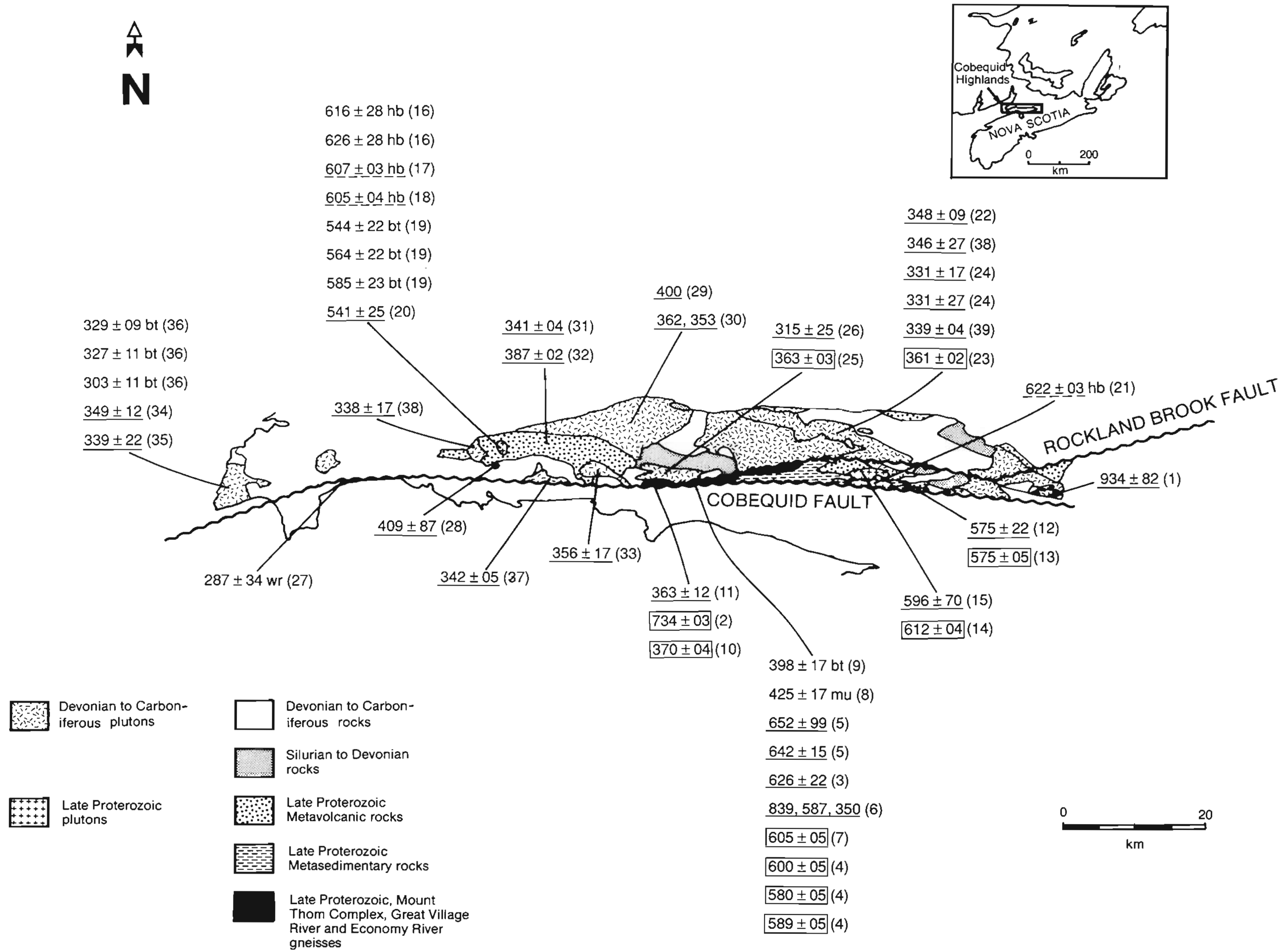

Figure 1. Geochronological compilation and geological map of the Cobequid Highlands (adapted from Donohoe and Wallace, 1982; Pe-Piper et al. 1989a; Murphy et al. 1988; and Keppie et al. 1990). Solid underlined ages are Rb-Sr whole rock/mineral isochrons; dashed underlined ages are $^{40}Ar/^{39}Ar$; boxed ages are U-Pb zircon; and other ages are K-Ar (hb = hornblende, bt = biotite, mu = muscovite, wr = whole rock). Numbers in brackets refer to source of data, see Table 1.

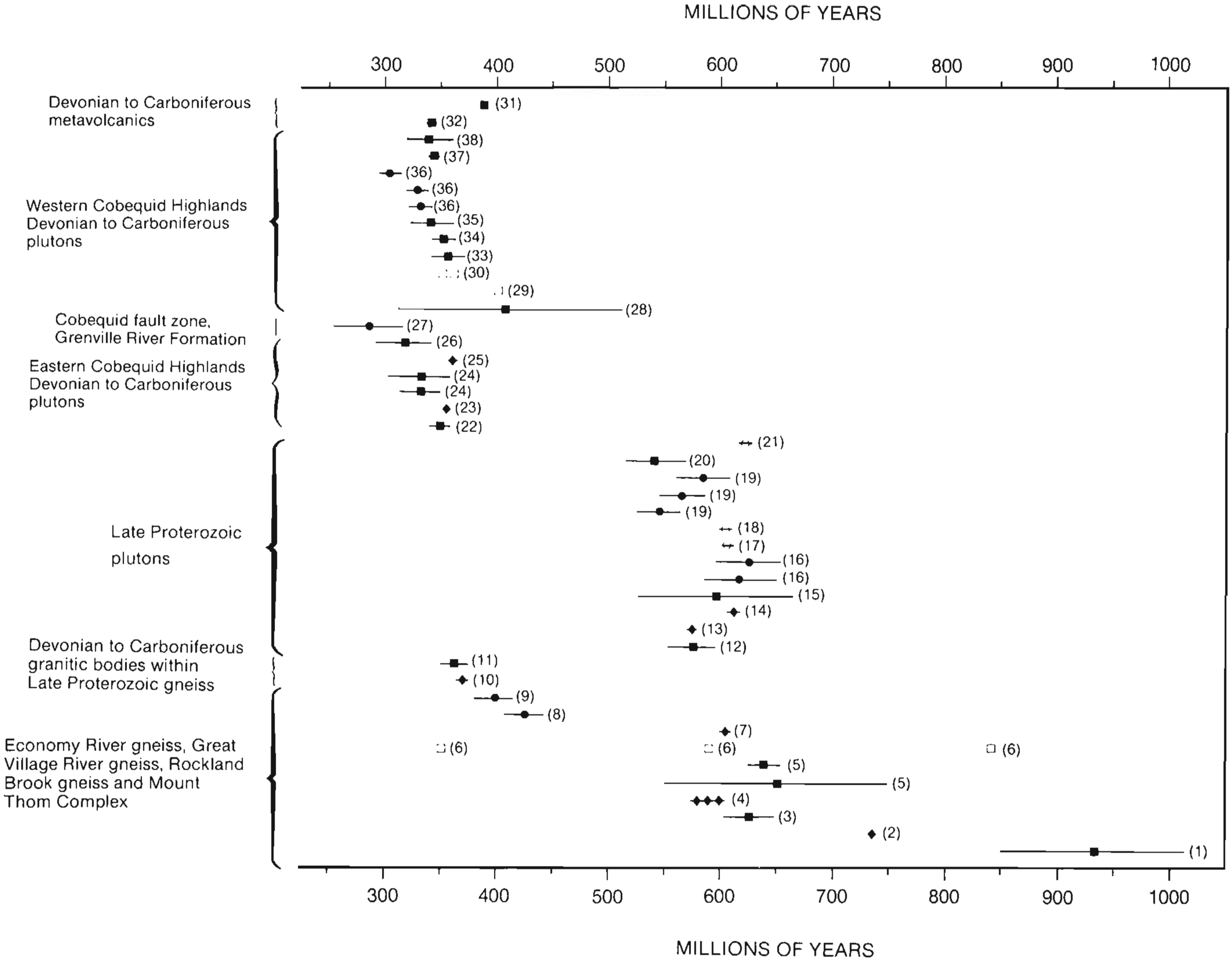

Figure 2. Summary of age data (with error limits) for the Cobequid Highlands. Filled squares are Rb-Sr whole rock/mineral isochron ages, open squares are Rb-Sr ages where no isochron formed, filled diamonds are U-Pb zircon ages, filled circles are K-Ar whole rock/mineral ages and open circles are $^{40}Ar/^{39}Ar$. Numbers in brackets refer to source of data in Table 1.

ca. 630-570 Ma Avalonian orogenic cycle. This data also indicate that the Great Village Gneiss is younger than the Gamble Brook Formation.

Deposition of the Folly River Formation post-dated the deformation associated with the ductile shear zone between the Gamble Brook Formation and the Great Village River Gneiss (Murphy et al. 1988). Rb-Sr whole-rock dates on the syn- to late-tectonic granite gneisses within the shear zone (6, Table 1, Gaudette et al. 1984)) ranged from ca. 839 Ma to 350 Ma. U-Pb data on these rocks yield a 605 +/− 5 Ma age (7, Table 1, Doig et al. 1989, and manuscript in preparation). A minimum age for the Folly River Formation is provided by the post-tectonic intrusion of the Debert River pluton. This pluton yielded an imprecise Rb-Sr date of 596+/− 70 Ma (15, Table 1, Donohoe et al. 1986). U-Pb data yield a 612 +/− 4 Ma age (14, Table 1, Doig et al. 1989 and manuscript in preparation). Together these isotopic data tightly constrain deposition and deformation of the Folly River Formation to about 610 Ma. The close temporal relationship between deposition, deformation and intrusion supports the interpretation that the formation was deposited in a strike-slip tectonic regime (Nance and Murphy, 1988, 1990; Pe-Piper and Murphy, 1989).

Plutonic rocks of the eastern Cobequid Highlands range from granites to appinitic gabbros in composition. Recent $^{40}Ar/^{39}Ar$ (Keppie et al. 1990) and U-Pb (Doig et al. 1989 and manuscript in preparation) data indicate that the plutons range from ca. 622 Ma (Frog Lake Pluton, 21, Table 1) to ca. 575 Ma (McCallum Settlement Pluton, 12, Table 1, Doig et al., manuscript in preparation). The Frog Lake pluton intrudes the Gamble Brook Formation but its age relative to the Folly River Formation cannot be determined on the basis of field relationships.

Table 1. References and sources of age data for geochronological compilation of the Cobequid Highlands.

EASTERN COBEQUID HIGHLANDS

Location/Unit/Rock Type	Age (Ma)	Technique	Reference
1. Mount Thom orthogneiss	934 ± 82 (0.7113)	Rb-Sr (W)	Gaudette et al. 1984
2. Economy River orthogneiss	734 ± 3	U-Pb (Z)	Doig et al. 1989 and 1990a
3. Great Village granite gneiss	626 ± 22 (0.7057)	Rb-Sr (W)	Gaudette et al. 1984
4. Great Village River gneiss	600 ± 5 580 ± 5 589 ± 5	U-Pb (Z)	Doig et al. in prep.
5. Rockland Brook granite-gneiss	652 ± 99 (0.7051) 642 ± 15 (0.7053)	Rb-Sr (W)	Gaudette et al. 1984
6. Portapique River granite-gneiss	*839 (0.7003) *587 (0.7046) *350 (0.7088)	Rb-Sr (W)	Gaudette et al. 1984
7. Great Village River gneissic granodiorite	605 ± 5	U-Pb (Z)	Doig et al. 1989; Doig in prep.
8. Great Village River gneiss	425 ± 17	K-Ar (m)	Wanless 1972; Keppie and Smith 1978
9. Great Village River gneiss	398 ± 17	K-Ar (b)	Wanless 1972; Keppie and Smith 1978
10. Sheared syenogranite in Great Village River gneiss	370 ± 4	U-Pb (Z)	Doig et al 1989
11. Granitic body within Economy River gneiss.	363 ± 12 (0.7063)	Rb-Sr (W)	Gaudette et al. 1984
12. McCallum Settlement pluton (granite)	575 ± 22 (0.7065)	Rb-Sr (W)	Gaudette et al. 1984
13. McCallum Settlement pluton (granite)	575 ± 5	U-Pb (Z)	Doig et al. in prep.
14. Debert River pluton	612 ± 4	U-Pb (Z)	Doig et al. 1989 and in prep.
15. Debert River granite	596 ± 70 (0.7059)	Rb-Sr (W)	Donohoe et al. 1986; Donohoe and Wallace 1985
16. Jeffers Brook diorite	616 ± 28 626 ± 28	K-Ar (h)	Wanless et al. 1973; Keppie and Smith 1978.
17. Jeffers Brook diorite	607 ± 3	$^{40}Ar/^{39}Ar$ (h)	Keppie et al. 1990
18. Jeffers Brook diorite	605 ± 4	$^{40}Ar/^{39}Ar$ (h)	Keppie et al. 1990
19. Jeffers Brook diorite	544 ± 22 564 ± 22 585 ± 23	K-Ar (b)	Wanless et al. 1973; Keppie and Smith 1978
20. Jeffers Brook diorite	541 ± 25 (0.7060)	Rb-Sr (b)	Cormier 1979
21. Frog Lake diorite	622 ± 3	$^{40}Ar/^{39}Ar$ (h)	Keppie et al. 1990
22. Hart Lake-Byers Lake granite	348 ± 9 (0.7046)	Rb-Sr (W)	Donohoe et al. 1986
23. Hart Lake-Byers Lake granite	361 ± 2	U-Pb (Z)	Doig et al. 1990
24. Hart Lake-Byers Lake granite	331 ± 17 (0.7076) 331 ± 27 (0.7097)	Rb-Sr (W)	Cormier 1979
25. Pleasant Hills granite	315 ± 25	U-Pb (Z)	Doig et al. 1990
26. Pleasant Hills granite	315 ± 25 (0.7082)	Rb-Sr (W/k/q-p)	Cormier 1980
27. Fault Zone, Grenville River Formation	287 ± 34	K-Ar (W)	Wanless et al. 1967
28. Davidson Brook granite	409 ± 87 (N/A)	Rb-Sr (W)	Pe-Piper et al. 1989
29. Gilbert-Wyvren (East) granite	*400	Rb-Sr (W/b)	Pe-Piper et al. 1989
30. Gilbert-Wyvren (West) granite	*362 *353	Rb-Sr (W/k/b)	Pe-Piper et al. 1989
31. Fountain Lake (old) volcanics	**387 ± 2 (0.7075)	Rb-Sr (W)	Cormier 1982 and Pe-Piper et al. 1989.
32. Fountain Lake (young) volcanics	**341 ± 4 (0.7075)	Rb-Sr (W)	Cormier 1982 and Pe-Piper et al. 1989.
33. North River granite	**356 ± 17 (0.7049)	Rb-Sr (W/k)	Pe-Piper et al. 1989
34. Cape Chignecto granite (diorite)	349 ± 12 (0.7052)	Rb-Sr (W/k)	Pe-Piper et al. 1989
35. Cape Chignecto granite	339 ± 22 (0.7064)	Rb-Sr (W)	Cormier 1979
36. Cape Chignecto deformed granite-diorite phases	329 ± 9 327 ± 11 303 ± 11	K-Ar (b)	Waldron et al. 1989
37. West Mooose Rive granite (gabbro-diorite)	342 ± 5 (0.7084)	Rb-Sr (W/k)	Pe-Piper et al. 1989
38. Hanna Farm granite	338 ± 17 (0.7069)	Rb-Sr (W/k)	Pe-Piper et al. 1989

(Sample types: W - whole rock; k - K feldspar; q-p - quartz-plagioclase; b - biotite; m - muscovite; h - hornlende; Z - zircon)

* - Rb-Sr isochron not formed (minimum ages)
** - Rb-Sr errorchron

Precambrian rocks in the western Cobequid Highlands have been studied in detail by Pe-Piper and Piper (1987, 1989). The Jeffers Group consists of arc-related interlayered mafic to felsic volcanic rocks and turbidites (Pe-Piper and Piper 1987, 1989). A minimum age for the deposition and deformation of the group is provided by the age of the Jeffers Brook pluton, which post-tectonically intrudes the volcanic and turbidite sequences. K-Ar and Rb-Sr data for the diorite pluton (16, 19, 20 Table 1) produced ages ranging from ca. 626 +/− 28 Ma to 541 +/− 25 Ma. More precise $^{40}Ar/^{39}Ar$ age determinations on hornblendes (17, 18 Table 1, Keppie et al. 1990) suggest that the age of the diorite is ca. 605 Ma. A maximum age for the deposition of the Jeffers Group cannot, at present, be determined by geochronological data. However, on the basis of regional stratigraphic relationships, the Jeffers Group is interpreted to correlate with the Folly River Formation (Murphy et al., manuscript in preparation).

PALEOZOIC ROCKS

Cambrian to Lower Ordovician rocks are not exposed in the Cobequid Highlands. Silurian to middle Devonian rocks predominantly consist of shallow marine to subaerial siliciclastic rocks. The depositional ages of these rocks are reliably inferred from fossil data (Donohoe and Wallace, 1982, 1985). Upper Devonian to Upper Carboniferous rocks unconformably overlie middle Devonian rocks (Donohoe and Wallace 1982; Murphy et al. 1988). To the north of the Rockland Brook Fault they consist of bimodal volcanic rocks and minor interlayered clastic sediments of the Fountain Lake Group. Rb-Sr data from drillholes within the Fountain Lake Group yield a well constrained isochron of 341 +/− 4 Ma (Pe-Piper et al. 1989a) and a suggestion that some of these volcanics may be as old as 387 +/− 2 Ma (31, 32, Table 1, Pe-Piper et al. 1989a). This supports the drillhole data of Chatterjee (1984) that suggest that an unconformity may exist within the group. However, at present, field relationships are equivocal. These rocks are currently being investigated by U-Pb dating techniques. Preliminary data are not encouraging and indicate that most of the zircon crystals are xenocrysts and are more likely to give the age of the basement rocks than the age of crystallization of the Fountain Lake rhyolites (Doig et al., 1990). To the south of the Rockland Brook Fault, Upper Devonian to Upper Carboniferous rocks are predominantly clastic sediments, many of which are reliably dated by paleontological data (Donohoe and Wallace, 1982).

Paleozoic plutonic rocks are widespread within the Cobequid Highlands. These are predominantly bimodal in composition (gabbro and granite); intermediate compositions being attributed to mixing between gabbroic and granitic magmas (Pe-Piper et al., 1989b). The granitic rocks display intraplate characteristics and are thought to be related to intra-continental rifting (Pe-Piper et al. 1989b). K-Ar and Rb-Sr age data range from 409 +/ − 87 Ma (Davidson Brook granite, 28, Table 1, Pe-Piper et al. 1989a) to 315 +/ − 25 Ma (Pleasant Hills Pluton, 26, Table 1, Cormier, 1980); with many clustering around 340 to 360 Ma. U-Pb investigations of these plutons is

Table 2. Stratigraphy of Precambrian to Lower Carboniferous rocks in the Cobequid Highlands (modified after Donohoe and Wallace, 1982 and Murphy et al., 1988). (F — dated by fossils).

NORTH OF ROCKLAND BROOK FAULT		SOUTH OF ROCKLAND BROOK FAULT	
STRATIGRAPHY	PLUTONS	STRATIGRAPHY	PLUTONS
	(a) gabbro, (b) diorite, (c) granite		(a) gabbro, (b) diorite, (c) granite
Fountain Lake Group (a) mafic (b) felsic		Nuttby Formation	
----------unconformity	-------------------------------	unconformity-------------	
Portapique River and Murphy Brook Formations			
Wilson Brook (F) Formation		Sediments (F) (Undivided)	
-----------unconformity	-------------------------------	unconformity------------	
(a) Jeffers Group (ca. 610 Ma), (b) Warwick Mountain Formation, (c) Dalhousie Mountain Formation.	Diorite-granite (ca. 605 Ma)	Folly River Formation (ca. 610 Ma)	Granite-diorite-gneiss (ca. 625-575 Ma), Great Village River gneiss. (ca. 600-580 Ma)
			Economy River Gneiss (734 Ma)
		Gamble Brook Formation >734 Ma	
		<--- Mount Thom Complex ? --->	

the subject of ongoing research. A pilot study was initiated to ascertain the reliability of Rb-Sr age data. Two plutons with well constrained Rb-Sr isochrons were selected; the Hart Lake-Byers Lake and Pleasant Hills plutons, which have Rb-Sr ages of 348 +/− 9 Ma (22, Table 1, Donohoe et al. 1986), and 315 +/− 25 Ma (26, Table 1, Cormier 1980) respectively. Zircons from these plutons show a degree of inheritance and yield 363 +/−3 and 361 +/− 2 Ma ages respectively. These ages compare favourably with the 370 +/− 4 Ma age for a dyke of syenogranite that intrudes the ca. 734 Ma Economy River Gneiss (10, Table 1, Doig et al. 1989). These preliminary data suggest that Devono-Carboniferous plutonism in the Cobequid Highlands may span a much narrower time interval than was hitherto realized.

CONCLUSIONS

K-Ar and Rb-Sr geochronological data have provided a broad framework for the geological history of the Cobequid Highlands. However, recent U-Pb and $^{40}Ar/^{39}Ar$ age data provide new insights into the timing of magmatism and deformation, thus placing constraints on tectonic models. The most significant results are summarized as follows:

1: The Great Village River Gneiss was emplaced between ca. 580 and 600 Ma. Therefore it should be considered part of the ca. 630-570 Ma Avalonian orogenic cycle and is not basement to the Avalon terrane as it was previously interpreted.

2: The 734 +/− 2 Ma age of the Economy River Gneiss provides a minimum age for the Gamble Brook Formation and confirms the antiquity of the formation relative to the Avalonian orogenic cycle.

3: The age of the basement to the Avalon terrane in the Cobequid Highlands is unknown. The Mount Thom complex is a possible candidate although its age cannot, at present, be established using U-Pb techniques.

4: A U-Pb pilot study into the age of Devono-Carboniferous plutons suggest that they may not have as wide an age range as previous Rb-Sr data would suggest.

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Syn- and post-kinematic intrusions of gabbro and peridotite into layered gabbroic cumulates in the Bay of Islands ophiolite Newfoundland: genesis of anorthosite by reaction, and troctolite by hybridization

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Abstract

The layered mafic cumulate unit of the North Arm Mountain massif is composed of early, isoclinally-folded, leuco- to mela-olivine gabbro, gabbro, leuco- to mela-troctolite and anorthosite; with late-kinematic, intra-cumulate intrusions of wehrlite, feldspathic wehrlite, troctolitic gabbro, olivine micro-gabbro and gabbro-norite. Most rocks are deformed. Some peridotite intrusions have sharp, discordant contacts, others have anorthositic reaction rims or hybrid margins. Many layered troctolitic and melagabbro sequences include lenses of feldspathic wehrlite. These peridotites commonly contain rafts and clots of troctolite or gabbro, suggesting that the troctolitic and melagabbro layered sequences formed when primitive replenishments intruded as near-concordant, intra-cumulate sills. The textures suggest that the replenishing magmas mixed with loosely consolidated feldspathic cumulates to yield hybrid troctolites and olivine gabbros. The Bay of Islands layered mafic cumulates did not form in an open, replenished chamber, but as multiple, intra-cumulate, syn-kinematic intrusions.

Résumé

L'unité litée mafique du massif de North Arm est composée de séquences précoces plissées, de gabbro à olivine et de troctolite (leuco- à méla-), de gabbro et d'anorthosite. Ces séquences plissées sont recoupées par des filons et filons-couches tardi-cinématiques, intra-cumulatifs, composés de werhlite et de werhlite felspathique, de gabbro troctolitique, de micro-gabbro à olivine et de gabbro-norite. La plupart des roches sont déformées. Plusieurs intrusions de péridotite ont des bordures réactionnelles anorthositiques, ou des faciès hybrides de contact. D'autres ont des contacts intrusifs francs. Plusieurs séquences troctolitiques et mélagabbroïques contiennent des lentilles de péridotite feldspathique, qui elles contiennent des xénolithes anguleux de troctolite et de gabbro. Les textures impliquent que ces péridotites représentent des filons-couches intra-cumulatifs de magma primitif. Suite à l'injection, les cumulats felspathiques hôtes se sont désagrégés, donnant lieu à des troctolites et gabbros hybrides. Les cumulats mafiques stratifiés de Bay of Islands ne se sont pas formés aux parois d'une grande chambre magmatique ouverte aux injections de magmas primaires, mais reflètent plutôt une série d'intrusions syn-cinématiques, injectées dans des cumulats partiellement consolidés.

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INTRODUCTION

This paper reports on an investigation of the layered gabbros from the North Arm Mountain massif of the Bay of Islands ophiolite (Fig.1). We examine the structures and contact relationships characterizing the layered mafic cumulates (unit iii) that overlie the mixed mafic-ultramafic cumulates (unit ii) of the lower crust (Fig.1). Our objective is to clarify the nature of the magma chamber processes responsible for these rocks, with the hope that the insights gained thereby will help in understanding the processes by which the oceanic crust forms at spreading ridges. The North Arm massif layered gabbros have previously been interpreted as in-situ cumulates precipitated on the floor and walls of a large, cylindrically-shaped, long-lived, periodically replenished magma chamber (Komor et al., 1987; Komor and Elthon, 1990; Casey et al., 1983). Our results are inconsistent with the existing model, requiring instead a more complex scenario of multiple, syn-kinematic, intra-cumulate intrusions.

GEOLOGICAL CONTEXT

The Bay of Islands ophiolite was obducted onto the North American platform around 469 ±5 Ma ($^{40}Ar/^{39}Ar$ Dallmeyer and Williams, 1975), at least 10 Ma after the plutonic rocks had completely solidified (487 +1.9/−1.2 U-Pb age on trondhjemites by Dunning and Krogh, 1985). During obduction, the ophiolite was tilted and broken up by steeply-dipping faults, exposing a near-continuous cross-section through the crust (Fig.1). These steeply-dipping faults appear to be the only obduction-related structures developed; all other structures are considered to be ridge-related.

STRUCTURE

Most of the rocks have ubiquitous isoclinal folds (Fig. 2, 4, 9), boudins (Fig. 5, 6), open folds and incipient axial plane thrusts (Fig.7), transposed and repeated contacts (Fig.9), and the development of mineral foliations and stretching lineations (Fig.3: c.f. Calon et al., 1988). Mineral foliations are typically parallel to lithological layering (Fig.3a), except in fold noses, where the mineral foliations transect layering. Fold closures in the layered gabbros have 1-20 mm wavelengths (Fig. 2, 4, 9). Similar folds in the mixed mafic-ultramafic cumulates underneath have wavelengths up to 50m (Bédard, 1990). In the layered gabbros, dykes show every gradation from apparently undeformed examples which are discordant to layering and mineral foliations (Fig. 8, 10), to strongly transposed and folded equivalents that have acquired the dominant layering orientation (e.g., compare Fig. 8 and 9). This demonstrates that the magmatism was syn-kinematic (Casey 1980; Calon et al., 1988). With progressive deformation, the more feldspathic facies tend to break up into complexly-folded boudins (Fig. 5, 6).

Late mylonitic shear zones transect the layering at various angles, or follow specific horizons within the layered sequence. The presence of shear zones suggests that deformation was increasingly focused into discrete high-strain zones as time went on. In places, structures resembling incipient axial plane thrusts are observed (Fig.7). Intra-cumulate thrust imbrication provides a mechanism for tectonic repetition of right-way-up sequences.

The question that remains to be answered concerns the origin of, and the mechanisms responsible for deformation of the gabbroic cumulate rocks.

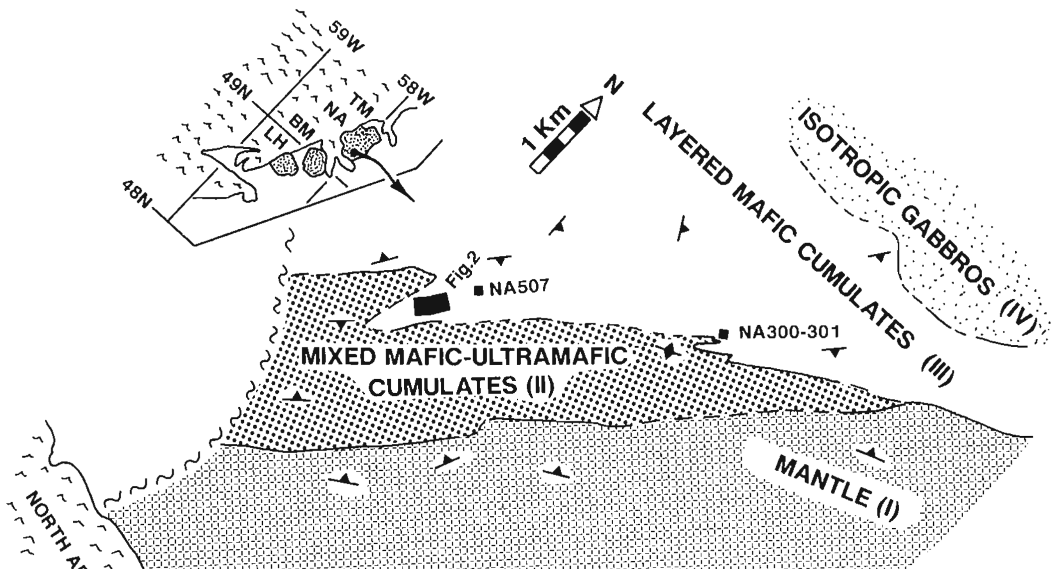

Figure 1. Generalized map (adapted from Casey, 1980; Riccio, 1976) of the study area. NA300-301 is the location of the detailed traverse of Komor and Elthon (1990), NA507 that of Komor et al. (1987). Inset: location map of Bay of Islands Complex, western Newfoundland, NA = North Arm Mountain, TM = Table Mountain, LH = Lewis Hills, BM = Blow-me-Down.

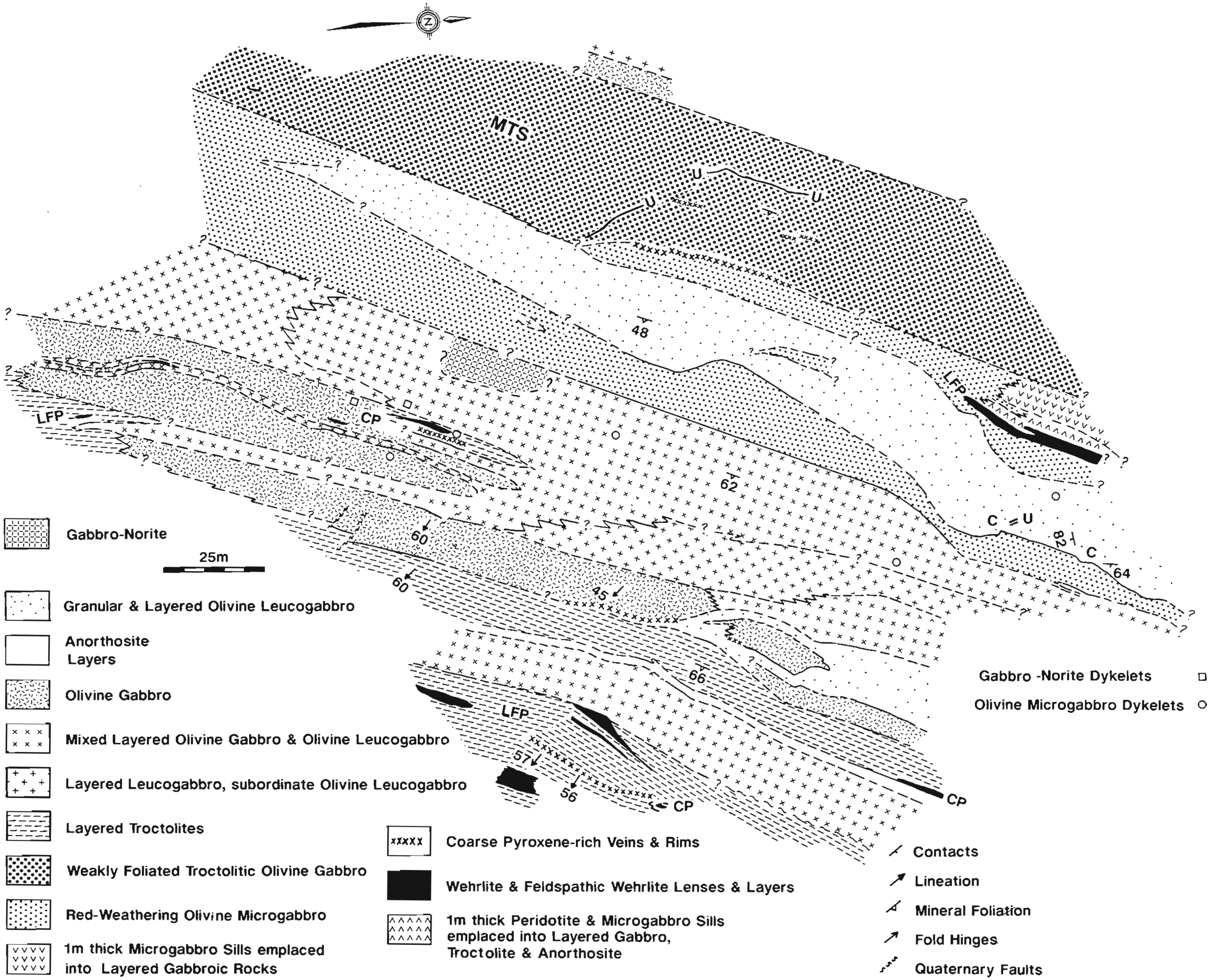

Figure 2. Simplified detail map of layered gabbroic cumulates. See Figure 1 for location. Mapping was done on a chained grid at 1/250 scale. "U" are peridotite dykes, "CP" are wehrlitic lenses within layered troctolites or gabbros, while "LFP" are leopard-textured feldspathic peridotite lenses within layered rocks. "C" are the contacts of an olivine microgabbro sill.

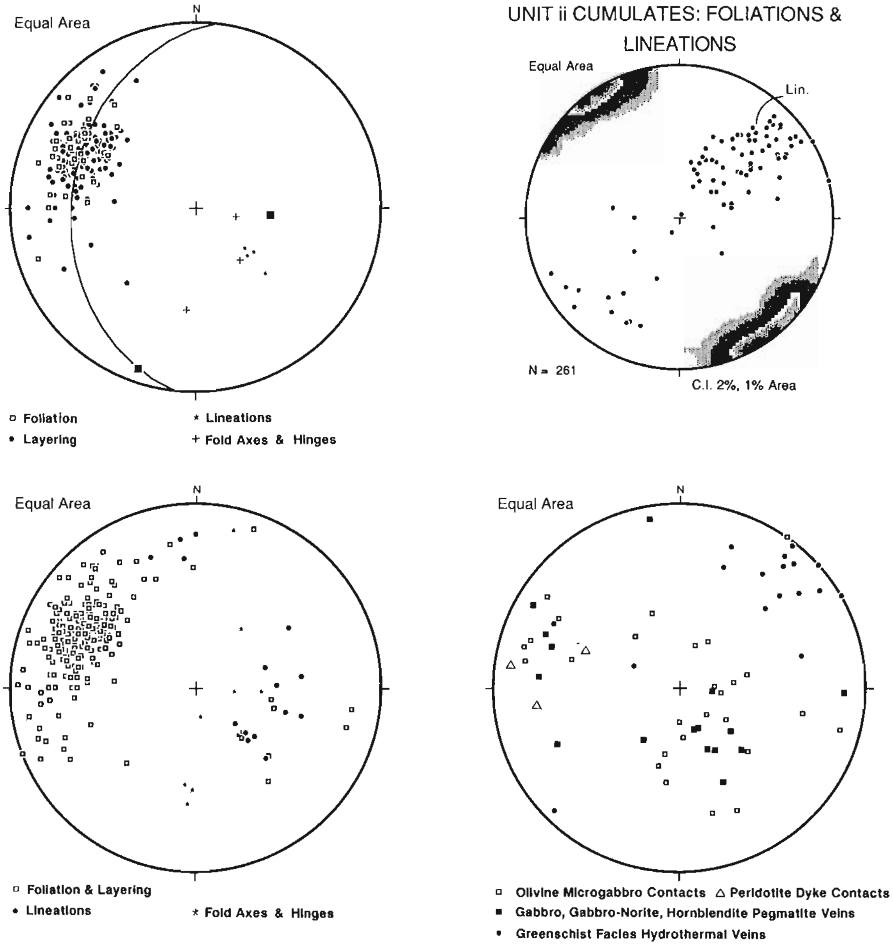

Figure 3. Equal Area projections of structural data plotted using STEREONET, by R.W. Allmendinger.
a) Poles of mineral foliation and layering, mineral lineations, hinges and fold axes from the detailed grid of Figure 2.
b) Poles of mineral foliation and layering, mineral lineations, hinges and fold axes from all layered gabbros above the last thick ultramafic intrusion (includes data from Fig. 3a).
c) Foliations and layering contours and lineations in the underlying mixed mafic-ultramafic cumulates (unit ii). Mantle rocks have similar structural orientations. Adapted from Bédard (1990).
d) Contacts of late, peridotite and olivine microgabbro sills and associated dykelets, gabbro and hornblende pegmatite veins, and hydrothermal alteration veins. Data from the detailed grid of Figure 2 only.

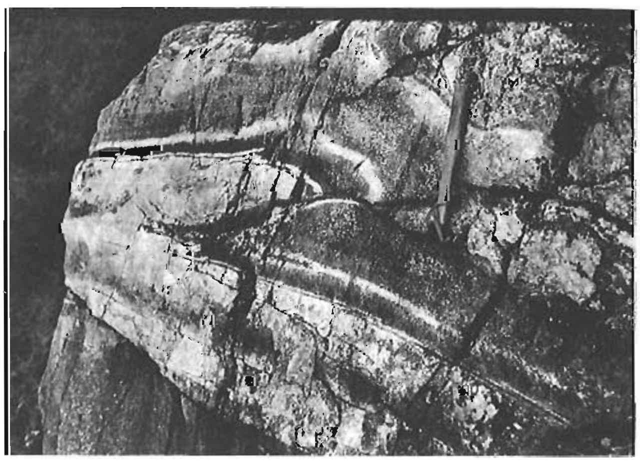

Figure 4. Small-scale fold closure in anorthositic interlayers. NA301 section of Komor and Elthon (1990).

Figure 5. Anorthositic boudins.

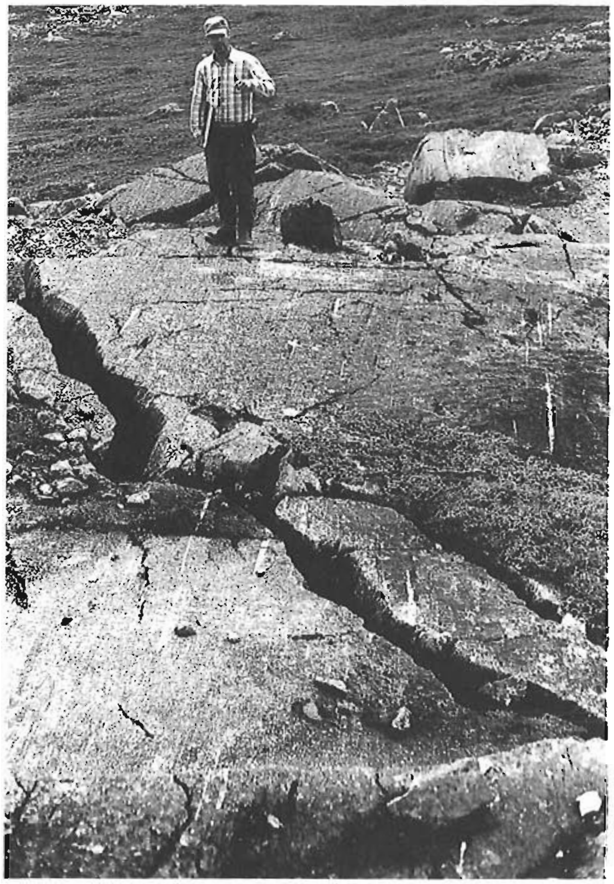

Figure 6. Weakly layered olivine gabbro cumulates containing numerous anorthosite boudins.

Figure 7. Open fold with axial shear-zone, interpreted to be an incipient décollement. Upper part of outcrop NA301 of Komor and Elthon (1990).

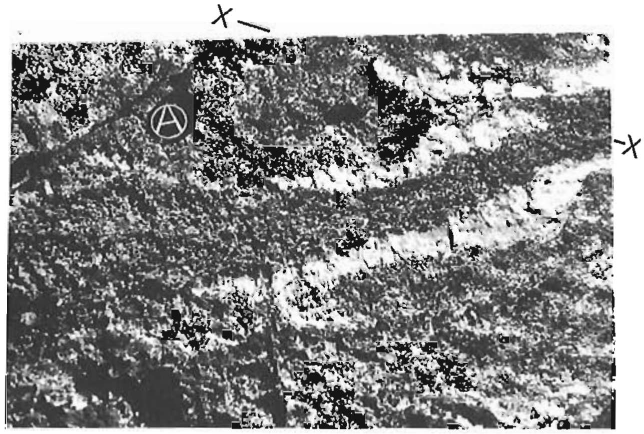

Figure 8. Feldspathic peridotite dyke with prominent anorthositic reaction rims. X-X is the foliation/layering orientation in the host olivine gabbros.

Figure 9. Folded equivalent of Figure 6.

Figure 10. Dykelet of pyroxene-phyric olivine microgabbro near the contact of a larger microgabbro sill.

LITHOLOGIES

A) *Layered olivine gabbros.* This facies consists of thickly-layered (0.5-1m), foliated and lineated olivine gabbros. The proportions of the three phases vary widely. Some sections are relatively massive, with thickly-bedded, fairly homogeneous olivine gabbro. Other sections are more heterogeneous, with finer-scale alternations of olivine gabbro, leucogabbro to leucotroctolite, anorthosite and troctolite (Fig.6). Some laminae or layers can be followed for 10s to 100s of metres; but more typically, individual layers branch out or terminate in isoclinal fold closures along-strike. Graded "depositional" beds have not yet been observed.

B) *Layered troctolites.* The layered troctolite facies is composed of foliated and lineated medium grained troctolite and melatroctolite layers (typically 0.5-1m thick), with subordinate interlayers of olivine gabbro and internally folded anorthositic boudins (Fig. 5). Feldspar in the melatroctolites generally occurs as anorthositic clots (Fig.11). The thicker troctolite sequences commonly contain lenses of wehrlite (CP on Fig.2) or leopard-textured feldspathic wehrlite (LFP on Fig.2) and melatroctolite (Fig.11).

C) *Leucogabbros.* Layered sequences 1-25 m thick are dominated by foliated leucogabbro, olivine leucogabbro, and leucotroctolite. Isoclinally folded anorthositic schlieren and boudins are common. Interlayers of gabbro and olivine gabbro are found along-strike from lenses of more massive olivine gabbro. The leucogabbros are generally more altered than are the other lithologies, with closely-spaced swarms of greenschist-facies veins, or of hornblende-gabbro or gabbro-norite pegmatite. Locally, gabbro-norite lenses are found.

D) *Massive troctolitic gabbro sill.* A 20m-thick tabular body of troctolitic gabbro (MTS on Fig.2) is almost massive (no layering was observed), and is only weakly foliated. This is in marked contrast with the surrounding lithologies, which are very heterogeneous, extensively veined and altered, and exhibit prominent deformation-induced mineral foliations. The massive troctolitic gabbro body is interpreted to represent a late-kinematic sill. It is cut by wehrlite dykes, and by "veinlets" characterized by coarse grained clinopyroxene. These veinlets appear to be related to the olivine microgabbro sills. Confirmation of these inferred relationships awaits geochemical data.

E) *Ultramafic dykes and anorthositic rims.* Narrow (0.2-1m), discordant dykes of wehrlite to feldspathic wehrlite occur sparsely throughout the lower part of the layered gabbro unit (U on Fig.2, Fig.8). Some ultramafic dykes have sharp, fine grained contacts. However, in many cases the contacts are lined either by leopard-textured rocks, or by anorthosite. Veins of anorthosite may emanate from the main anorthosite rims into the host rocks. The occurrence and morphology of the anorthosite (Fig. 8) suggests that they formed as reaction rims between the peridotite intrusions and the host gabbros. Some ultramafic dykes and their anorthosite rims have been isoclinally folded (Fig. 4, 9), producing pseudo-rhythmically layered sequences. The common anorthosite boudins (Fig. 5, 6) may represent disrupted anorthosite reaction rims.

F) *Leopard-textured melatroctolites, melagabbros and feldspathic wehrlites.* Leopard-textured facies commonly occur as small patches at the contacts of discordant peridotite dykes, or as thicker (0.2-3m) concordant layers or zones within the layered troctolites and olivine gabbros (LFP on Fig.2; Fig. 12, 13, 14). Plagioclase is typically rounded or scalloped and much of it occurs as clots of gabbro or anorthosite (Fig.11). Cr-spinel is commonly found as an accessory, either as discontinuous rims around feldspar grains, as discontinuous rims around anorthositic lenses, or as disseminations in the peridotite matrix.

Where deformation is not too severe, outcrop-scale textures and structures are diagnostic of intrusion by olivine + spinel ± clinopyroxene-saturated magmas into, and incomplete hybridization with poorly-consolidated gabbroic cumulate hosts (Fig. 11-14). In one place, the outcrop-scale textures strongly suggest that the anorthositic schlieren in the peridotite (Fig.11) originated through marginal disaggregation of a 1.5m thick raft of leuco-

Figure 11. Partially disaggregated schlieren and clots of anorthositic troctolite in a hybrid melatroctolite.

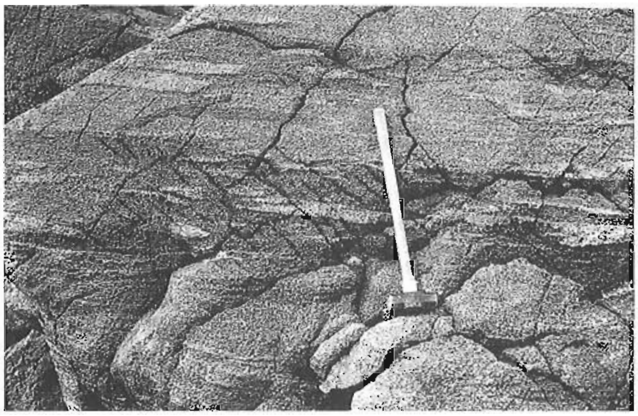

Figure 12. Hybrid troctolites. Note the angular troctolitic rafts at the top left of the photo, and the ovoid, irregularly-shaped troctolitic clots near the hammer handle (handle is 0.8m long).

Figure 13. Brecciated rafts of leucogabbro in a feldspathic peridotite matrix.

Figure 14. Detail of Figure 13. Note that the peridotite-gabbro contact is discordant to layering in the raft.

olivine gabbro. In other places, layered troctolite, leucogabbro and anorthosite occur as brecciated, angular rafts in a heterogeneous medium- to coarse-grained, leopard-textured feldspathic wehrlite or melatroctolite (Fig. 12-14). The internal layering in the rafts is clearly truncated by the peridotite (Fig. 13, 14), implying that the peridotite is intrusive. The rafts probably represent layers that were too well consolidated to mix with the invading magmas.

G) *Olivine microgabbro and gabbro-norite intrusions.* Sill-swarms of massive to weakly foliated, red-weathering olivine microgabbro (Fig.2) have been emplaced into the layered gabbroic cumulates. Individual microgabbro sills can be up to 15m thick, but closely-spaced swarms of thinner (< 1m) sills also occur. Sill contacts are typically sharp, with 1-2m jogs that may reflect control by pre-existing joints (C on Fig.2). Schlieren of anorthositic and coarse grained gabbroic rocks occur within the microgabbroic intrusions, especially near the contacts. These coarser schlieren are interpreted to be enclaves of the host rocks. Thin (1-5cm) dykelets of microgabbro (Fig.10) are composed either of pyroxene-phyric olivine microgabbro that is very similar to the thicker microgabbro bodies, or of gabbro-norite (± amphibole and oxides). Pyroxenite pegmatites may develop at the contacts of microgabbro dykelets. Gabbroic and hornblende-gabbro to gabbro-norite pegmatite veins have the same orientation as the dykelets (Fig.3) and may also belong to the microgabbro suite.

H) *Pyroxenites.* Thin (30cm) layers of granular clinopyroxene-rich gabbro are observed locally. They are laterally discontinuous, but are only weakly foliated. Their origin is problematic.

I) *Hydrothermal-mylonitic shear zones.* Hydrothermally altered greenschist to amphibolite facies metagabbros (200-300m domains) are developed around mylonitic shear zones (0.1-3m wide) throughout the layered gabbro unit (Rosencrantz, 1983). These arcuate hydrothermal alteration/shear zones dissect the layered gabbroic cumulates into kilometre-scale domains. The shear zones appear to bottom out at the uppermost thick ultramafic layer (Bédard, 1990), and extend up as breccia zones into the overlying sheeted dykes and lavas (Rosencrantz, 1983). The shear zones also localized the emplacement of thin feldspathic peridotite intrusions, the development of hornblende-plagioclase, pyroxenitic or gabbroic pegmatites (Fig.14 in Bédard, 1990), and the deposition of sulphides (up to 40 vol%). The association of mylonitic shear zones and hydrothermal alteration suggests that seawater migrated down into major crustal faults, all the way to the gabbro/peridotite transition.

ORIGIN AND NATURE OF DEFORMATION IN THE LAYERED GABBROS

In the detail map area (Fig.2), fold axes and lineations are tightly clustered (Fig.3a), suggesting a fairly uniform stress field. This uniformity, the fact that the mineral foliation transects lithological layering in hinge zones, rather than wrapping around the hinges, and the ubiquitous boudinage of anorthositic rocks, together imply that the deformation is not solely the result of viscous, syndepositional slumping as proposed by Komor et al. (1987), but also involves plastic deformation in response to an applied shear stress.

On the outcrop scale, the structures seen in the layered gabbroic cumulates (unit iii) are virtually identical to those of the underlying ultramafic cumulates and mantle tectonites (Bédard, 1990). In the ultramafic rocks of units i and ii (Fig.3c), the highest concentration of stretching lineations (L2) is normal to the plane of the sheeted dykes (Casey, 1980; Rosencrantz, 1983). The deformation seen in the ultramafic rocks has been interpreted Casey, 1980; Girardeau and Nicolas, 1981; Nicolas and Violette, 1982; Calon et al., 1988; Bédard, 1990) to be the result of subhorizontal asthenospheric flow away from the spreading ridge. Does the deformation of the gabbric cumulates of unit iii have the same origin?

Poles to mineral foliations and layering from all of the layered gabbros are plotted together with lineations, hinges and fold axes on Figure 3b. Comparison of Figures 3b and 3c shows that the foliations and layering orientations in the layered gabbros (unit iii) are distinct from that of the underlying ultramafic cumulates and mantle tectonites. The orientation of mineral stretching lineations, hinges and fold axes from unit iii show considerable scatter (Fig.3b), but most plunge steeply to the east, with a smaller number plunging shallowly to the north. Lineations in the underlying units (units i and ii) plunge mostly northeast (Fig.3c). The structures in the layered mafic cumulates are not parallel to those of the underlying units. There are at least three potential explanations for this:

1) It is due to an unrecognized obduction-related fault between the two domains. In the study area, the outcrop along Liverpool Brook is poor, and so this hypothesis cannot be evaluated.

2) The layered mafic cumulates may have been deformed at the same time as the underlying rocks in response to shear between the lithosphere and the asthenosphere (e.g. Nicolas et al., 1988). The progressive upward rotation of foliations and changes in the lineation directions might then reflect strain refraction caused by the upwardly-increasing rigidity of the rocks, in conjunction with the approach to the free upper surface.

3) Rosencrantz (1983) suggested that differential subsidence and rotation of a thick magma chamber lid (lavas and sheeted dykes) was controlled by slip along the arcuate shear zones that dissect the layered gabbros. This provides a possible explanation for the non-concordance of lineations and foliations in the gabbroic versus the ultramafic cumulates; differential lid subsidence could have caused deformation of partially consolidated gabbroic cumulates, while ultramafic cumulates were principally affected by shear between the asthenosphere and lithosphere.

In the area mapped in detail (Fig.2), poles to the contacts of post-kinematic intrusions, dykes, and gabbroic pegmatite veins define two distribution maxima at 90° to one another (Fig.3d). The angular contacts of some sills (C on Fig.2) suggest that the orientations of these post-kinematic intrusive contacts are controlled by pre-existing structures, perhaps joints. Late, greenschist-facies, hydrothermal veins define a third cluster that is perpendicular to the other two. The poles to the hydrothermal veins are parallel to the maximum density of lineations in the subjacent mantle and lower crustal rocks (compare Fig. 3c and d). If this parallelism is not fortuitous (i.e. hypothesis 1 is incorrect), then the brittle failure which overprinted the plastic deformation structures of the gabbroic rocks may have been caused by a stress-field similar to that responsible for deformation in the ultramafic cumulates and mantle tectonites. This may reflect propagation of the asthenosphere-lithosphere shear couple up into the largely solidified, previously deformed, gabbroic cumulates of the upper crust. The late intrusive contacts and veins would then reflect control by (a) bedding-joints (roughly paleo-horizontal, perpendicular to the lithostatic load), and (b) a complementary paleo-vertical joint plane. These two orientation clusters would represent the exploitation by late intrusions of a joint set conjugate to the shear couple between the asthenosphere and the lithosphere. The hydrothermal veins would have developed subsequently, when the differential movement of crust and mantle had ceased, allowing the formation of relaxation fractures with an orientation perpendicular to σ_1 (= lineation direction).

FORMATION OF LAYERED GABBROS

Previous model

Casey et al. (1983), Komor et al. (1987) and Komor and Elthon (1990) have proposed that basaltic magmas produced by high pressure fractional crystallization (Elthon et al., 1982, 1984) were injected into large, long-lived, steady-state magma chambers, where extraction of gabbroic cumulates yielded MORB-like residual lavas. The relatively uniform layered gabbros were interpreted by them to represent in-situ precipitates against chamber walls and floor, extracted from well-stirred, convecting magmas; while the rhythmically layered melagabbro-anorthosite sequences were interpreted to represent localized replenishment events in sheltered "re-entrants" at chamber walls.

Re-interpretation.

The results presented above and in Bédard (1990) suggest a different interpretation.

1) Consequences of deformation.

Recognition of the intense deformation that has affected the layered gabbros has profound implications for magma chamber models, since it invalidates the assumption that all rocks accumulated sequentially from the bottom up. For example, Komor et al. (1987, NA507 section on Fig.1) document the existence of reversed and normally zoned cryptic stratigraphies with cycle length-scales of 10-30m. This is close to the scale of the fold closures we have mapped in very similar rocks (Fig.2) less than 1 km along-strike from the NA507 section. We would re-interpret the symmetrical reversed and normally-zoned cryptic cycles of NA507 as tectonic repetitions. Similarly, the 37 replenishment events proposed by Komor and Elthon (1990), are probably tectonic repetitions (Fig. 4 and 7 are from the same outcrop they studied) of a few intra-cumulate replenishment events (cf Fig. 8, 9). The so-called chamber-wall "re-entrant" discussed by Komor and Elthon (1990) is an intrusive peridotite at the core of a kilometre-scale fold, not an original chamber-floor feature.

2) A single replenished chamber or multiple intrusions?

Our detailed lithological mapping documents the existence of numerous and voluminous intrusions of wehrlite, troctolitic gabbro, olivine microgabbro and gabbro-norite. These sills and dykes were emplaced into already-deformed layered gabbroic cumulates. The nature of the layered cumulates that host the intrusions is difficult to determine because of the intense deformation. They may be in situ precipitates (as proposed by Komor et al. (1987)), or they may simply be earlier, more deformed intrusions.

Dykes and subconcordant sills of wehrlite and feldspathic wehrlite crosscut the fabric of layered gabbroic rocks (U on Fig.2). The small peridotite intrusions of unit iii contrasts with the voluminous peridotite intrusions of the underlying unit (ii). This is not surprising, since the unit ii/iii contact is defined as the last thick peridotite layer. One problem with the interpretation of the peridotites of unit ii as intrusions is their cumulate nature, i.e. where have the residual magmas gone to? The abundance of late gabbroic and gabbro-norite intrusions in the layered mafic cumulate unit (Fig.2) suggests that the magmas residual from crystallization of the peridotite intrusions of unit ii were injected up into the overlying gabbroic cumulates of unit iii.

The thicker sequences of layered troctolite and olivine gabbro commonly enclose (< 3m) lenses of wehrlite (CP on Fig.2) and heterogeneous, leopard-textured, feldspathic wehrlite (LFP on Fig.2). The evidence presented above suggests that they are intra-cumulate sills (cf. Bédard et al., 1988). The less deformed examples have textures that suggest a hybrid origin (Fig. 11-14). We propose that the replenishing magmas mixed with loosely consolidated, partially molten host cumulates. It remains to be determined whether the feldspathic clots in the peridotite are restites from anatexis induced by the replenishments, or are cumulus clumps disrupted by the forceful injection of magma into a loosely consolidated cumulate layer, or both. Some of the common layered troctolites and olivine gabbros may represent more deformed, well homogenized hybrids (Fig.6). The anorthositic boudins are particularly useful petrogenetic tracers in this regard, since they are interpreted to be fossil reaction rims produced through partial assimilation reactions, subsequently disrupted by deformation.

3) Origin of leucogabbros and anorthosites.

Anorthosite rims form at some of the peridotite-gabbro contacts (Fig.8). The morphology of these rims suggests an origin as by products of the assimilation of gabbros by intrusive peridotites. A similar origin was proposed for pyroxenite rims in the lower crust by Bédard (1990). With progressive deformation the dykes and their anorthositic rims are repeated tectonically and form pseudo-rhythmically layered sequences (Fig.9, cf. Fig.3a and f in Komor and Elthon, 1990). If deformation proceeds while a significant viscosity contrast remains between the anorthosite rims and the partially molten gabbroic hosts, the more competent anorthosite will break up into isolated boudins (Fig. 5, 6). The thicker, layered leucogabbro sequences may represent more diffuse, yet analogous reaction zones formed around large intracumulate replenishments.

CONCLUSIONS

The lower part of the layered mafic cumulates (unit iii, Fig.1) of the North Arm Mountain massif is strongly affected by ridge-related plastic deformation. The orientation of structural fabrics in the gabbros appears to differ from those of the underlying ultramafic-mafic cumulates and mantle tectonites. If the difference is real, and not the result of obduction-related faulting, then this suggests that the two lithological associations (layered gabbros versus mixed mafic-ultramafic cumulates) were deformed in response to different stresses. Abundant late-kinematic intrusions of wehrlite, troctolitic olivine gabbro, olivine microgabbro and gabbro-norite have been documented. Peridotite intrusions commonly have hybrid or anorthositic rims which may have formed as byproducts of assimilation reactions. Primitive replenishments also appear to have intruded previously deformed cumulate rocks as near concordant sills, where they mixed with partially molten host cumulates to form hybrid troctolites and olivine gabbros. The evidence does not support a model of a single, long-lived, replenished chamber; but rather, seems to record a history of multiple, intra-cumulate, syn-kinematic intrusions.

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Structural relationships between some gold occurrences and fault zones in the Bathurst area, northern New Brunswick[1]

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Abstract

Structural characteristics of gold occurrences have been studied along the Elmtree and Rocky Brook-Millstream fault zones in northern New Brunswick. The Elmtree deposit is located within a small lenticular mafic body intruded within the black phyllites of the Elmtree Formation. Like the surrounding rocks, it displays evidence of both sub-horizontal and sub-vertical tectonic movements and a contractional duplex geometry related to the Elmtree fault is inferred. Gold showings of the Rocky Brook-Millstream fault system occur in various splay-related structural settings along the main fault zone. All ore-related structures are compatible with a dextral faulting event along the Rocky Brook-Millstream fault.

Résumé

Dans le nord-est du Nouveau-Brunswick, les caractéristiques structurales de plusieurs indices aurifères situés en bordure des failles d'Elmtree et de Rocky Brook-Millstream ont été étudiées. Le gisement d'Elmtree est encaissé par une petite lentille de roches intrusives mafiques recoupant les phyllades de la Formation d'Elmtree. Au même titre que les roches encaissantes, le gisement d'Elmtree présente des signes de mouvements tectoniques sub-horizontaux et sub-verticaux. Une géométrie de duplex par contraction, associé à la faille d'Elmtree, est proposée. Les indices d'or du système de faille de Rocky Brook-Millstream sont associés à des zones de cisaillement secondaires localisées en bordure de la principale surface de décollement et sont en outre compatibles avec un mouvement de décrochement dextre le long du système de faille de Rocky Brook-Millstream.

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INTRODUCTION

During the summer of 1990, a Canada-New Brunswick MDA-II project on gold mineralization was initiated in northern New Brunswick. This project, mainly concerned with structural relationships between gold occurrences and fault zones, is part of a large-scale study of gold deposits in the Canadian Appalachians (Dubé, 1990).

Several types of epigenetic gold occurrences have been recognized in the New Brunswick Appalachians by Ruitenberg et al. (1990). They identified the Bathurst area as a locality where quartz-carbonate vein-type deposits occur. The purpose of this paper is to present preliminary results of a structural study of vein-type gold occurrences along the Elmtree and Rocky Brook-Millstream fault zones. Four deposits (Elmtree, Stephens Brook, Bradley Brook, and Rocky Brook Shaft) are described as typical examples.

REGIONAL GEOLOGICAL SETTING

The studied gold occurrences are located between 10 and 25 km northwest of Bathurst, near the Nicholas Denys and Alcida settlements (Fig. 1). The area is underlain by Ordovician to Devonian rocks belonging to the Ordovician Miramichi and Elmtree terranes and post-Ordovician Matapedia cover sequence (Ruitenberg et al., 1990). Recently, van Staal and Fyffe (in press) proposed including the Elmtree Group as a formation within the Middle Ordovician Tetagouche Group and restricting the use of Miramichi Group to designate clastic rocks stratigraphically underlying the Tetagouche Group. Accordingly, Ordovician rocks of the area belong to the Fournier and Tetagouche groups (Fig. 1). Post-Ordovician cover rocks are included within the Silurian Chaleur Group and the Devonian Dalhousie Group (Davis, 1977). All the rocks are intruded by late Devonian felsic stocks (Fig. 1).

The Elmtree and Rocky Brook-Millstream breaks are major fault systems in northern New Brunswick (Ruitenberg et al., 1990). The Elmtree fault (Fig. 1) lies along the contact between the Elmtree Formation to the NW and the Chaleur Group to the SE. This contact is interpreted as an angular unconformity by van Staal and Fyffe (in press) but it is best described as a faulted unconformity (Hoy, 1985). In the Bathurst area, the Rocky Brook-Millstream fault zone (RBMFZ) is divided into a main break and a south branch zones (Philpott, 1987; Fig. 1). It is interpreted as a major transpressional dextral fault zone (Fyffe and Fricker, 1987; Ruitenberg et al., 1990; van Staal and Langton, 1988).

GOLD ALONG THE ELMTREE FAULT

The Elmtree deposit (site 1 on Fig. 1), held by Corona Corporation, is a hydrothermally altered and mineralized gabbroic body occuring within the black slates and

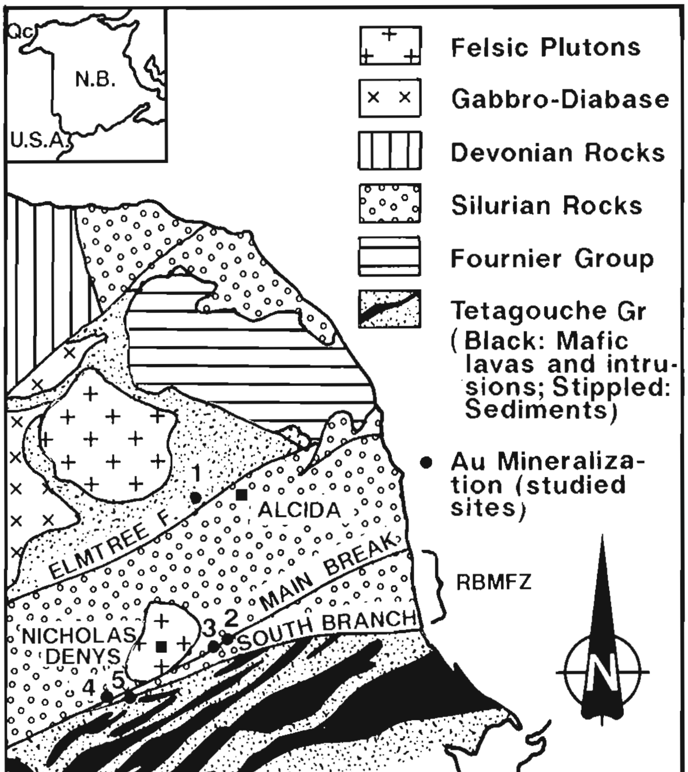


Figure 1. Regional geology of the northwestern part of the Bathurst area. Numbers refer to visited Au occurrences of this study. After Davies, 1977; and van Staal and Fyffe (in press).

phyllites of the Elmtree Formation, north of the Elmtree fault (Fig. 2). It is the most significant gold-only deposit in the studied area. The drill-indicated geological reserve of the deposit is estimated to be 350 000 tons grading 4.46 Au g/t (Dubé, 1990).

An interpretative geological map of the Elmtree deposit area is shown in Figure 2. The Elmtree fault is not clearly exposed in this area but can be seen approximately 1.5 km east of Alcida at the Elmtree Formation and Chaleur Group contact (Fig. 3). The dominant structural feature of this outcrop is a metre-wide brittle-ductile shear zone along the Elmtree-Chaleur contact. Kinematic indicators such as C-S fabrics, dragging of bedding and/or schistosity of both units, and marker offsets indicate a dextral strike-slip movement. Synthetic low-angle and parallel shears and micro-faults are compatible with the kinematic interpretation of the main décollement surface. High-angle sinistral brittle faults (Fig. 3), essentially developed within the Chaleur Group as a result of a high competency contrast with the Elmtree Formation, are interpreted as antithetic fractures (R'-type of Freund, 1974) associated with the Elmtree fault.

North of the Elmtree deposit (Fig. 2), rocks of the Elmtree Formation enclose a small body consisting of greywackes, conglomerates and ultramafic rocks. Van Staal and Fyffe (in press) tentatively interpreted the greywacke-conglomerate-ultramafic association as outlier of overthrusted Pointe Verte Formation.

The West Gabbro Zone (WGZ)

At the WGZ (Fig. 2), the mafic body of the Elmtree deposit is a sill-like gabbroic intrusion outcropping for about 600 m along strike and varying in width from 3 to 40 m (Hoy *in* McCutcheon et al., 1988). It appears to be conformable with the structural trend of the surrounding rocks. The gabbro is bounded by black argillites of the Elmtree Formation which are locally tranformed into hornfels according to McCutcheon et al. (1988). Three trenches, perpendicular to the structural trend (Fig. 2), provide good exposure of the gabbro, and have been mapped in detail during this study (Fig. 4, 5, 6).

Paktunc and Ketchum (1989) and Hoy *in* McCutcheon et al. (1988) have noted that the gabbro is zoned towards the centre. Three phases have been recognized: (1) a marginal fine grained ophitic gabbro grading into (2) a medium grained ophitic-subophitic gabbro and (3) a coarse grained cumulate anorthositic gabbro towards the centre (Paktunc and Ketchum, 1989). Our observations are essentially similar to those of earlier workers except that we have mapped a quartz-bearing,

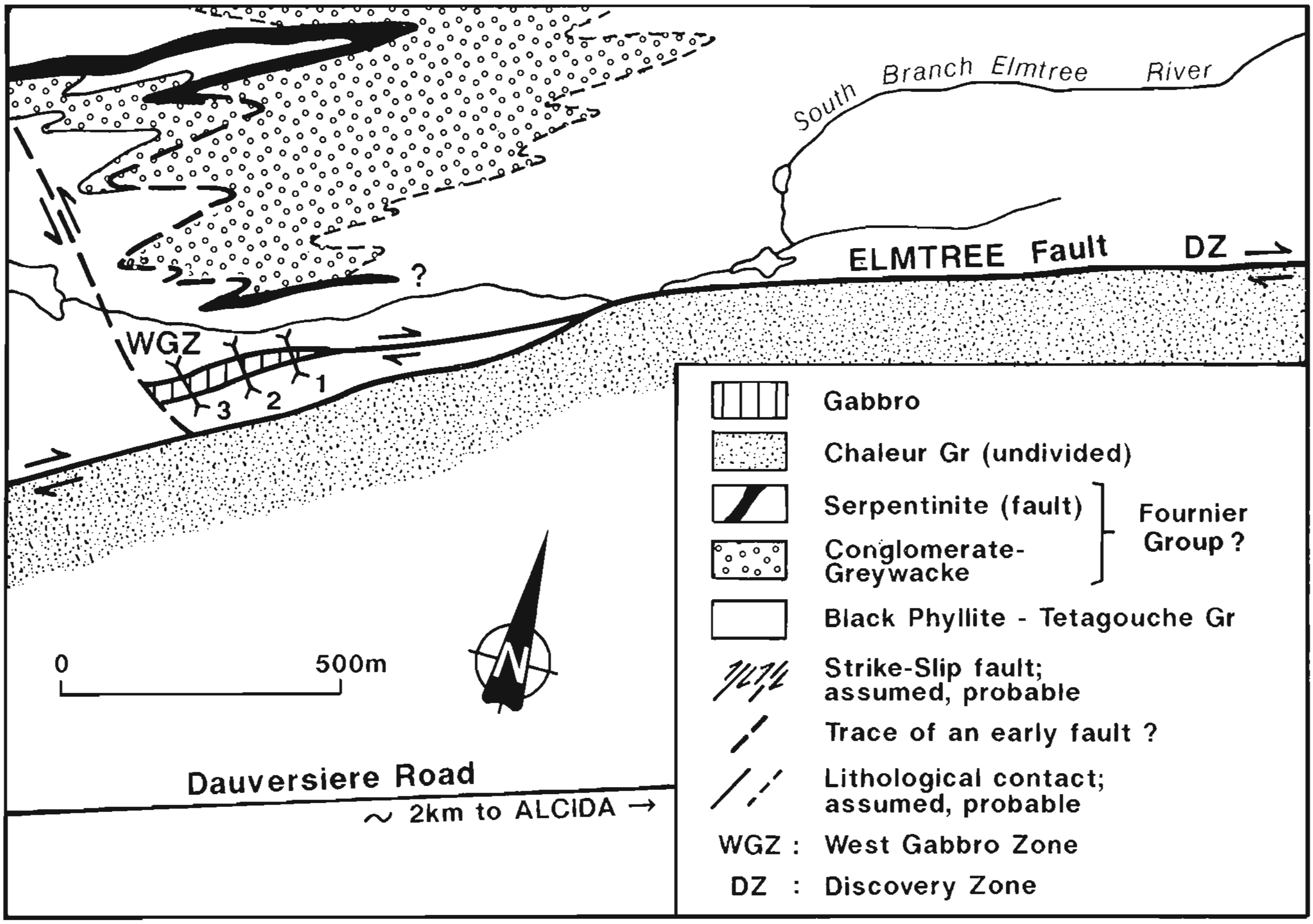


Figure 2. Schematic and interpretative geological map of the Elmtree deposit area. Numbers along the West Gabbro Zone (WGZ) refer to trenches mapped during this study (*see* text). Modified from Hoy (1985).

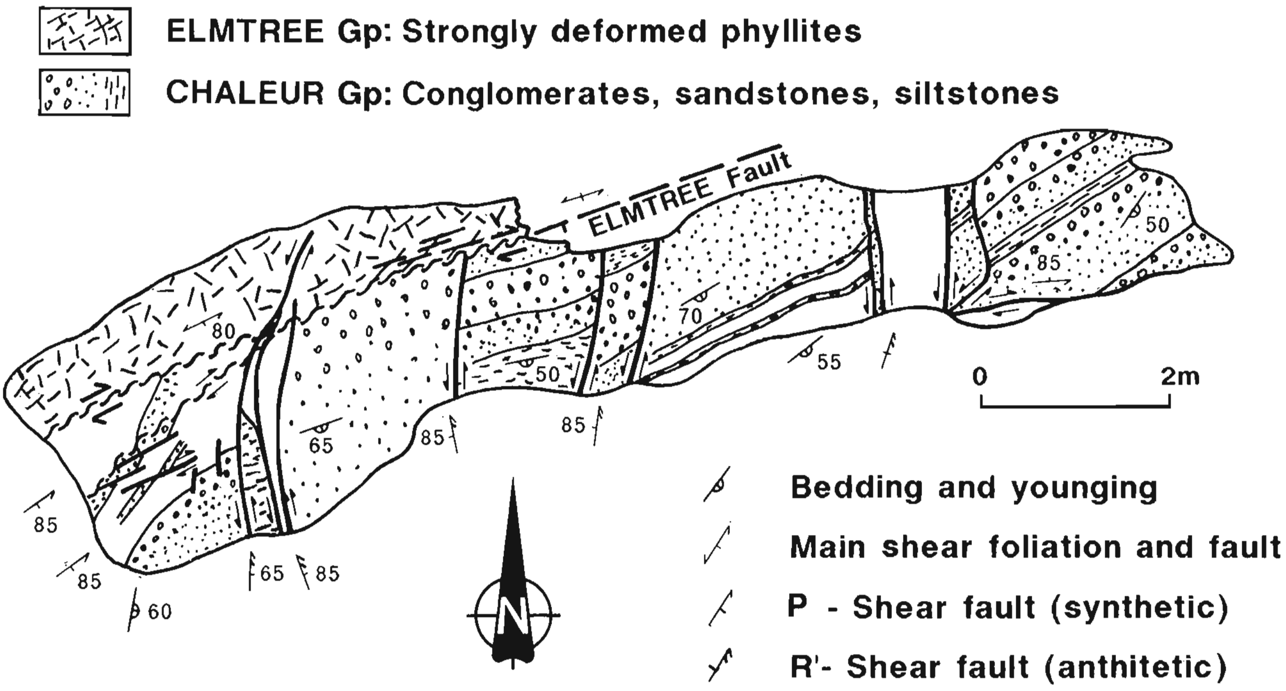


Figure 3. Geological sketch of an outcrop along the Elmtree fault on the Corona property. See text for a descriptive interpretation.

iron-rich phase in the central part of the anorthositic gabbro. The differentiation is visible in all trenches but is best developed in trench #3 (Fig. 6). The reader is referred to Paktunc and Ketchum (1989) for complete macroscopic and microscopic mineralogical descriptions and relations. As suggested by these authors, we believe that petrographical variations within this gabbro are related to flow differentiation.

Structure and ore morphology

Although there are good indications that the Elmtree break is a dextral strike-slip fault on the Corona property (Fig. 3), the geometrical relationships of shears, quartz veins and fractures in the WGZ are not perfectly coherent with a horizontal transcurrent faulting model. The structural framework of the WGZ is dominated by an anastomosing pattern of shear zones mainly developed within the central coarse grained part of the intrusion. Two types of shear zones have been recognized and divided into high-strain (HSSZ) and medium-strain shear zones (MSSZ). Both types are present in trenches #1 and #2 (Fig. 4, 5) while trench #3 displays only the second type (Fig. 6). HSSZ are associated with strongly chloritized and carbonatized rocks and occur at the northwestern and southeastern margins of the exposed gabbro. Along strike projection of the margins from trenches #1-2 to #3 suggest that these are not intersected by the trench, which could explain the absence of HSSZ at trench #3.

At the northwestern margin of trenches #1 and #2, HSSZ are marked by 1 to 1.5 m wide altered phyllonitic rocks trending WSW with a subvertical dip. The phyllonites are associated with dismembered and foliation-parallel quartz veins (Figs 4, 5). In trench #1, an apparent dextral offset of some quartz veins is visible. Subvertical NW trending tension veins are compatible with a dextral strike-slip but striations on shear veins plunge moderately (65°) to the SE. In trench #2 (Fig. 5), a dextral component is suggested by bending of the shear foliation along low-angle synthetic shears, and by asymmetric porphyroclasts of dismembered quartz veins. Subhorizontal shear-related striations are present. The kinematics of the southeastern HSSZ are less constrained. In trench #1, an apparent dextral offset of a folded quartz vein was noted but down-dip striations and moderately dipping C-S-like fabrics suggest a south-over-north vertical component. In trench #2, the shear zone is 1 m wide and contains lensoid-shaped pods of altered gabbro. No unambiguous kinematic indicators have been observed. It is suggested that the HSSZ represent dextral wrench-dominated or oblique-reverse southeast-over-northwest fault zones. The fact that they are often associated with folded and faulted quartz veins as well as foliation-parallel shear veins suggests that their development slightly succeeds the onset of vein formation.

The MSSZ occurs between HSSZ in trenches #1 and #2 and characterize trench #3. They form the anastomosing pattern noted by Paktunc and Ketchum (1989). MSSZ are underlined by centimetre to decimetre-wide deformation corridors associated with a NE-SW trending centimetre-spaced shear foliation. The latter possesses a well-developed anastomosing character related to foliation moulding around resistant felsic granules. Outside the shear zone, the gabbro is slightly deformed or undeformed. A weak schistosity, trending approximately N240/70-90°NW, is defined by flattened minerals. No lineations are visible. This schistosity correlates with the regional foliation of the surrounding rocks. Paktunc and Ketchum (1989) suggested that the two fabrics could represent a C-S fabric indicating a dextral shearing along localized zones but we did not observe any clear indication of such a relationship. MSSZ are associated with shear-related quartz veins (central and oblique shear veins) which are frequently striated (Fig. 4 to 6). All striations indicate a steeply plunging tectonic movement vector (Fig. 7) suggesting a dominance of vertical rather than horizontal movement along the MSSZ, but there is no visible lineation in the surrounding gabbroic rocks to corroborate this. However, both sub-vertical and sub-horizontal extension veins and fractures are associated with MSSZ and suggest a complex combination of horizontal and vertical movement.

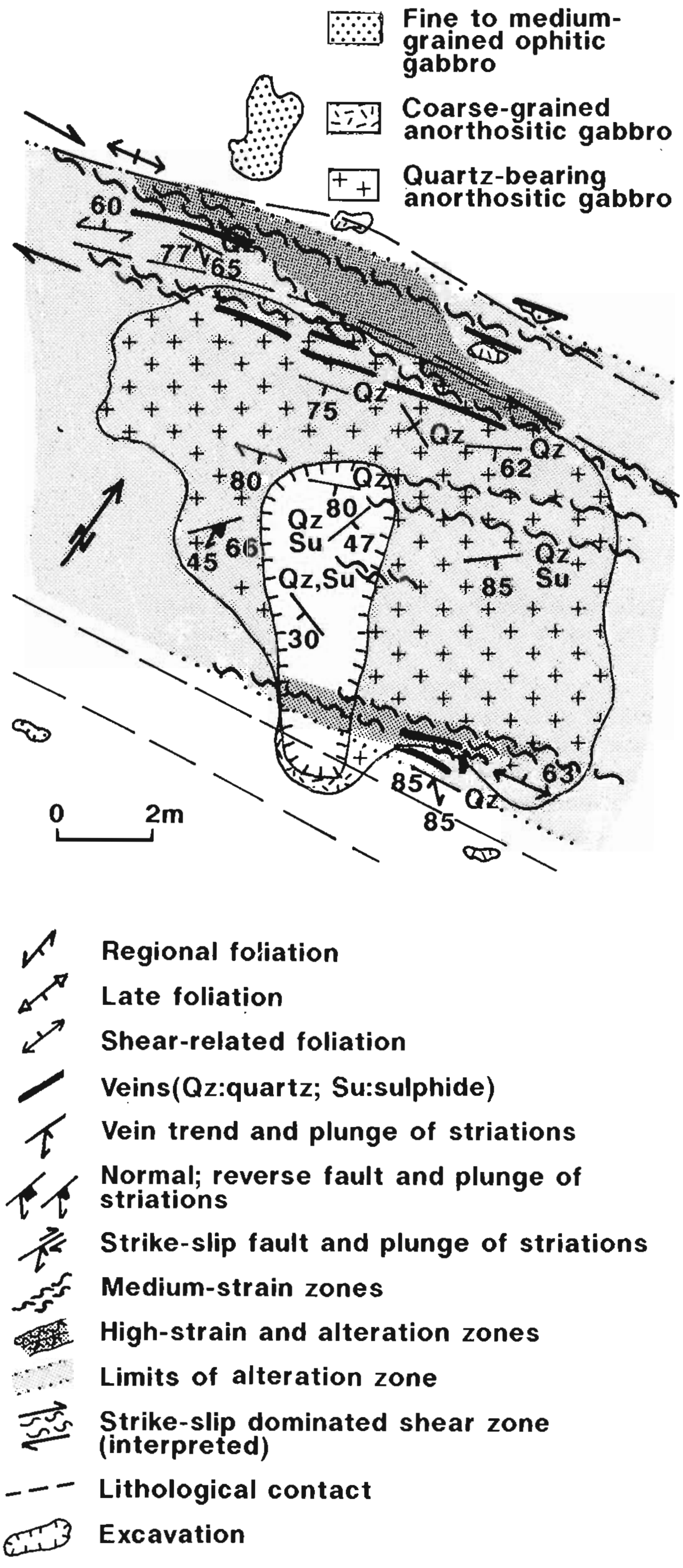


Figure 4. Detailed sketch of trench #1 at the West Gabbro Zone of the Elmtree deposit. See Figure 2 for location.

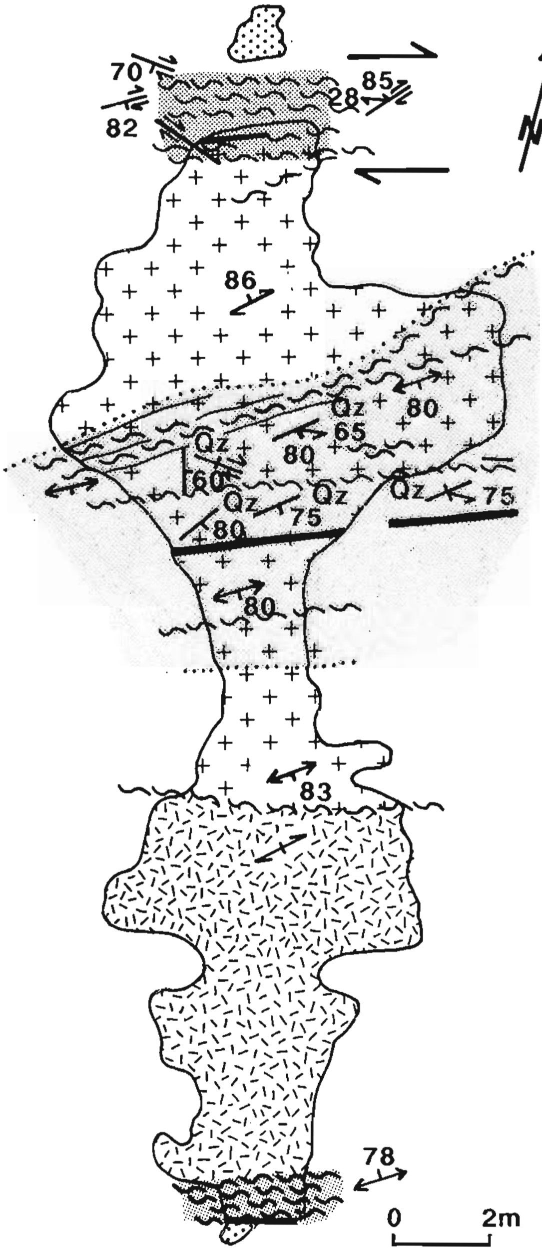


Figure 5. Detailed sketch of trench #2 at the West Gabbro Zone of the Elmtree deposit. Same legend as Figure 4. See Figure 2 for location.

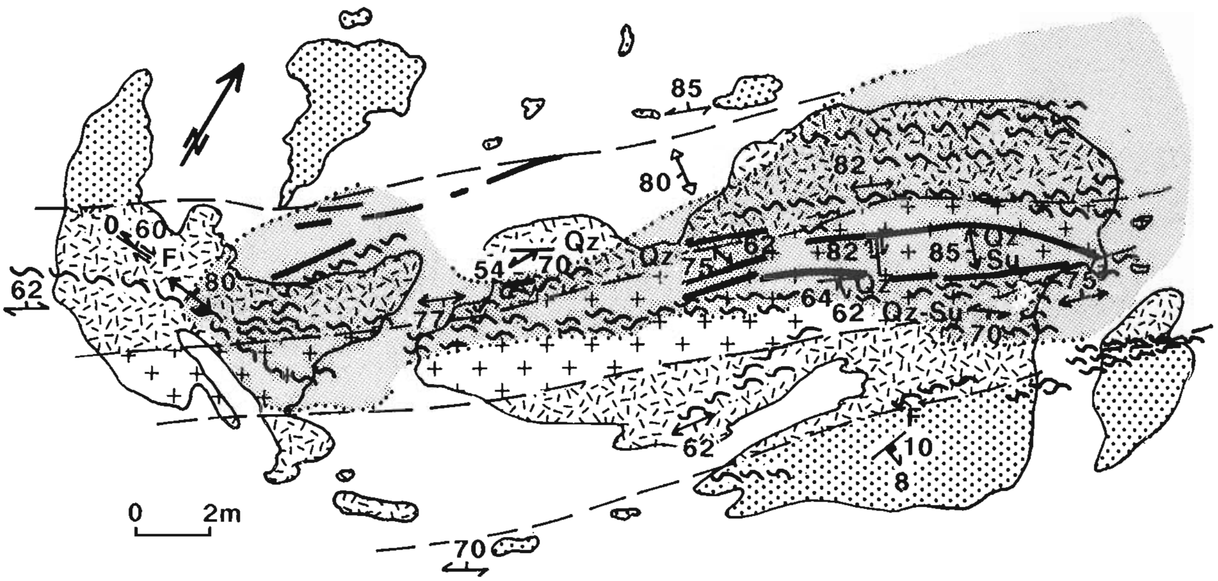


Figure 6. Detailed sketch of trench #3 at the West Gabbro Zone of the Elmtree deposit. The width of quartz veins is slightly exagerated. Same legend as Figure 4. See Figure 2 for location.

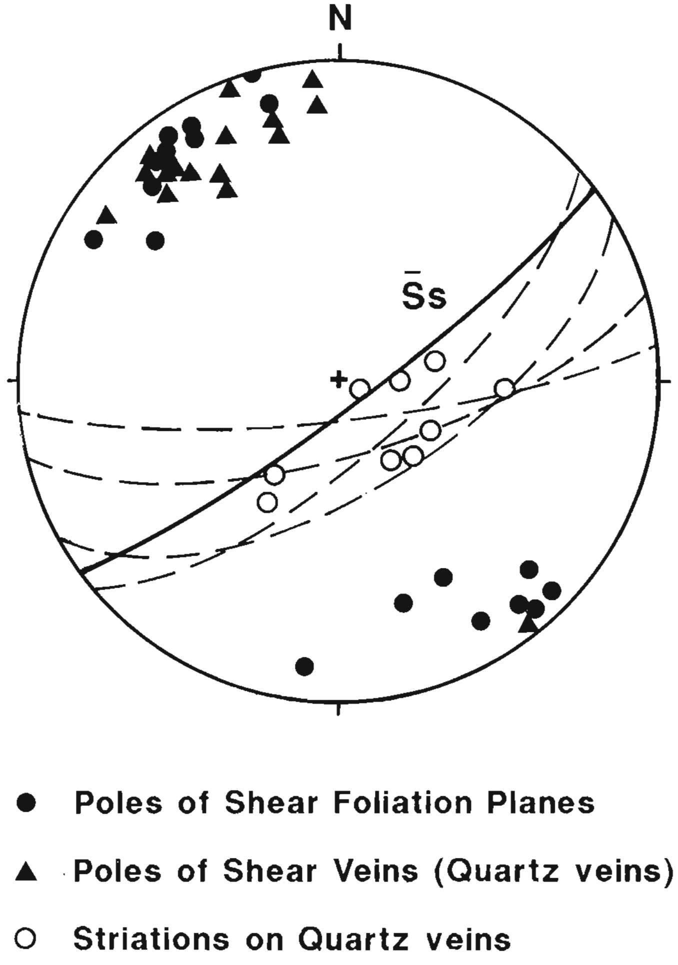


Figure 7. Lower hemisphere equal-area stereographic plot for shear-related planar and linear fabrics from the West Gabbro Zone of the Elmtree deposit. Ss represents the mean strike and dip of shear foliations. Dashed lines are average trends of shear veins (quartz veins).

A spaced cleavage locally overprints the shear foliation. It is best developed in trench #3 and trends approximately NW (Fig. 6). The exact relationship of this cleavage with earlier fabrics is not understood but it could be related to a late deformational event (Paktunc and Ketchum, 1989) or to sinistral faulting at the southwestern extension of the gabbro (Fig. 2).

The highest gold values are reported from sulphide-rich zones cutting the central quartz-bearing iron-rich anorthositic gabbro in proximity to quartz veins and from intensely sheared and altered zones (HSSZ; Hoy, 1985). The most abundant sulphides (arsenopyrite, pyrite and pyrrhotite) are dissiminated throughout the altered zones (Watson, 1988) and are also concentrated in foliation-parallel veinlets in both types of shear zones as well as in high-angle veinlets and fractures in MSSZ and related alteration zones.

Discussion and interpretation

Gold is structurally-controlled as indicated by the close relationship between shear zones and mineralization. Both types of shear zones are mineralized although the HSSZ carry the highest gold values (D. Hoy, pers. comm., 1990). The HSSZ appear as wrench-dominated deformation zones whereas available data suggest that MSSZ are probably related to a dominantly vertical movement component. Vertical movements along the Elmtree fault are also suggested by down-dip stretching lineations defined by elongated pebbles in Silurian conglomerates from the Discovery zone (Fig. 2). Because at this stage there is no clear evidence for more than one stage of progressive deformation within the WGZ, we suggest that both the HSSZ and MSSZ are coeval and related to a dextral transpressive tectonic regime along the Elmtree fault. The Elmtree deposit is interpreted as a contractional duplex (Woodcock and Fisher, 1986) bordered by dextral strike-slip faults (Fig. 2). The inferred

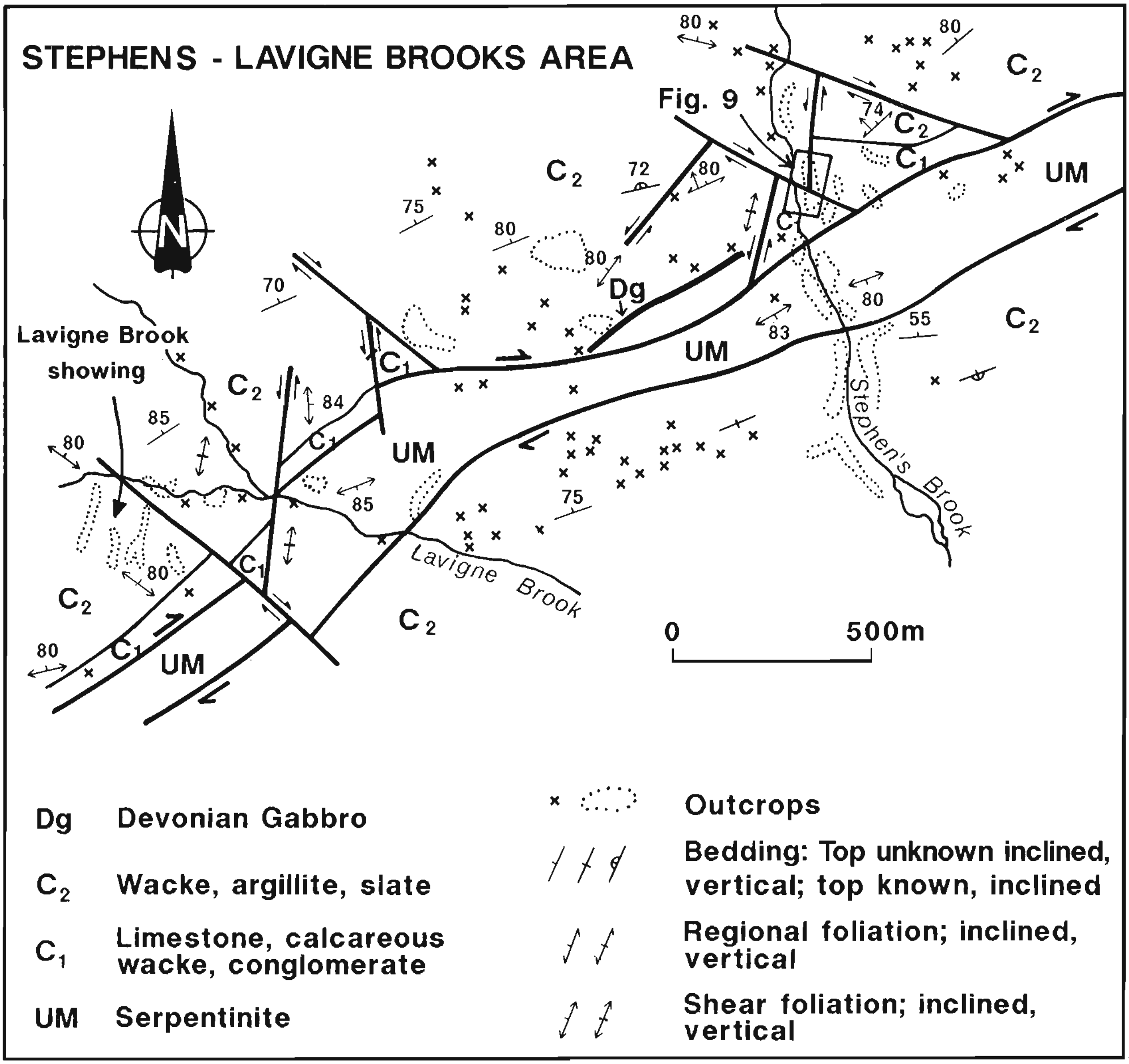


Figure 8. Geological map of the main break of the Rocky Brook-Millstream fault system between the Stephens and Lavigne brooks. Modified from Philpott (1987).

deformational regime agrees with the transpressionnal character generally attributed to tectonic events leading to dextral strike-slip faulting in northern New Brunswick (van Staal and Langton, 1988).

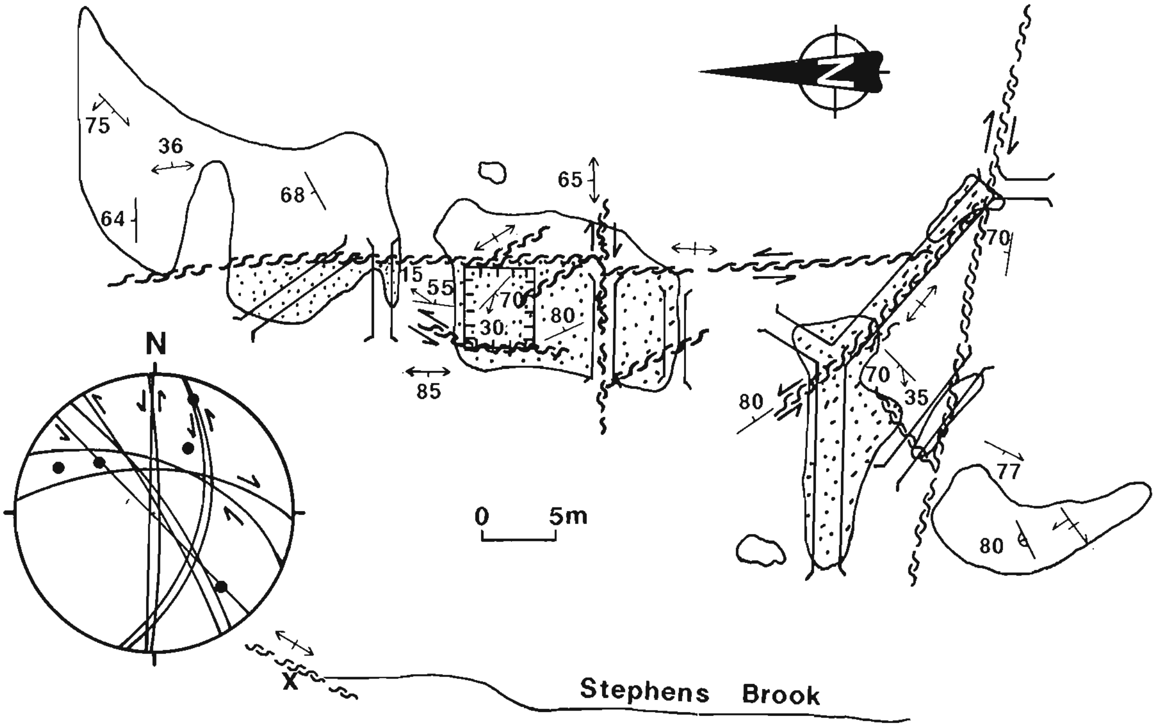
Gold mineralization along the Elmtree fault is similar to several Archean gold deposits hosted by differenciated gabbroic sill such as the Norbeau mine (Dubé et al., 1989), the San Antonio mine (Poulsen et al., 1986) or the Golden Mile dolerite in Kalgoorlie, Australia (Phillips, 1986). Elmtree is a typical example of a gabbroic host rock exerting a rheological and a chemical control on gold mineralization. The competency contrast between the gabbro and the surrounding rocks causes a more brittle response in the gabbro, allowing the preferential development of brittle-ductile shear zones in the latter, whereas its iron-rich composition favoured the precipitation of gold (Dubé et al., 1987).

GOLD ALONG THE RBMFZ

In the Bathurst area, gold occurrences along the RBMFZ are restricted to the main break and hosted by Silurian rocks of the Chaleur Group. We have visited the Stephens-Lavigne-Bradley Brook showings (sites 2, 3 and 4 on Fig. 1). Along the south branch of the RBMFZ, gold appears as a secondary mineralization associated with base metal deposits. For the purpose of our study, the Rocky Brook Shaft deposit (site 5 on Fig. 1) is taken as a case study.

THE MAIN BREAK

The main break is underlain by tectonic slivers of foliated serpentinite and serpentinite breccias interpreted as fault products. Along Stephens Brook, a central zone of brecciated serpentinite is bordered to the NW and the SE by



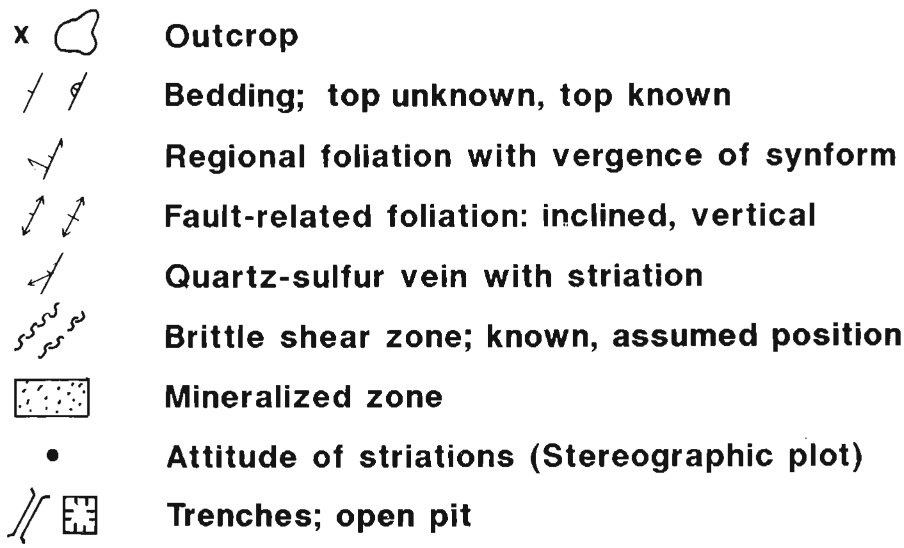


Figure 9. Detailed sketch of the Stephens Brook prospect. The lower hemisphere equal-area stereographic plot shows the trend of various shear zones and associated striations. Arrows indicate the sense of shear. See Figure 8 for location and text for detail.

massive serpentinites in faulted contact with sedimentary rocks of the Chaleur Group. In the latter, younging directions are opposite on both sides of the main break which suggests a corresponding antiformal structure (Philpott, 1987). Breccias consist of a serpentine or serpentine-carbonate matrix supporting centimetre-scale fragments of ultramafic and calcareous compositions. The protolith of serpentinite appears to be mainly harzburgitic in composition. In many aspects, these rocks are similar to the serpentinite occurring north of the Elmtree deposit.

A modified version of Philpott's (1987) compilation map is presented in Figure 8. It shows the geology along the main break between the Stephens and Lavigne brooks. As previously recognized, the main break is shown as a NE trending dextral strike-slip fault. Synthetic and antithetic strike-slip faults are interpreted as related subsidiary faulting (Fig. 8). Gold occurrences are associated with dextral (Lavigne Brook) and sinistral splays (Stephens Brook) of the main break. The Bradley Brook showing (Fig. 1) is associated with extensional veins. Shear vein-type occurrences will be examplified by the Stephens Brook prospect in the following section.

The Stephens Brook prospect

The mineralization is located east of Stephens Brook, where an open pit and several trenches provide excellent exposure of the ore zone (Fig. 9). Gold assays reveal up to 9.3 g/t Au over 30 cm to 1.2 m (Burton, 1985).

The mineralization is related to a north-south, sinistral, brittle shear zone (Fig. 9). The shear zone is underlain by massive to semi-massive arsenopyrite and quartz veins or by brecciated host rocks within an arsenopyrite-rich matrix. The movement sense of the fault zone is indicated by bending of the regional foliation and bedding in the host rocks, by local asymmetrical folds, and by striations and steps in some shear veins. The sulphide-rich zone is located on the western side of the north-south shear zone (Fig. 9). Some low-angle synthetic shears are present and contribute to more shearing and brecciation within the mineralized zone. Between these shears, the rock is commonly fractured or brecciated and contains 5 to 10% arsenopyrite. The mineralized zone is silicified. Minor reverse and normal faults are present. Northeast of the mineralized zone, the host rocks comprise interbedded sandstones, wackes and shales whereas to the SW, they comprise a well-bedded sequence of calcareous siltstones, graded sandstones and conglomerates. To the south, the mineralized zone is abruptly crosscut by an E-W trending dextral shear (Fig. 9) that is interpreted as a synthetic fault related to the main break.

The lower hemisphere equal area plot shown in Figure 9 presents a compilation of these structures. Subvertical north-south sinistral faults are associated with subvertical NW-SE synthetic faults and crosscut by east-west dextral faults steeply dipping to the north. Shear-related lineations plunge weakly to moderately (Fig. 9). It seems that the ore fluids, originating from the main break, travelled along the north-south sinistral fault system, which was subsequently crosscut by dextral faults associated with a continuous deformation along the main break.

Discussion

Along the main break, gold and base metal showings do not occur in the same structural settings and do not bear the same fluid composition. At Lavigne Brook (Fig. 8), gold is related to WNW-ENE dextral brittle shears and subsidiairy NW-SE synthetic shears. In contrast to the Stephens Brook area, at Lavigne Brook north-south sinistral shears appear as crosscutting structures within the mineralized zone. The ore mineralogy at Lavigne Brook is also slightly different from the Stephens Brook area and consists of pyrite-chalcopyrite-sphalerite. The local occurrence of garnet within the host rocks and skarns suggests a possible magmatic component to the ore fluid composition, originating from the Devonian Nicholas Denys Intrusion (Fyffe et al., 1981). As already noted by Ruitenberg et al. (1990), at Bradley Brook, the mineralization is located within NW-SE extensional veins of arsenopyrite-galena and quartz.

Structural analysis of gold occurrences along the main break reveals that ore fluids were trapped in various settings related to splays along the main décollement zone. This is attributed to continuous hydrothermal activity, possibly of a valve and pump mechanical-type (Sibson et al.,1988), along the main break. Various compositions of mineralizing fluids are inferred from variable ore mineralogy in different sites and could be possibly correlatives with distinctive structural traps.

THE SOUTH BRANCH

The south branch of the RBMFZ marks the contact between the Chaleur Group and the Tetagouche Group (Fig. 1). This fault is slightly oblique to the main break. As opposed to the latter, the south branch is not underlain by any type of exotic rocks (ultramafics or skarns) and is mapped as a sharp fault zone frequently offset by WNW-ESE dextral faults (Philpott, 1987). We believe that the south branch represents a subsidiary fault within the RBMFZ system. Because the main break is associated with relatively wide slivers of ultramafic rocks and abundant products of dynamic hydrothermal activity, it is suspected to have been the locus of a much greater lateral transport than the south branch. Although it is not identified as an important gold-bearing structure, the south branch is associated with some interesting base metal (with minor gold) occurrences that are exemplified by the Rocky Brook Shaft deposit.

The Rocky Brook Shaft deposit

The Rocky Brook Shaft deposit is located approximately 400 m south of the main break at, or near, the contact between the Chaleur Group and the Tetagouche Group (Fig. 10). It is mapped as a Zn-Pb-Au massive sulphide deposit enclosed within Tetagouche Group argillites

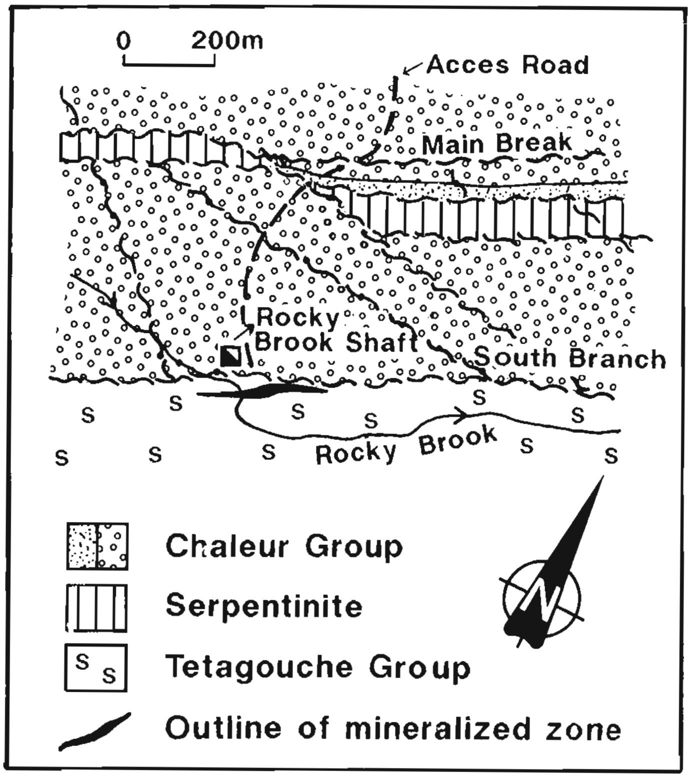


Figure 10. Location and geological map for the Rocky Brook Shaft deposit. After Philpott (1987).

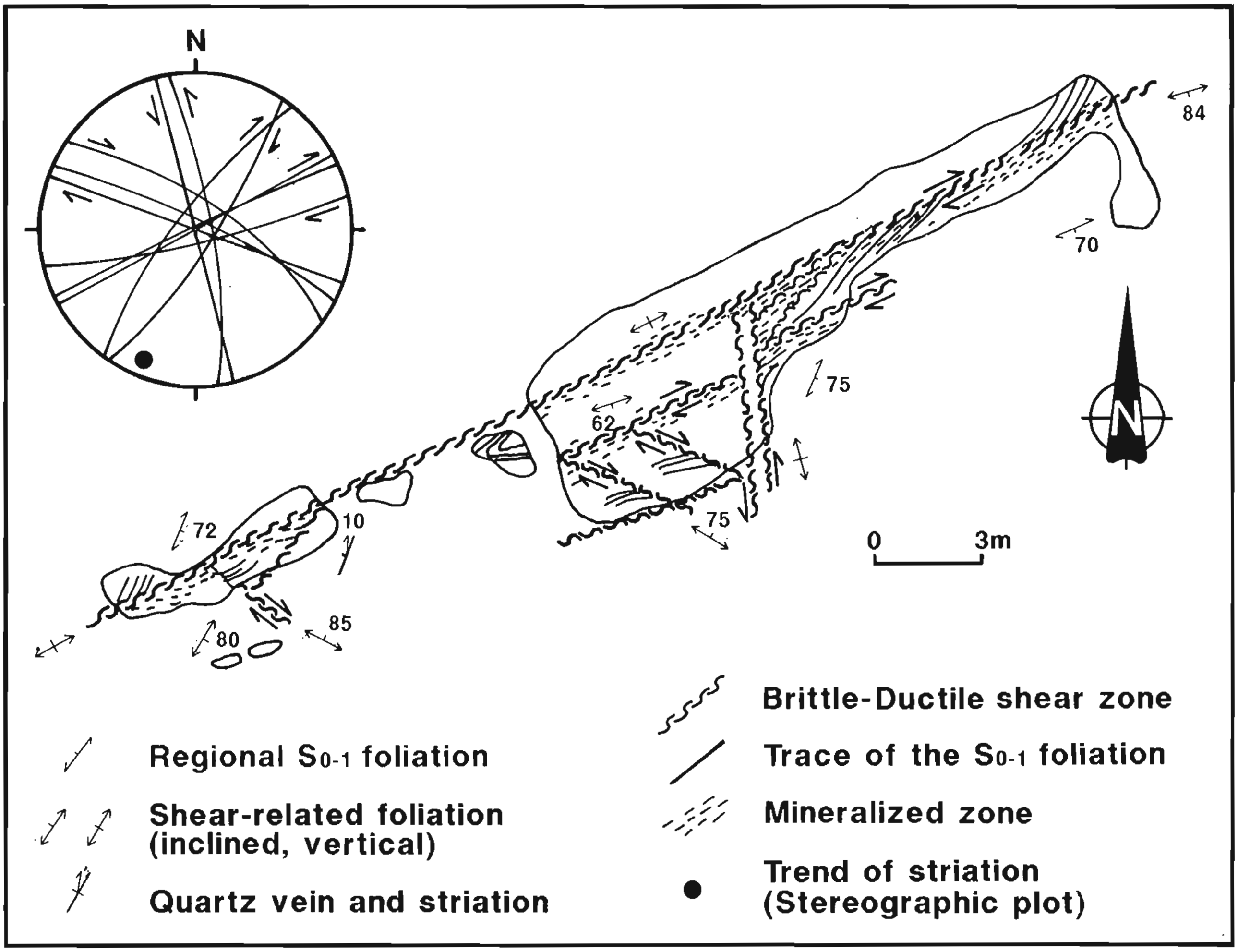


Figure 11. Detailed sketch of a trench from the Rocky Brook Shaft deposit. The lower hemisphere equal-area stereographic plot shows the trend of various prominent shear zones and associated striations. Arrows indicate the sense of shear. See text for detail.

(Philpott, 1987). In the field, the mineralization consists of several discontinuous lenses of massive to semi-massive sulphides associated with WSW trending shear zones. Several trenches are exposed around the old shaft and the structural geometry of the most representative is sketched in Figure 11.

The sulphide mineralization occurs along a 0.5 m wide dextral brittle-ductile shear zone associated with the development of a foliation and brecciation. The orientation of the shear zone is WSW-ENE (Fig. 11). Centimetre-wide low-angle and high-angle synthetic shears are associated with the main shear zone. A minor set of north trending antithetic (sinistral) shears crosscuts the mineralized zone. All the shears are subvertical or steeply inclined, and a few subhorizontal striations indicate lateral motion along these zones (*see* the stereographic plot of Fig. 11). Bending and drag folding of the regional foliation were used as kinematic indicators. Host rocks are siliceous mudrocks and well-foliated slates belonging to the Tetagouche Group.

Sulphides plus quartz occur as massive to semimassive veins from a few centimetres to approximately one metre wide. Central and oblique shear veins contain up to 80% sulphides. Stringers and veinlets of quartz and sulphides lie between the shear zones. Some of the latter are NNW trending extensional veins. The ore mineralogy mainly comprises pyrite, sphalerite, chalcopyrite and galena. Galena may be locally more abundant than sphalerite. Silicified host rocks adjacent to shear zones contain up to 50% dissiminated sulphides, which contribute to a well-developed gossan-type alteration.

Discussion

The Rocky Brook Shaft mineralization is similar to that at the Lavigne Brook showing. Both types of mineralization are associated with dextral shearing and crosscutting sinistral shears. However, the ore zone at Lavigne Brook is related to a splay of the main break whereas the Rocky Brook Shaft deposit seems to occur exactly on the south branch fault zone. Also, no skarn-type alteration or mineralogy is present around the Rocky Brook Shaft deposit.

CONCLUSION

Gold occurrences along the Elmtree and Rocky Brook-Millstream fault zones of northern New Brunswick are structurally controlled and associated with both vein formation and wallrock alteration. Detailed studies along the Elmtree fault suggest a transpressive deformational regime implying the coexistence of strike-slip and dip-slip structures at the Elmtree deposit. Ongoing laboratory studies on several structurally-oriented samples from the WGZ will hopefully constrain the contractionnal duplex hypothesis proposed in this paper. Fieldwork on gold and base metal occurrences along the RBMFZ confirms the strike-slip nature of this fault system. Mineralization is associated with dextral and sinistral splays of the main dextral fault zone. The south branch of the RBMFZ is interpreted as a dextral splay of the main break. Although not clearly demonstrated in this study, a transpressive deformationnal regime is also attributed to the RBMFZ (van Staal and Langton, 1988; van Staal et al., 1990)

Gold occurrences of this study are similar to the structurally-controlled mesothermal vein-type according to Dubé's (1990) classification. The Elmtree deposit belongs to the altered wallrock subtype as does the Stog'er tight prospect (Dubé, 1990) in Newfoundland. They are spatially related to major fault zones generated respectively during the Acadian orogeny in New Brunswick and the Acadian and/or Taconain orogenies in Newfoundland.

ACKNOWLEDGMENTS

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Detailed gravity traverses in the Appalachian Dunnage and Gander terranes, Northern New Brunswick

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Thomas, M.D., Halliday, D.W., and O'Dowd, D.V., Detailed gravity traverses in the Appalachian Dunnage and Gander terranes, northern New Brunswick; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 101-109, 1991.

Abstract

Detailed gravity surveys in the Dunnage and Gander terranes of the northern Appalachians of New Brunswick have defined more precisely several anomalies associated with Ordovician rocks. The anomalies are associated with the basaltic rocks of the Nine Mile synform, rhyolitic rocks of the Tetagouche antiform, and a structural package of predominantly basaltic rocks near Camel Back Mountain on the western flank of the antiform. A small positive anomaly, 4 mGal amplitude, has been outlined over ophiolitic rocks of the Elmtree inlier. It signifies that these rocks are quite thin, probably no more than a few hundred metres thick. The gravity surveys have also improved definition, locally, of prominent negative anomalies associated with the Antinouri Lake and Pabineau granites. The improved resolution of anomalies in the region will result in gravity models which are better constrained.

Résumé

Des levés gravimétriques détaillés des zones Dunnage et Gander de la région nord des Appalaches au Nouveau-Brunswick ont permis de définir de manière plus précise plusieurs anomalies associées à des roches de l'Ordovicien. Ces anomalies sont associées aux roches basaltiques de la synforme de Nine Mile, aux roches rhyolitiques de l'antiforme de Tétagouche et à un ensemble de roches principalement basaltiques situé près du mont Camel Back sur le flanc ouest de l'antiforme. Une petite anomalie positive d'une amplitude de 4 mGal a été délimitée sur des roches ophiolitiques de la boutonnière d'Elmtree. Cela signifie que ces roches sont très minces et que leur épaisseur ne dépasse guère quelques centaines de mètres. Les levés gravimétriques ont également permis d'améliorer par endroits la définition des anomalies négatives remarquables associées aux granites d'Antinouri Lake et de Pabineau. La meilleure résolution des anomalies de la région permettra de construire des modèles gravimétriques mieux définis.

[1] Geophysics Division

INTRODUCTION

Recent investigations into the structure and tectonic history of the Appalachian orogen in the vicinity of Bathurst, northern New Brunswick (van Staal, 1986, 1987; van Staal et al., 1988) have outlined a need for further insight into the third dimension of various crustal units. In an attempt to meet this requirement, detailed gravity surveys were carried out along several traverses crossing selected geological targets. Several detailed ground magnetic traverses were also surveyed. The magnetic investigations are the subject of a separate report (Thomas et al., 1991).

Aside from the geological interest, positioning of traverses was guided by rock density considerations, i.e. the likelihood of the geological target producing a discrete gravity anomaly, and/or the existence of gravity anomalies as indicated by regional gravity data (station spacing about 5 to 10 km) in the National Gravity Data Base. A total of 376 new observations was made, principally along 7 traverses totalling about 160 km in length. A few observations were positioned to improve the regional coverage, and on an opportunistic basis, at bench marks used for altimeter ties. The traverse locations are shown on the geology map (Fig. 1), on the combined Bouguer gravity anomaly-simplified geology map (Fig. 2) and on the Bouguer anomaly map (Fig. 3).

Geological features crossed by the traverses include:

1. The Ordovician Elmtree inlier.
2. The Rocky Brook-Millstream fault.
3. A fold-thrust package on the south side of the Rocky Brook-Millstream fault comprising rocks of the Ordovician Fournier and Tetagouche Groups.
4. The major, subparallel Tetagouche antiform and Nine Mile synform involving rocks of the Fournier and Tetagouche Groups.
5. The Devonian Pabineau granite.

In this report, we present a new regional gravity map of the area incorporating the 1990 detailed survey data, gravity profiles along the traverses, and a preliminary assessment of the geological significance of the new data.

GEOLOGICAL SETTING

A simplified geological map of the region based on Davies (1979), and modified by van Staal et al. (1990) and van Staal and Langton (1990) is shown in Figure 1. The region contains the boundary between the Dunnage and Gander tectono-stratigraphic zones of the Appalachian orogen as defined by Williams (1979), which, in New Brunswick, was thought to lie at or near the Rocky Brook-Millstream fault. The Dunnage zone lay to the north of the fault, and the Gander zone to the south. The Dunnage zone, in general terms, was thought to represent the vestiges of the Iapetus Ocean, e.g. island arc assemblages that developed on oceanic crust, whereas the Gander zone was generally interpreted as the eastern continental margin of the Iapetus Ocean, which locally may have been of Andean-type (Williams, 1979). Although initially referred to as zones, these tectono-stratigraphic units are now commonly termed terranes.

Williams and Hatcher (1983) noted that, in New Brunswick and southward, where volcanic rocks similar to those of the Dunnage terrane overlie clastics of the Gander terrane, the distinction between the two terranes can only be made at deeper stratigraphic levels. Van Staal (1987) was of a similar opinion, concluding that, because "Gander" rocks represent basement to the "Dunnage" rocks, a tectono-stratigraphic boundary between the two terranes in this area has no real significance. More recently, van Staal et al. (1990) added support to this idea by demonstrating that the Ordovician Fournier Group is represented on both sides of the Rocky Brook-Millstream fault, thereby eliminating the fault as the Dunnage-Gander boundary. Notwithstanding the demise of the fault as the accepted boundary, for the purpose of describing the geology of the area, it provides a convenient line of subdivision. In the following descriptions, therefore, the terms Dunnage and Gander terranes equate with regions north of and south of the fault, respectively.

The Dunnage Terrane

The Dunnage terrane contains the Ordovician Elmtree inlier, near the coast of Chaleur Bay, which is surrounded on most sides by Silurian sedimentary rocks of the Chaleur Group, although a large intrusion of Devonian gabbro and diabase flanks it to the west (Davies, 1979). Two groups of rocks are recognized in the inlier (van Staal and Langton, 1990). The Fournier Group is Middle Ordovician (Sullivan et al., 1990) and lies structurally above the older Tetagouche Group. The Tetagouche Group is represented by the sedimentary Elmtree Formation, formed largely of slate and shale. The Fournier Group comprises basalt, gabbro, diabase, minor plagiogranite, greywacke and slate, which are assigned to two formations. An association of pillow lavas, gabbros, amphibolites, trondhjemites, sheeted dykes, greywacke and slate, which defines the Deveraux Formation (van Staal and Langton, 1990), was interpreted as an ophiolitic sequence by Pajari et al. (1977). The Pointe Verte Formation includes sedimentary rocks and alkali basalt.

The Silurian Chaleur Group forms an east-northeast-trending belt separating the Elmtree inlier and various Silurian and Devonian rocks to the north from Ordovician rocks of the Miramichi Highlands to the south. The Chaleur Group consists predominantly of sedimentary rock types, including conglomerate, sandstone, siltstone, argillite, greywacke and limestone. Locally it is overlain unconformably by Carboniferous sandstones, conglomerates, siltstones and shales. Davies (1979) showed the belt as being fault-bounded on both sides, the southern fault being part of the Rocky Brook-Millstream fault system.

The Elmtree inlier is intruded by the Devonian Antinouri Lake granite, and the Chaleur Group by the Devonian Nicholas Denys granite.

The Gander Terrane

In the Bathurst area, the Gander terrane is represented by Cambrian-Ordovician sedimentary rocks of the Miramichi Highlands. These rocks were previously known

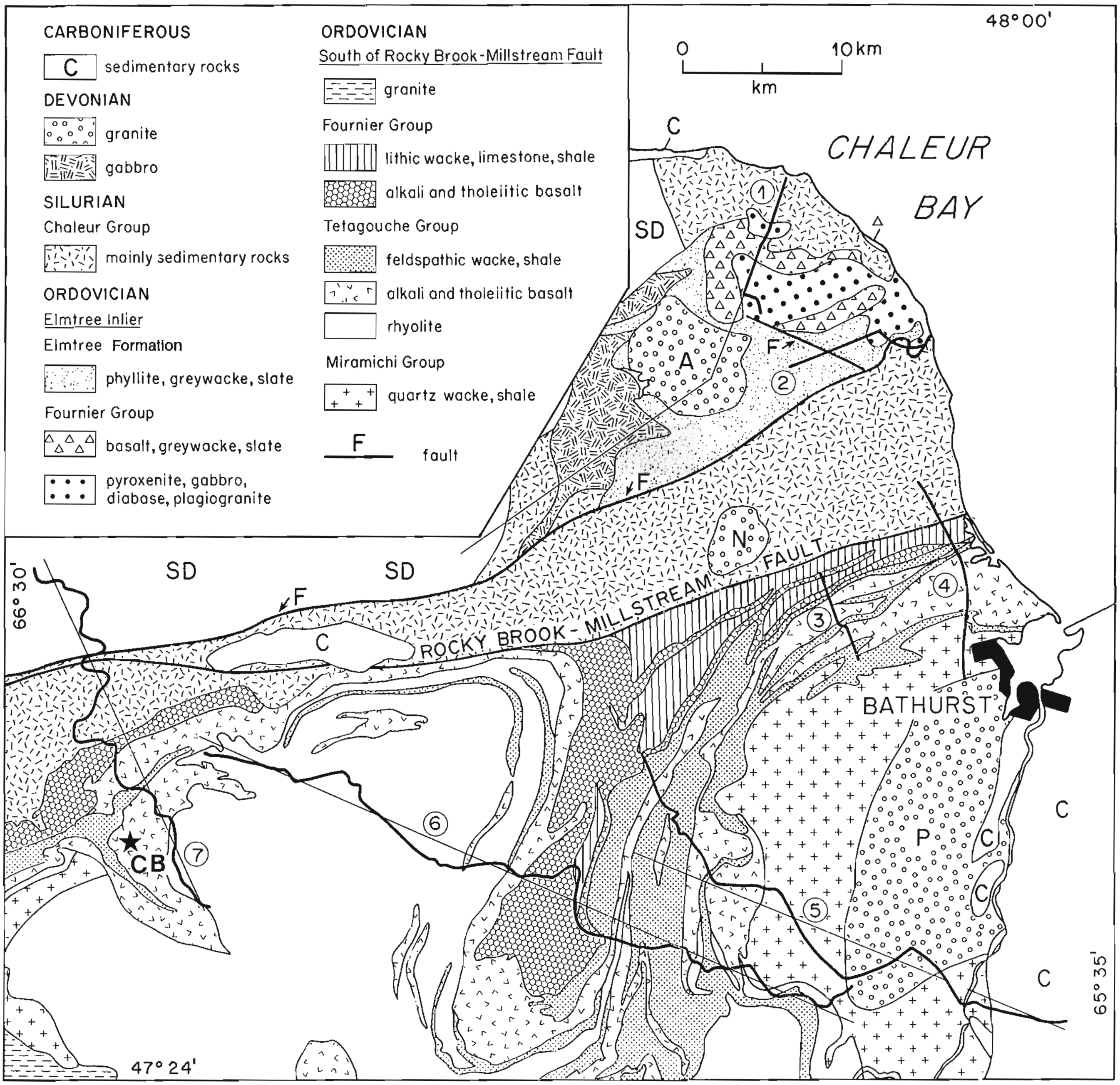

Figure 1. Geological map of region spanning Dunnage-Gander boundary zone near Bathurst, based on Davies (1979) and van Staal and Langton (1990). Devonian granites: A, Antinouri Lake; N, Nicholas Denys; P, Pabineau. SD, undivided Silurian and Devonian sedimentary and volcanic rocks. CB, Camel Back Mountain. Heavy black lines labelled 1 through 7 are detailed gravity traverses. Accompanying fine lines for traverses 5, 6 and 7 are corresponding profile lines onto which gravity values have been projected; profiles 1, 2, 3 and 4 follow closely their respective traverses. The fine line extending southwestward from traverse 1 marks the extension of profile 1 using regional gravity data.

informally as the Tetagouche Group (e.g. van Staal, 1987; van Staal et al., 1988). A more recent geological analysis and compilation (van Staal and Langton, 1990) subdivides the former Tetagouche succession into three groups. The oldest is the Miramichi Group, which is Arenigian and older. The two younger groups, a redefined Tetagouche Group and the Fournier Group, have similar ages extending from Llanvirnian to Caradocian. The Miramichi Group, comprising quartz wacke and shale, is restricted to the southeast corner of the area where it flanks the Devonian Pabineau granite. It is succeeded to the immediate north and west by relatively narrow belts of sedimentary (feldspathic wacke and shale) and volcanic (tholeiitic and alkali basalt) assemblages of the Tetagouche Group. Farther west, in the central part of the region, lies the relatively broad Nine Mile synform.

Its flanks comprise sedimentary and volcanic rocks of the Tetagouche Group, whereas the core is underlain by tholeiitic and alkali basalt and lithic wacke, limestone and shale of the Fournier Group. Westward again is the broad Tetagouche antiform built predominantly of rhyolite and associated epiclastic sedimentary rocks of the Tetagouche Group. The northwestern and western flanks of this antiform comprise mainly basaltic rocks of both the Tetagouche and Fournier groups.

Structurally the region is dominated by the Tetagouche antiform and the Nine Mile synform, which are the latest folds formed during the polyphase structural history. Much narrower folds occurring in the southeast of the region are the Mud Lake, Pabineau and Brunswick antiforms and the Pabineau and Brunswick synforms. The axes of these folds trend variously between north and northeast.

Many of the lithological units on van Staal and Langton's (1990) geological compilation are bounded by thrusts, which, for clarity, are not indicated on the simplified geological map of Figure 1. The down dip direction of the thrusts, notwithstanding various excursions around antiforms and synforms, is towards the north. In the area north of Bathurst, the structure of the Ordovician rocks had been viewed as a north-younging homocline, but recent mapping by van Staal et al. (1988) indicated that stratigraphy is repeated and inverted, characteristics attributed to early D_1-related, east or southeast directed thrusting and folding. This structural package is traceable westward to the region of Camel Back Mountain. Basalts and associated sediments of the upper part of the Tetagouche Group (old definition) are interpreted structurally to overlie silicic volcanics and sediments of the lower part, the contact being an F_1-related thrust (van Staal, 1987; van Staal et al., 1988).

Plate Tectonic Models

Rast and Stringer (1980) interpreted the Elmtree ophiolitic complex of the Dunnage terrane as remnant oceanic crust of a southeast-dipping subduction zone, and the rocks of the Tetagouche Group (old definition) in the Gander terrane as representing a complementary volcanic arc. Subduction was terminated in late Lower Ordovician time through collision of the Iapetian mid-oceanic ridge with the Tetagouche arc.

Van Staal (1987) favoured a model in which both the Fournier and Tetagouche groups evolved in a marginal back-arc oceanic basin developed above a southeast-dipping subduction zone that closed the Iapetus Ocean to the northwest. Northwestward subduction leading to closure of the marginal basin was proposed to explain the occurrence of blue schist metamorphism in the Tetagouche rocks, and southeastward thrusting of these same rocks above silicic volcanics comprising the bulk of the lower part of the Tetagouche Group.

GRAVITY SURVEYS

Gravity observations were made with a Lacoste and Romberg gravity meter, No. 172, and tied to National Gravity Base Station 9243-73 located in Bathurst. Most stations were distributed along 7 traverses having a total length of approximately 160 km; altogether 376 observations were made. Station spacing was typically 0.5 km, but ranged from 0.2 to 0.7 km , and occasionally was as large as 1.0 km. Vertical control was provided in most cases by altimetry tied to both federal and provincial benchmarks; a single Atmospheric Instrumentation Research, dual diaphragm altimeter was used. Horizontal positions were obtained using 1:50 000 scale National Topographic System maps and 1:12 500 scale New Brunswick Department of Natural Resources forest development survey maps; in a few cases horizontal positions were determined using a Trimble Pathfinder global positioning system receiver. The estimated accuracy of elevations obtained by altimetry is ± 3 m, and of horizontal positions scaled from maps ± 20 m.

Bouguer gravity anomalies were computed using a standard uniform rock density of 2.67 g/cm^3 and sea level datum. Consideration of errors from all sources indicates that anomalies are accurate to within ± 1 mGal. This value does not take into account terrain corrections, which have not been computed. It is estimated that most corrections in this region will not exceed 1 mGal.

BOUGUER GRAVITY FIELD AND ITS RELATIONSHIP TO GEOLOGY

The regional gravity field (Fig. 3) is defined primarily by gravity observations spaced some 5 to 10 km apart. Along some highways spacing is closer, ranging from about 1 to 4 km, and the new traverses have stations at intervals ranging from about 0.2 to 0.7 km. In Figure 2 the gravity field is superimposed on a greatly simplified geological map to facilitate comparison. The pattern of Bouguer contours is characterized by a series of roughly oval-shaped anomalies, the major axes of which strike approximately northeast, in harmony with associated geological features. Because the gravity field is changeable across the region (values range from about -40 to $+5$ mGal), it is difficult to determine a local "background" level, with which anomalous portions may be compared. There is, however, a small area in the central part of the region (Fig. 3) where gravity anomaly values are confined to a relatively narrow range between -15 and -20 mGal. From this area, values fall rapidly northeastward into the area of the Antinouri Lake granite, and rise sharply southward to produce the anomaly associated with the Nine Mile synform. A smaller area characterized by values lying between -15 and -20 mGal lies northwest of Bathurst. From this area values decrease both northward and southward into the regions of the Antinouri Lake and Pabineau granites, respectively, and rise to the southwest over the Nine Mile synform. As a first approximation, the -20 mGal contour level is used to represent datum. Six principal gravity anomalies are identified in the area, most of which can be unequivocally associated

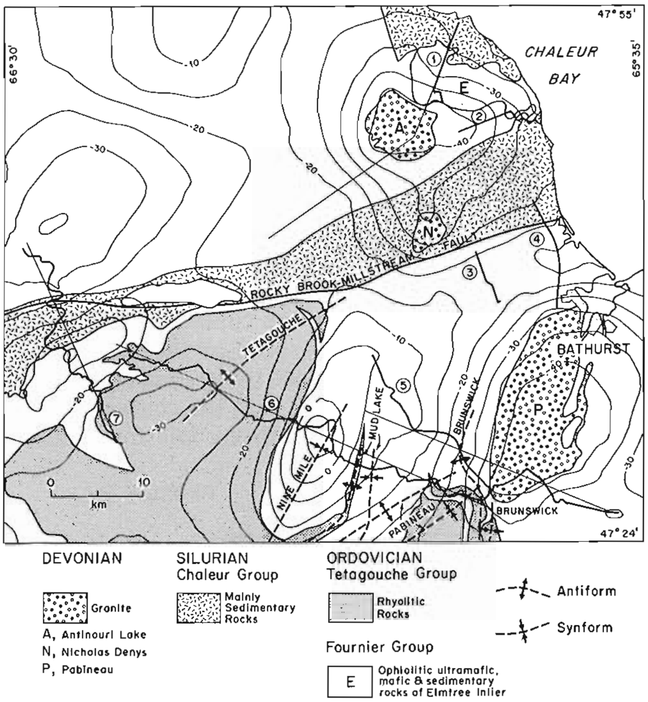

Figure 2. Geological map simplified from Figure 1 with addition of major antiforms and synforms (van Staal, 1986) and with Bouguer gravity anomaly contours (interval 5 mGal) superimposed. Heavy lines labelled 1 through 7 are detailed gravity traverses; accompanying fine lines are positions of corresponding profiles as described in caption for Figure 1.

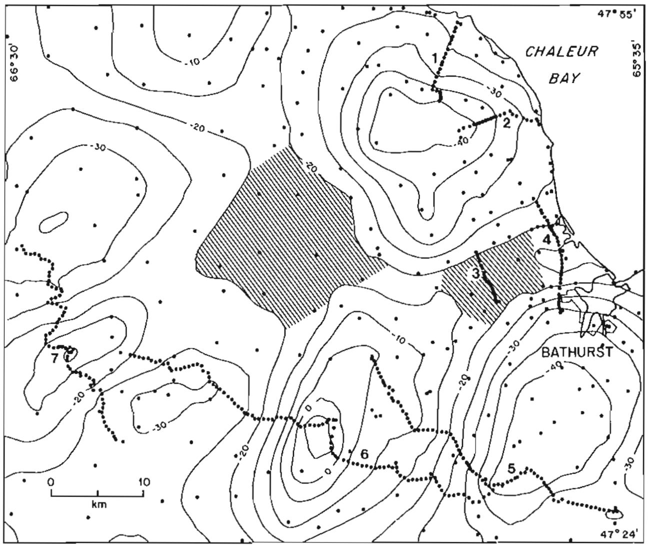

Figure 3. Bouguer gravity anomaly map contoured at a 5 mGal interval, based on pre-1990 regional gravity data and the data obtained along the detailed gravity traverses surveyed in 1990. Station locations indicated by dots. Shaded areas are taken as the "background" gravity field, with which anomalous regions are compared.

with a geological unit or package. The names of the anomalies, their wavelength, amplitude and interpreted geological source are tabulated in Table 1.

Antinouri Lake Low

This roughly circular negative anomaly is centred on the Antinouri Lake granite. A southward deflection of the −35 mGal contour, in particular, and of other contours towards and around the Nicholas Denys granite is noted. These granites, and possibly other buried granites, are probably linked at relatively shallow depth below the region and give rise to the anomaly. The geochemistry of the two granites differ (J. Whalen, pers. comm., 1990) so they may well represent different phases of a composite intrusion. The gravity field, defined essentially by regional data available prior to the 1990 survey, shows the flanks of the anomaly to be relatively smooth and undisturbed by any marked perturbations which could be attributed to a particular geological source.

With respect to the Fournier Group of the Elmtree inlier it is somewhat surprising that the presumed relatively dense, gabbroic and basaltic rocks, which comprise much of the group, do not produce a sizable positive signature. Fyffe and Noble (1985) argued that "A lack of an associated gravity anomaly over the Fournier Group and its intrusion by a stock of Devonian granite suggest that the complex is allochthonous . . .". Detailed gravity traverse 1, striking north-northeast, was run across the western part of the group to determine whether a possible signal had been missed by the coarse spacing of regional data. Observations were made at intervals of 0.5 km, which compares with spacings of about 5 and 2.5 km for regional data along the same line. Unfortunately, lack of suitable access prevented observations being made across the southwest margin of the group.

Table 1. Principal gravity anomalies

Anomaly	Wavelength (km)	Amplitude (mGal)	Geological Source(s)
Antinouri Lake low	30	−25	Antinouri Lake and Nicholas Denys granites
Pabineau low	30	−25	Pabineau granite
Nine Mile high	20	+25	Nine Mile synform
Tetagouche low	20	−10	Tetagouche antiform
Camel Back high	8	+10	mafic volcanics of the Fournier and Tetagouche Groups
Rocky Brook low	25	−15	rhyolitic rocks of Chaleur Group?

The data for traverse 1 are incorporated in profile 1 (Fig. 4), which extends southwest across the Antinouri Lake low. The profile varies smoothly across its entire length, and a significant perturbation attributable to the Fournier Group is not apparent. However, a small positive anomaly with an amplitude of about 4 mGal does correlate with mafic igneous components of the group. Using assumed mean densities of 3.00 and 2.75 g/cm^3 for these rocks and for surrounding sedimentary rocks, respectively, a rough estimate of the thickness of ophiolitic rocks is 400 m. Whether the base of the group represents a décollement or thrust, as per the allochthonous model suggested by Fyffe and Noble (1985), or an intrusive contact related to the Antinouri Lake granite is difficult to ascertain on the basis of the gravity data.

Traverse 2 crosses the Elmtree inlier from west to east, falling mainly along metasedimentary rocks of the Elmtree Formation. It was positioned to improve definition of the Antinouri Lake low and to investigate a northwest-trending fault that partially bounds the southwest margin of the Fournier Group. Observations were made at 300 m intervals across the fault and up to 700 m elsewhere. The corresponding profile shown in Figure 5 has the form of a simple gradient, and there is no signal associated with the fault. A small positive culmination with an amplitude of about 1 mGal coincides with the Deveraux gabbro at the eastern end of the profile, signifying that the gabbro is extremely thin in this location. This is consistent with the structural interpretations of van Staal and Langton (1990).

Before discussing other regional anomalies, two traverses (3 and 4) crossing Ordovician rocks between the Pabineau granite and the Rocky Brook-Millstream fault near Bathurst are described. These traverses were selected to determine whether gravity would be useful for investigating relatively fine structure within a complex fold-thrust tectonic package described by van Staal et al. (1988). Detailed ground magnetic traverses were carried out along the same routes (Thomas et al., 1991). The lithological and tectonic units in the area are quite narrow, ranging from about 200 m to 3.5 km in width.

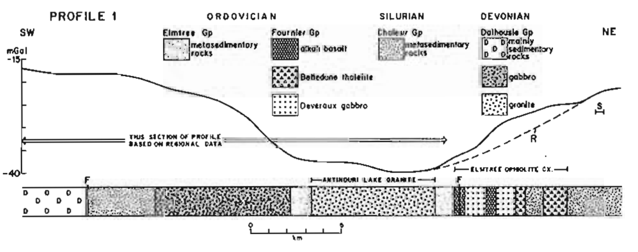

Figure 4. Gravity profile and geology along traverse 1. Geology based on Davies (1979) and van Staal and Langton (1990); F, fault. Detailed gravity data available only at northeast end of profile; horizontal bar labelled S indicates the spacing between gravity stations for the detailed portion. Dashed curve labelled R is estimated regional gravity trend; subtraction of this curve from the observed gravity profile indicates that a residual anomaly of about 4 mGal amplitude is associated with the Elmtree ophiolite complex.

Station spacing along traverse 3 ranges generally between 200 and 400 m, whereas that along the longer traverse 4 is generally 400 m.

Profiles for both traverses have the form of a broad positive signature (Figs. 6, 7), and attain peak values of about -15 mGal over a unit of alkali basalt within the Tetagouche Group (van Staal and Langton, 1990). Apart from this correlation there are no distinct perturbations of the curves that can be further investigated in terms of local geological structure. In profile 4, values drop off fairly rapidly southward from the contact between the alkali basalt and sedimentary rocks of the Miramichi Group. This feature may be related to the Pabineau granite, rather than being associated with the contact itself. Where profile 4 crosses the Rocky Brook-Millstream fault

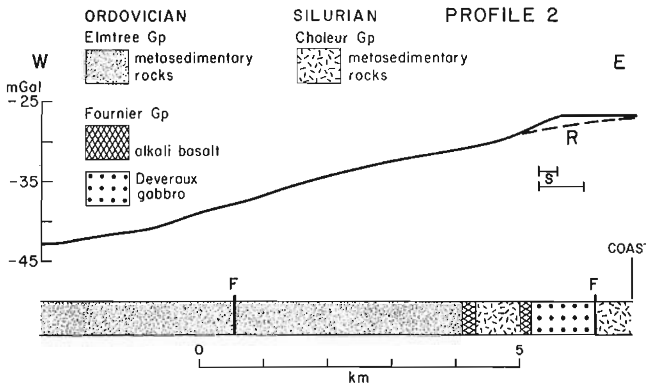

Figure 5. Gravity profile and geology along traverse 2. Geology based on van Staal and Langton (1990); F, fault. Horizontal bars labelled S indicate minimum and maximum spacings between gravity stations. Dashed curve labelled R is estimated regional gravity trend; subtraction of this curve from the observed profile indicates that a residual anomaly of about 1 mGal amplitude is associated with the Deveraux gabbro of the Elmtree ophiolite complex.

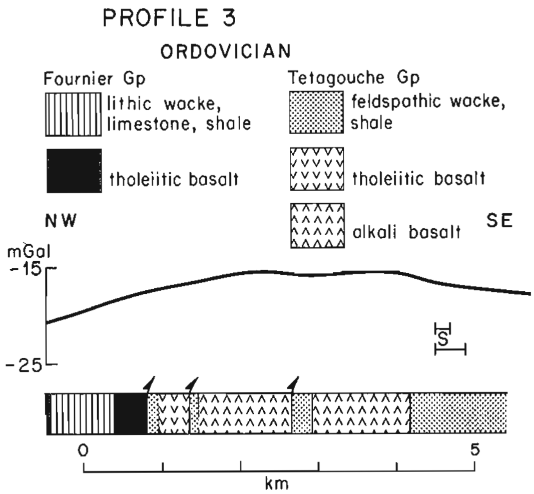

Figure 6. Gravity profile and geology along traverse 3. Geology based on van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. Horizontal bars labelled S indicate minimum and maximum spacings between gravity stations.

gravity values vary smoothly, indicating that the mean densities of sedimentary rocks of the Chaleur and Fournier Groups, which are in contact across the fault, are very similar.

Pabineau Low

The Pabineau low is centred on the Pabineau granite, and strikes north-northeast in sympathy with the trend of the granite, which is therefore interpreted as the cause of the low. The low, which extends beyond the margin of the granite in all directions, indicates its subsurface distribution. If the −20 mGal contour is a reliable estimate of the level of the background field, the granite extends laterally to the vicinity of this contour. Gravity traverse 5 was positioned to investigate the western margin of the granite and to better define the steep gradient mutual to the Pabineau low and the adjacent Nine Mile gravity high to the west. Observations were spaced 0.5 km apart.

The corresponding profile (Fig. 8), is dominated by the gradient, which is more or less constant across its width. Although the gradient crosses sedimentary and volcanic units, there are no prominent perturbations that correlate with them. However, construction of a regional trend (Fig. 8) indicates that a small positive anomaly with an amplitude of about 3 mGal is associated with basaltic rocks at the western end of the profile. The profile attains a minimum value of about −45 mGal over the Pabineau granite, before abruptly rising eastward across sedimentary rocks of the Miramichi Group to attain a local peak at a level of about −32 mGal. The eastern gradient may signify an eastward increase in density of the sedimentary rocks, the presence of a buried, relatively dense body, such as a mafic igneous unit, or a fault downthrowing the upper surface of the granite on its eastern side.

Nine Mile High

The Nine Mile gravity high correlates closely with alkali and tholeiitic basalts of the Fournier and Tetagouche Groups occurring in the core of the Nine Mile synform (van Staal, 1986). The axis of this elliptical anomaly coincides approximately with the axis of the synform (Fig. 9). Traverse 6 was selected to better define both this anomaly and the Tetagouche low to the west, and to determine whether individual geological units produce significant gravity signatures. Observations were made at intervals of 0.5 km.

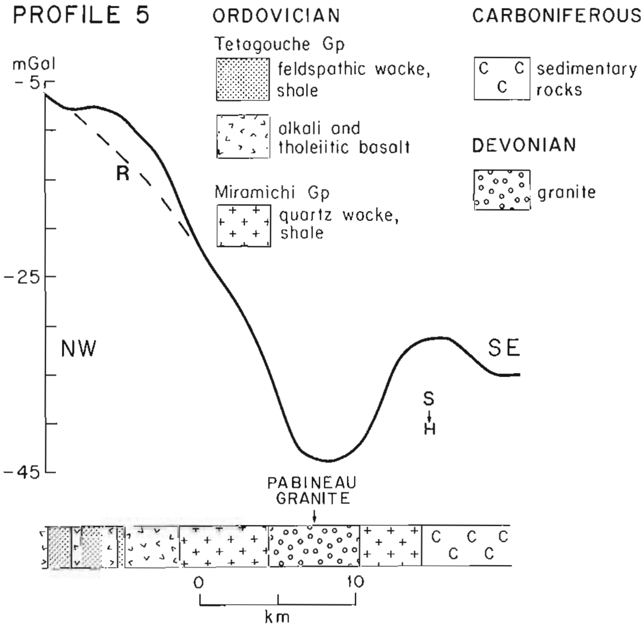

Figure 8. Gravity profile and geology along traverse 5. Geology based on van Staal and Langton (1990). Horizontal bar labelled S indicates spacing between gravity stations. Dashed curve labelled R is estimated regional gravity trend; subtraction of this curve from the observed profile indicates that a residual anomaly of about 3 mGal amplitude is associated with basaltic rocks of the Tetagouche Group.

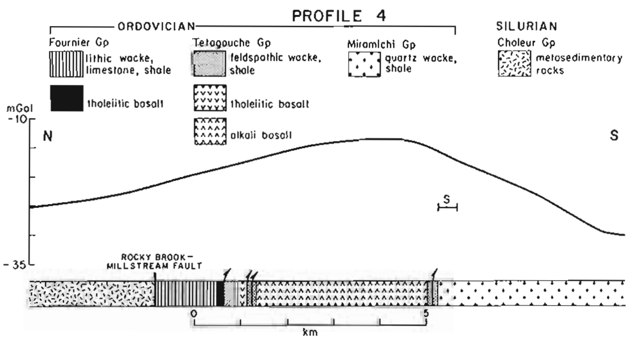

Figure 7. Gravity profile and geology along traverse 4. Geology based on van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. Horizontal bar labelled S indicates spacing between gravity stations.

Profile 6 reveals that a relatively broad development of basaltic rocks in the core of the synform can be correlated with the sharp peak of the Nine Mile high. Apart from this relationship, geology elsewhere apparently has little influence on the gravity field, although two minor, positive perturbations between the Mud Lake and Pabineau folds coincide with narrow bands of basaltic rocks.

An interesting characteristic of the profile is that it begins to rise eastward towards the peak from well within the outcrop of rhyolitic rocks to the west, suggesting that dense mafic volcanics underlie or are interbedded with the rhyolites in this region.

Tetagouche Low

The Tetagouche low is a relatively small anomaly falling within a broad expanse of rhyolitic rocks and centred essentially just west of the axis of the Tetagouche antiform (Fig. 2). The axis of the low is subparallel to that of the antiform. As previously noted, gravity values increase eastward across the rhyolitic unit towards the Nine Mile high. To a lesser degree they also increase westward across rhyolitic rocks towards the mafic volcanic assemblages near Camel Back Mountain. Again this implies that denser mafic volcanics underlie or are interbedded with the rhyolite at depth.

Camel Back High

The Camel Back high is a small, yet distinct anomaly whose axis lies over a tectonic package of predominantly mafic volcanic rocks of the Tetagouche and Fournier groups. The anomaly itself spans both these volcanic rocks and sedimentary rocks of the Silurian Chaleur Group, which are juxtaposed to the northwest. Gravity traverse 7 runs northwest across the anomaly; observations were made generally at 0.5 km intervals. The corresponding profile is illustrated in Figure 10.

Rhyolitic rocks occur along the southeast section of the traverse, and here the gravity field is fairly flat. It begins to rise northwestward from a point near the contact of the rhyolite with alkali basalts of the Tetagouche Group. It peaks over alkali basalts immediately south of the boundary with sedimentary rocks of the Chaleur Group; the basalts are the probable source of the anomaly. The field then diminishes northwestward into the area of the Rocky Brook low. The fact that the high extends over the Chaleur Group suggests that this sedimentary assemblage may be underlain by dense igneous rocks.

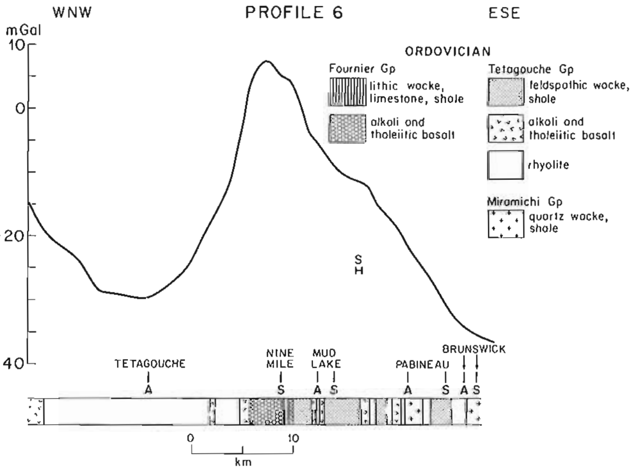

Figure 9. Gravity profile and geology along traverse 6. Geology based on van Staal and Langton (1990). Axes of antiforms (A) and synforms (S) are indicated. Horizontal bar labelled S indicates spacing between gravity stations.

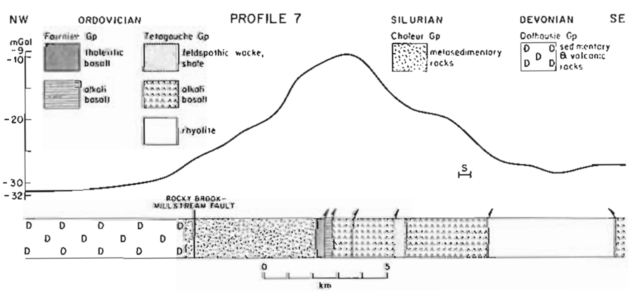

Figure 10. Gravity profile and geology along traverse 7. Geology after Davies (1979) and van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. Horizontal bar labelled S indicates spacing between gravity stations.

Rocky Brook Low

As the Tetagouche high gives way northwestward to the Rocky Brook low, the gravity field continues to decrease across sedimentary rocks of the Chaleur Group to a point just north of the Rocky Brook-Millstream fault, where it becomes flat again over Devonian sedimentary and volcanic rocks. The fault itself is not associated with a gravity signature. The source of the low may be the extensive developments of rhyolitic rocks of the Chaleur Group, which occur immediately northwest of the Devonian rocks.

FUTURE PLANS

The new, detailed gravity data have more precisely defined several anomalies associated with discrete geological units. This improved resolution will benefit model studies which will be carried out in the near future. An integral part of the gravity investigations was the collection of representative rock samples from the area. These will be used for measurements of rock densities, which along with structural data from ongoing studies by C. van Staal will be used to constrain gravity models. It is anticipated that these gravity studies will lead to a clearer picture of the third dimension of the crust in this region.

ACKNOWLEDGMENTS

We are most grateful to Cees van Staal for a thorough review of this report, and for his many suggestions, particularly with respect to the geology of the region. We thank also Warner Miles for providing various computer-generated gravity maps and profiles, and Lorne Campbell and Azad Rafeek for drafting the figures.

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A correlation technique for separating natural and man-made airborne gamma-ray spectra

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Grasty, R.L., and Multala, J. A correlation technique for separating natural and man-made airborne gamma-ray spectra; in Current Research, Part D, Geological Survey of Canada, Paper 90-1D p. 111-116, 1991.

Abstract

Analysis of airborne gamma-ray data collected in Finland following the Chernobyl nuclear accident has shown that cesium-134 and cesium-137 spectra can be successfully separated from natural gamma-radiation due to potassium-40 and decay products in the uranium and thorium series. In this technique, the natural gamma-ray contribution to the low energy region containing both man-made and natural radiation, was determined directly from the airborne measurements by a correlation technique and required no prior knowledge of the natural gamma-ray spectral shapes.

Résumé

L'analyse de données aérogammamétriques recueillies en Finlande à la suite de l'accident nucléaire de Chernobyl a montré qu'on peut réussir à séparer les spectres du césium 134 et du césium 137 du rayonnement gamma naturel attribuable au potassium 40 et aux produits de désintégration des familles de l'uranium et du thorium. Cette méthode part du principe selon lequel l'apport du rayonnement gamma naturel à la plage de faible énergie regroupant les rayonnements naturels et artificiels a été directement déterminé à partir des mesures aériennes à l'aide d'une méthode de corrélation n'exigeant aucune connaissance préalable des formes spectrales du rayonnement gamma naturel.

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INTRODUCTION

Following a nuclear reactor accident, such as the Chernobyl event of 1986, many man-made radioactive nuclides with a variety of gamma-ray energies and half-lives may contaminate the environment. The most significant gamma-ray emitting nuclides that may be present following such an accident are shown in Table 1 together with their gamma-ray energies and half-lives. The most abundant natural gamma-ray emitters from potassium-40 and decay products in the uranium and thorium series are also shown. The International Atomic Energy Agency has recommended three energy windows to monitor potassium-40, and decay products in the uranium and thorium series (IAEA, 1976). These energy windows which are shown in Table 2 are well above the energies of almost all man-made gamma-ray emitting nuclides.

To estimate the significance of any man-made contamination, the contribution due to natural gamma-radiation must first be removed. Grasty (1985) has described a spectral fitting method for separating man-made and natural gamma-ray spectra. However, the potassium, uranium and thorium spectral shapes must first be measured on large radioactive calibration pads and are specific to the detector system measured. If the detector system is changed or individual detectors drift relative to each other, the model spectra may no longer be applicable.

In 1989, analysis of airborne gamma-ray data collected in 1988 by the Geological Survey of Finland following the Chernobyl nuclear accident showed that the natural gamma-ray component could be removed from the airborne spectrum without prior knowledge of the gamma-ray spectral shapes of potassium, uranium and thorium. The contribution of the standard high energy potassium, uranium and thorium windows to the low energy region containing both man-made and natural radiation was first determined by a simple correlation technique. Once these contributions were known, the contribution of natural gamma-radiation to the low energy part of the gamma-ray spectrum could be removed.

Table 1. Principal man-made and natural radioactive nuclides

Nuclide	Principal Energies (keV)	Half-life* (days)
Man-made:		
Zr-95	724, 756	65
Nb-95	765	35
Mo-99	740	3
Ru-103	497	40
Ru-106	512	368
I-131	364	8
Te-132	230	3
Cs-134	605, 795	730
CS-137	662	11000
Ba-140/La-140	1596	13
Natural:		
K-40	1460	
Uranium Series		
Pb-214	352	
Bi-214	609, 1120, 1765	
Thorium Series		
Ac-228	911, 969	
Tl-208	583, 2614	

*Half-lives of natural emitters are not presented.

THEORY

In the correlation technique there are two basic assumptions. The first assumption is that the gamma-ray energies due to the fall-out are below the potassium, uranium and thorium windows normally used to monitor the concentrations of these elements on the ground (Table 2). Table 1 shows that lanthanum-140 with a gamma-ray peak at 1596 keV is the only man-made nuclide with an energy falling in the standard potassium, uranium and thorium windows. However, it has a half-life of 13 days (Table 1). Since our analyses were carried out on data collected two years after the Chernobyl accident, the lanthanum peak will be negligible.

The second assumption is that there is no correlation between fall-out and the potassium, uranium and thorium concentration of the ground. This assumption is expected to be valid since fallout deposition is controlled by atmospheric conditions and not by the radioactivity of the ground.

The count rate in any channel depends linearly on the ground concentrations of potassium, uranium and thorium. In addition, there will be a contribution in each channel due to fallout. There is also a linear relationship between the potassium, uranium and thorium window count rates shown in Table 2 and the ground concentrations of the three radioactive elements (IAEA, 1976). Consequently, there is a linear relationship between the potassium, uranium and thorium window count rates (R_K, R_U and R_T) and the count rate R(i) in channel i, given by:

$$R(i) = a_1(i) \times R_K + a_2(i) \times R_U + a_3(i) \times R_T + C(i) \quad \ldots . (1)$$

where $a_1(i)$, $a_2(i)$ and $a_3(i)$ are constants which give the count rate in channel i per unit count rate in the potassium, uranium and thorium windows and C(i) represents the contribution in channel i due to fallout. In this equation, it is assumed that the overwater background has been removed.

From flights over areas with different concentration of potassium, uranium and thorium, the four unknown constants ($a_1(i)$, $a_2(i)$, $a_3(i)$ and C(i)) in equation (2) can be determined by a simple linear regression for each chan-

Table 2. Spectral windows used to measure gamma-rays

Element Analyzed	Isotope Used	Gamma-ray Energy (keV)	Energy Window (keV)
Potassium	K-40	1460	1370 - 1570
Uranium	Bi-214	1765	1660 - 1860
Thorium	Tl-208	2614	2410 - 2810

nel of the spectrum. The value of C(i) will be the average contribution to fallout in channel (i) along the flight line that is analyzed. Once these coefficients have been determined, the contribution of potassium, uranium and thorium to any channel in the gamma-ray spectrum can be calculated by monitoring the potassium, uranium and thorium windows and multiplying them by the appropriate constants.

The same procedure can be applied not only to individual channels but also to a window covering a range of channels. However, in this case no information will be available as to the shape of the fallout spectrum which can be obtained from the calculated values of C(i) for each channel.

DEMONSTRATION

In 1989, the correlation technique was tested on airborne gamma-ray spectra acquired in 1988 by the Finnish Geological Survey over two areas in Finland which were contaminated by fall-out from Chernobyl. The two areas selected for study were in the Kokemaki area of southwestern Finland and in the Kaipola area of south central Finland. The Kokemaki area was moderately contaminated with a Cs-137 deposition from 11 to 45 kBq/m^2 and with an external dose rate from 0.028 — 0.11 μSv/h as measured by the Finnish Centre for Radiation and Nuclear Safety (Arvela et al, 1989). The Kaipola area was found to be considerably more contaminated with a Cs-137 deposition from 23 to 78 kBq/m^2 and an external dose rate from 0.056 to 0.19 μSv/h.

The Finnish Geological Survey gamma-ray spectrometer system has a total crystal volume of 25 litres of sodium-iodide and records 120 channels of gamma-ray data every second with each channel being approximately 25 keV wide. The flights were carried out at a nominal survey altitude of 30m (100 feet). Figures 1 and 2 are the average gamma-ray spectra from flights carried out in the two areas and show the higher contamination in the Kaipola area. The Cs-137 gamma-ray peak at 662 keV and the Cs-134 gamma-ray peak at 795 keV are clearly evident on the spectrum from the Kaipola area. Data from survey flights in both areas were analyzed. However, because of the large quantity of data collected on a single flight, the analyses were carried out on flight line sections no longer than 2000 seconds in length, corresponding to distances up to approximately 100km.

A background spectrum was first calculated from measurements over a lake which was located by visual inspection of the total count profile. This total count profile was the contribution of all counts between channels 10 and 111. The "Lake" spectrum was then subtracted from the spectrum recorded at each data point. The background-subtracted potassium, uranium and thorium windows as indicated in Figure 1 were then calculated for each one second data point and computer fitted by simple linear regression to equation (1) for all channels below the potassium window.

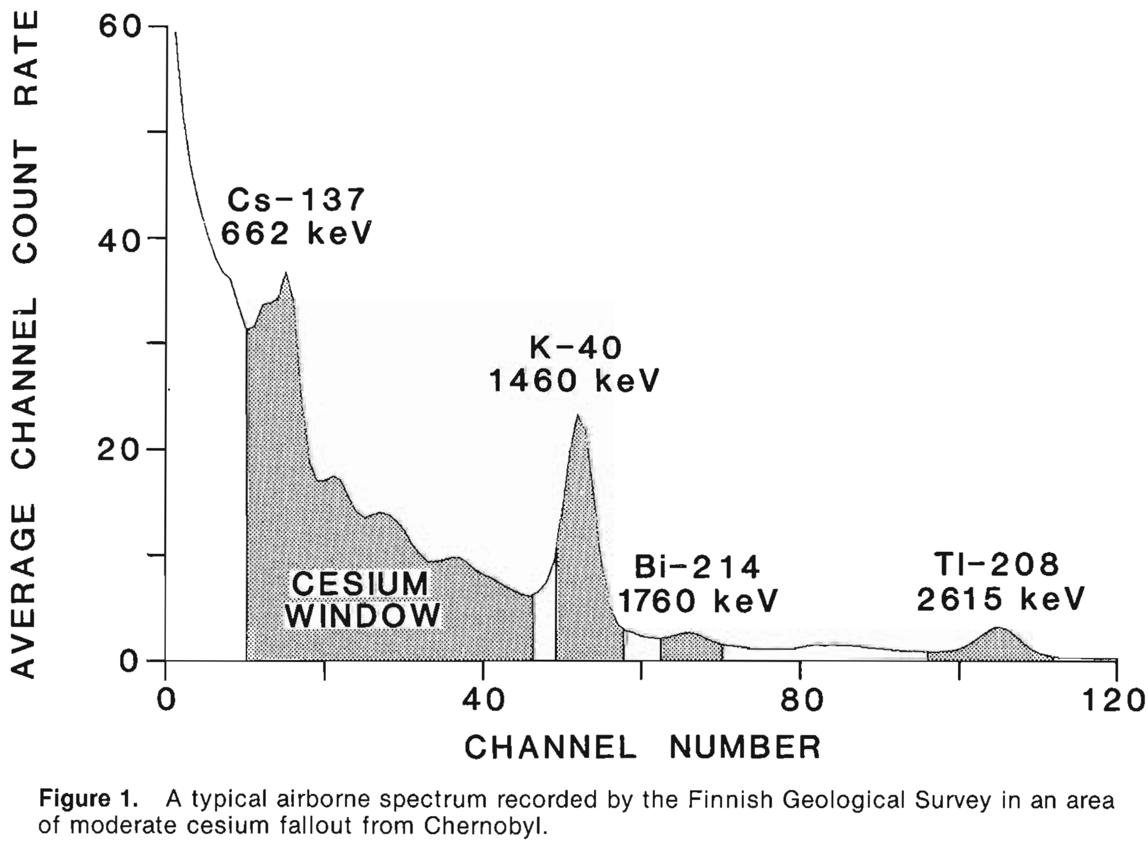

Figure 1. A typical airborne spectrum recorded by the Finnish Geological Survey in an area of moderate cesium fallout from Chernobyl.

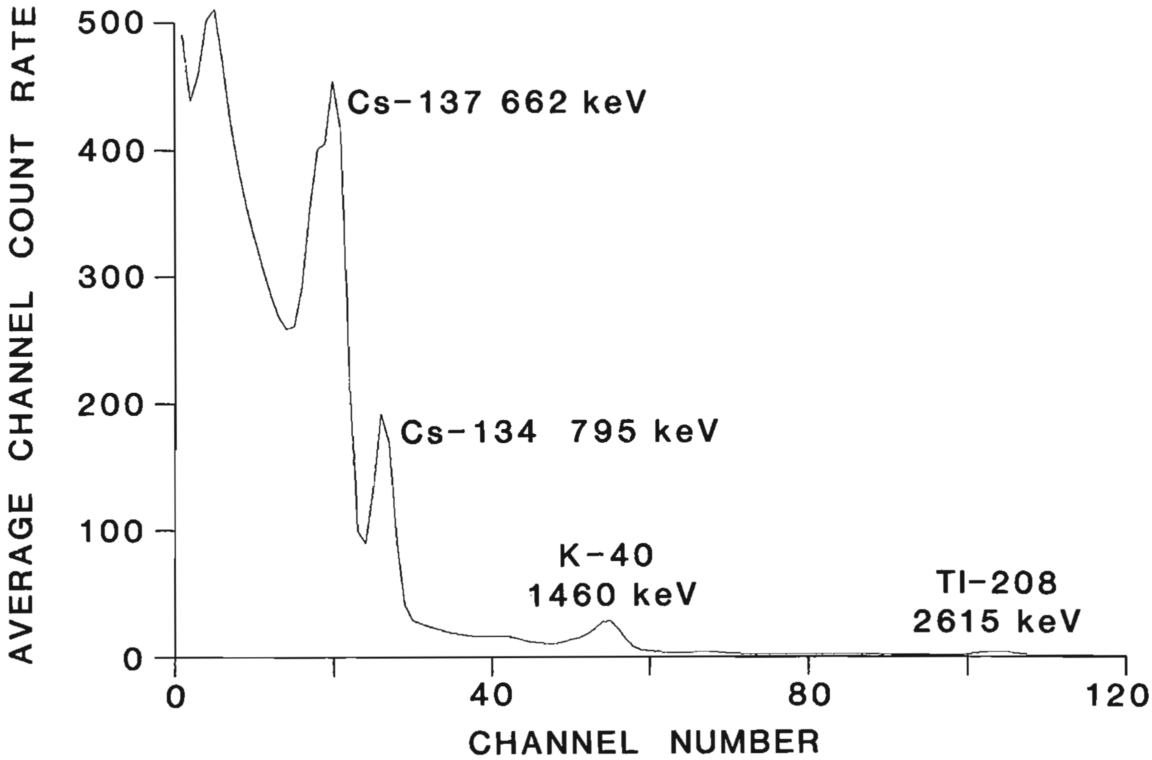

Figure 2. A typical airborne spectrum recorded by the Finnish Geological Survey in an area with high cesium fallout from Chernobyl.

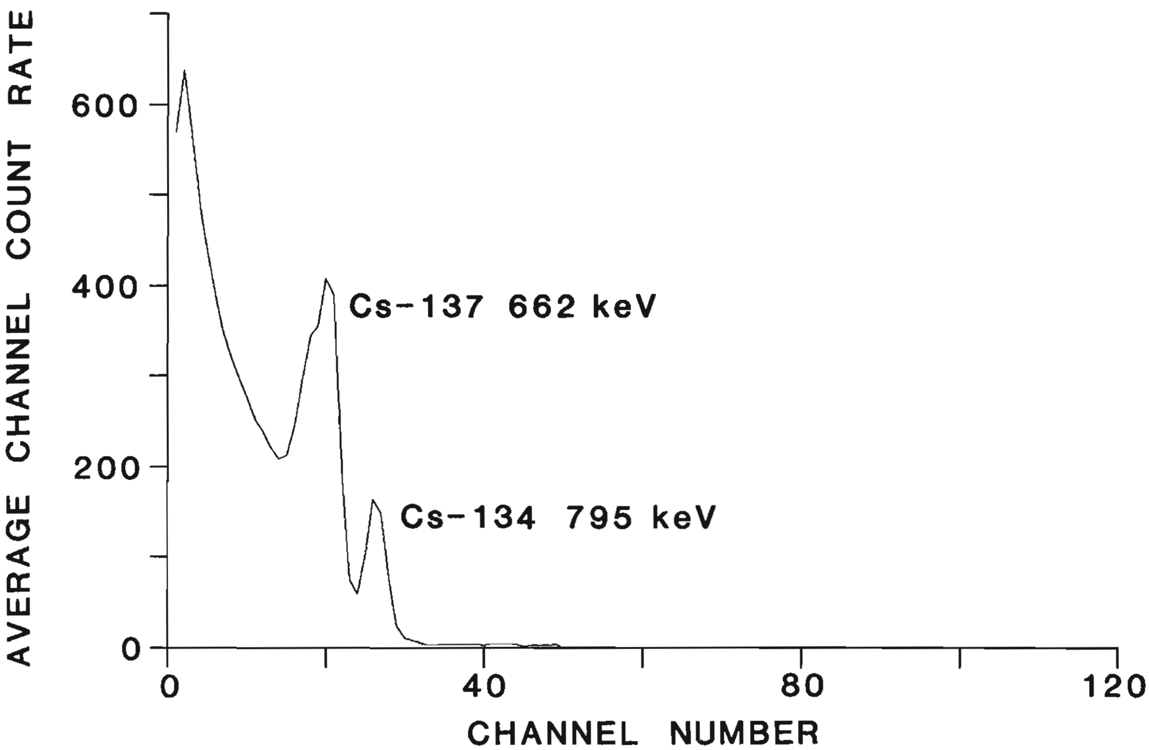

Figure 3. The computed Cs-134 and 137 spectra extracted from a gamma-ray spectrum in the high fallout region.

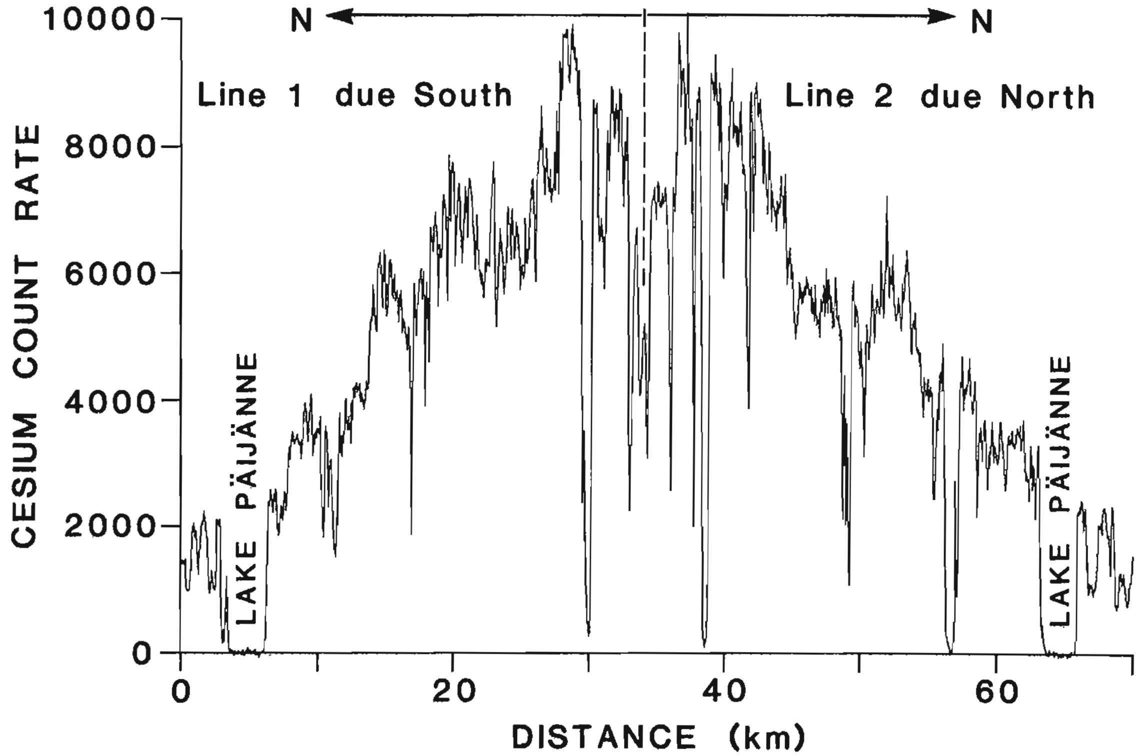

Figure 4. A profile flown in two directions in the area of high cesium comtamination showing the calculated cesium count-rate.

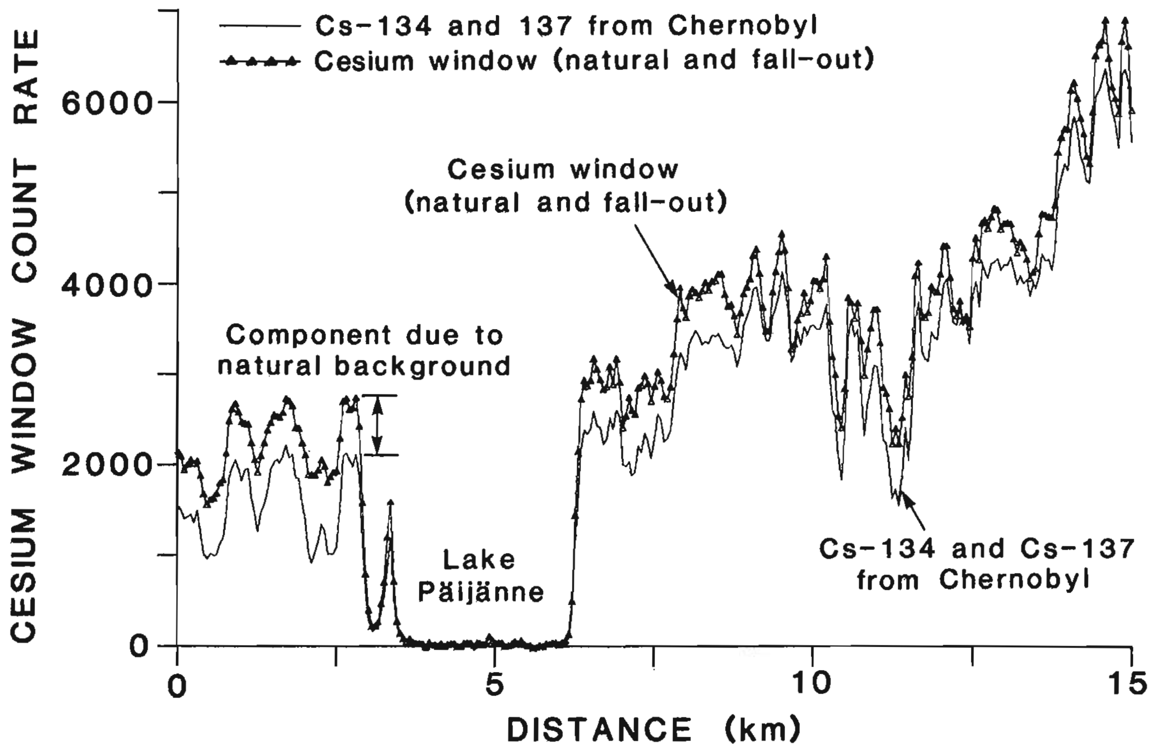

Figure 5. A calculated cesium profile in the area of high contamination showing the K, U and Th contribution to the cesium window.

Table 1 shows that provided sufficient time has elapsed for lanthanum-140 to decay, the maximum gamma-ray energy due to fallout originates from Cs-134 with an energy of 795 keV. Figure 3 shows a plot of the C(i) coefficients from 800 seconds of data collected in the Kaipola area. Above the Cs-134 gamma-ray peak, the C(i) coefficients are extremely low, demonstrating that the technique has successfully removed the contribution of potassium, uranium and thorium to the low energy part of the gamma-ray spectrum.

Once the technique had been demonstrated by production of the fallout spectrum, some flight line profiles from the same area were studied to investigate the contributions of cesium fallout and natural gamma-radiation to the observed counts in a wide cesium window.

Figure 4 shows two flight line profiles flown in opposite directions in the Kaipola area. The profiles show the calculated contribution of cesium fallout in a cesium window covering channels 10 to 47. The two flight lines were flown approximately 1200 m apart and show significantly increasing fallout from north to south. Part of this profile is shown again in Figure 5. This figure shows the total observed count rate in the wide cesium window from fallout and natural gamma-radiation to enable comparison with the contribution from fallout. As can be seen, the contribution from natural gamma-radiation stays relatively constant but the fallout contribution increases considerably along the flight line.

SUMMARY AND CONCLUSIONS

A simple procedure has been successfully tested on airborne gamma-ray spectra to separate man-made sources of radiation from those produced naturally from potassium-40 and decay products in the uranium and thorium series. The technique relies on the fact that almost all gamma-radiation from man-made sources resulting from nuclear accidents or atomic weapons fallout has a range of energies which is much lower than those produced naturally. By monitoring the high energy region due to natural gamma-radiation the low energy contribution from natural sources of gamma radiation can be removed from the measured spectrum.

The successful application of the technique requires a knowledge of the potassium, uranium and thorium gamma-ray spectral shapes. The gamma-ray spectral shapes were determined directly from airborne measurements by correlating three high energy windows for potasssium, uranium and thorium with the low energy region containing both man-made and natural radiation.

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A progress report on the structural control of gold mineralizations in the Cape Breton Highlands[1]

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Abstract

Two types of gold-bearing mineralization have been recognized in the Cape Breton Highlands. The first type consists of stratabound sulphide mineralization which predates deformation and metamorphism of their host rocks (e.g. Chéticamp area). They occur chiefly in the Aspy terrane. The other type is associated with quartz (± calcite) veins and occurs in both Aspy and Bras d'Or terranes (e.g. INCO, St. Anns area, Second Gold Brook). Available geochronological data suggest that stratabound mineralization was formed and remobilized at 430-400 Ma, hence their hosting structures may have formed during Acadian terrane amalgamation. In contrast, syntectonic veins at St. Anns may be as old as 517 Ma. Across Cape Breton Highlands, there is a gradual change from medium-to high-grade ductile conditions for mineralizations in the NW to low grade brittle fault controlled occurrences on the east coast. The geometry of the contractional tectonic regime responsible for structures hosting mineralization displays a similar variation: NNE-SSW to N-S contraction is apparent in the western Aspy terrane, whereas NW-SE oriented contraction is more likely farther east.

Résumé

Deux types de minéralisations aurifères ont été identifiées dans les hautes terres du Cap-Breton. Une première consiste en sulfures stratiformes antérieurs à la déformation et au métamorphisme de la roche hôte (p. ex. région de Chéticamp). On trouve les minéralisations de ce type principalement dans le terrane d'Aspy. Les minéralisations du deuxième type sont associées à des veines de quartz (± calcite) dans les terranes d'Aspy et de Bras d'Or (p. ex. INCO, région de St. Anns, Second Gold Brook). Les données géochronologiques disponibles semblent indiquer que les minéraux stratiformes ont été formés et remobilisés il y a 430 à 400 Ma et que leurs structures hôtes ont par conséquent pu se former pendant l'amalgamation du terrane d'Acadian. D'autre part, les veines syntectoniques à St. Anns peuvent avoir jusqu'à 570 Ma. Dans les hautes terres du Cap-Breton, des teneurs moyennes à élevées de minéralisation se manifestent dans des roches ductiles au nord-ouest, passant progressivement à des teneurs faibles dans des zones de roches cassantes contrôlées par des failles sur la côte est. La géométrie du régime tectonique de contraction responsable de la minéralisation des structures présente une variation semblable; une contraction NNE-SSO à N-S est apparente dans le terrane d'Aspy, alors que, plus loin à l'est, une contraction orientée NO-SE est plus vraisemblable.

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INTRODUCTION

The recent discovery of gold deposits in Newfoundland has led to an increased awareness of the potential for similar occurrences elsewhere in the Canadian Appalachians. A project aimed at investigating the relationships between major structures and gold occurrences on a regional scale in the Atlantic Provinces was initiated in 1989 with studies in Newfoundland (Dubé, 1990; Dubé at al., 1991), New Brunswick (Tremblay and Dubé, 1991) and Nova Scotia (this study) (Fig. 1).

This project "*Au and lineaments*", is concerned with known Au occurrences in the Cape Breton Highlands, Nova Scotia (Fig. 2). Through studies on appropriate scales the objective is to establish links between major regional structural features and known gold occurrences. To this end, studies of metallogeny and structural features were carried out at a number of gold occurrences, and detailed studies were conducted on the S-SW extension of the Eastern Highlands shear zone, which separates the Aspy and Bras d'Or terranes (Fig. 2).

REGIONAL GEOLOGY AND TECTONIC FRAMEWORK

Barr and Raeside (1986, 1989) and Barr et al. (1987a) divided Cape Breton Island into four terranes, each with distinct pre-Carboniferous lithological, geochronological, metamorphic and structural characteristics (Fig. 2). The terranes are separated by major fault zones.

The Blair River Complex (BRC) comprises variably retrogressed high grade gneisses and minor calcareous rocks. These rocks are intruded by a plutonic suite, including anorthosite, monzodiorite, syenite and granite (Raeside et al., 1986). U-Pb dating of zircons from the syenite yielded an age of 1040^{+18}_{-40} Ma, which is interpreted to represent a Grenvillian metamorphic overprint (Barr et al., 1987b).

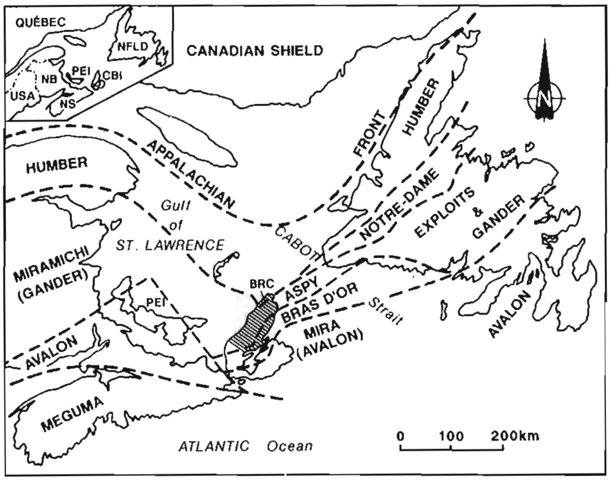

Figure 1. Tectonostratigraphic subdivisions in the Appalachians of Atlantic Canada (after Barr and Raeside, 1989). Ruled area is the Cape Breton Highlands, shown in Figure 2. Abbreviations: NFLD — Newfoundland, CBI — Cape Breton Island, PEI — Prince Edward Island, NB — New Brunswick, NS — Nova Scotia, BRC — Blair River Complex.

To the east and south, the BRC is separated from the Aspy terrane by the Wilkie Brook and Red River fault zones, respectively. The zones are composed of variably retrogressed mylonites, and are up to 1.3 km wide (Barr and Raeside, 1989).

The Aspy terrane comprises low- to high-grade metamorphic rocks intruded by extensive suites of mainly Devonian granitoids. The metamorphic rocks are late Ordovician to early Silurian, and include the Jumping Brook metamorphic suite, the Sarach Brook metavolcanics and the Money Point Group, which host some of the Au-mineralization described below.

The Bras d'Or terrane docked with the Aspy terrane along the Eastern Highlands shear zone during the late Silurian and Devonian (Reynolds et al., 1989; Dunning et al., 1990). The terrane is made up of late Proterozoic sedimentary rocks intruded by late Proterozoic to Cambrian dioritic to granitic plutons (Barr and Raeside, 1989; Dunning et al., 1990).

The Eastern Highlands shear zone (EHSZ) (Barr et al., 1987a; Barr and Raeside, 1989; Lin, 1990) separates the Aspy and Bras d'Or terranes. The shear zone is up to 0.8 km wide and is mainly characterized by mylonites showing a complex metamorphic and structural history. Along its northeastern extension (Fig. 2) it diverges into a number of low angle fault splays (Barr et al., 1987a; Barr and Raeside, 1989), and is cut by Devonian plutons (Dunning et al., 1990).

The Mira terrane in southeastern Cape Breton Island is composed of late Precambrian volcanics occurring in fault bounded blocks. A series of dioritic to granitic plutons intrude the volcanics and are in turn overlain by a sequence of late Precambrian to Cambrian supracrustal rocks (Barr and Raeside, 1989).

GOLD OCCURRENCES

Two types of gold occurrences have been recognized on Cape Breton Island. The first type consists of stratabound sulphides which predate deformation and metamorphism in the host rocks. Sulphides are abundant, and include arsenopyrite, chalcopyrite, pyrite, pyrrhotite, galena and sphalerite. This type occurs in the Aspy terrane (e.g. Chéticamp area) and along its eastern margin (Fig. 2). The second type of mineralization is associated with quartz (± calcite) veins, and occurs in both the Aspy and Bras d'Or terranes (INCO, St. Anns area, Second Gold Brook).

In text, tables and figures below, numbers in brackets refer to locations of occurrences (Fig. 2). Unless otherwise noted, all geochemical data are from Sangster (unpublished analyses) and the description of quartz veins follows the nomenclature of Hodgson (1989). Representative sulphide assemblages and microstructures are shown in Figure 6.

Stratabound sulphide occurrences

Gold mineralizations in the Chéticamp area are hosted by metavolcanics and clastic metasediments of the Jumping Brook metamorphic suite (JBMS). The

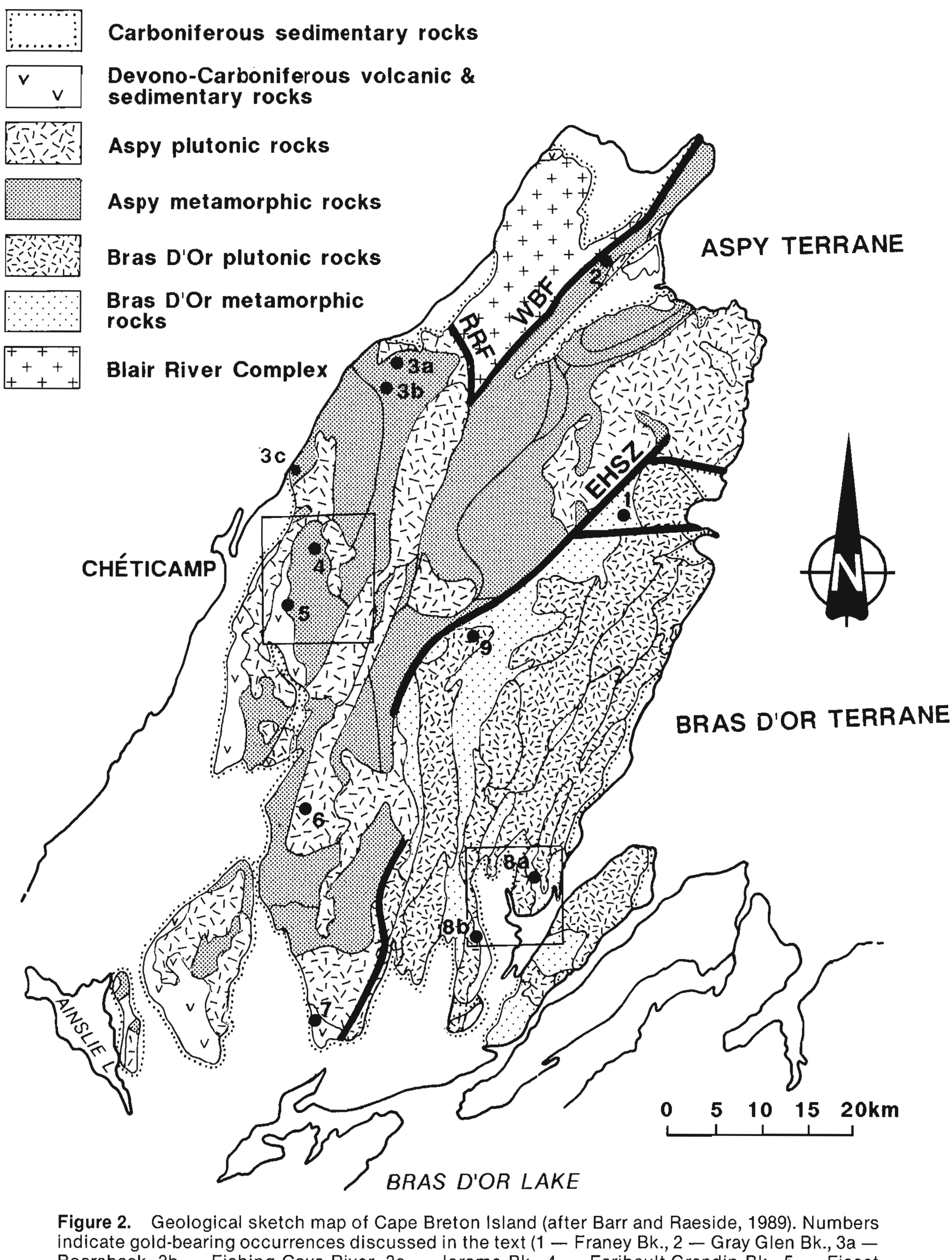

Figure 2. Geological sketch map of Cape Breton Island (after Barr and Raeside, 1989). Numbers indicate gold-bearing occurrences discussed in the text (1 — Franey Bk., 2 — Gray Glen Bk., 3a — Boarsback, 3b — Fishing Cove River, 3c — Jerome Bk., 4 — Faribault-Grandin Bk., 5 — Fisset Bk., 6 — Nile Bk., 7 — Second Gold Bk., 8a — St. Anns area, 8b — Goose Cove Bk., 9 — INCO). Boxes indicate location of index maps for the Chéticamp and St. Anns area (Fig. 3 and 7, respectively). Abbreviations: RRF — Red River fault zone, WBF — Wilkie Brook fault zone, EHSZ — Eastern Highlands shear zone.

occurrences include the Grandin Brook, Iron Cap Adit, Marleau, Mountain Top Adits, Grandin Cliff, Core Shack and Galena Mine showings. In the northwestern corner of the Aspy terrane, the Boarsback and Fishing Cove River occurrences were visited along the Cabot trail. A few other occurrences are known from trenches in the area, but these were not studied because trenches have been filled. The characteristics of selected occurrence is summarized in Table 1, and locations are shown in Figures 2 and 3.

JBMS comprise a lower tholeiitic metavolcanic suite (Faribault Brook metavolcanics), overlain by siliceous metasediments and felsic tuffs (Barren Brook schist), turbiditic metasediments (Dauphinee Brook schist), and the higher grade equivalents of these units (Corney Brook schist and Fishing Cove River schist) (Jamieson et al., 1990). All mineralization is hosted by quartz-rich schists and fine grained felsites belonging to the Barren Brook schist unit, except where otherwise indicated. Structural orientation data from occurrences along Grandin Brook are given in Figure 4a.

The mineralizations in the Faribault Brook-Grandin Brook-Fisset Brook area are generally similar, however, some important differences exist, and these may be interpreted in terms of the origin of the mineralization.

Along a N-S transect in the Faribault-Grandin Brook area, the orientation of foliations and lineations is constant (Fig. 4a), showing that the area, and more importantly, the mineralized zones, were uniformly effected by a regional structural event.

The two main variables are the relative proportions of sulphides and the thickness of the host sericite schist at each occurrence. Along the transect, there is a marked southward increase in thickness of the sericite schist and a concomitant change to more Cu-rich sulphide assemblages. If interpreted in terms of volcanic massive sulphide models (e.g. Lydon, 1989), the regional zonation would indicate that N to S variations correspond to changes between distal and proximal conditions, respectively. Similar suggestions regarding the syngenetic origin of mineralizations were made by Currie (1987) and Jamieson et al. (1989, 1990)

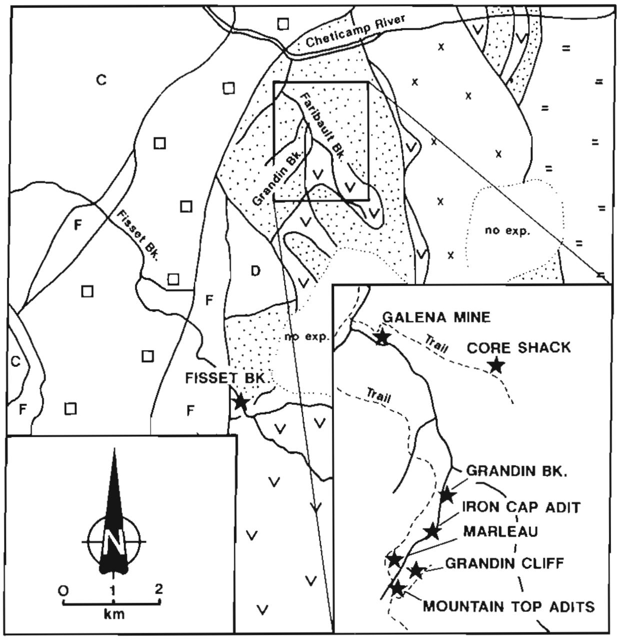

Figure 3. Location map of the Chéticamp River-Faribault Brook-Fisset Brook area, showing location of mineral occurrences (stars) and generalized geology (from Barr et al., in press). Legend: D — foliated diorite, boxes — Chéticamp pluton, v — mafic volcanics from Jumping Brook Metamorphic Suite (JBMS), dots — sediments of JBMS, = — Belle Côte Road orthogneiss, x — Salmon Pool Pluton, F — Fisset Brook Formation, C — Undivided Carboniferous sediments. See Figure 2 for location.

Table 1. Stratabound sulphide occurrences in the Aspy terrane

Occurrence	*Au-values*	*Sulfides (mode of occurrence)*	*Host rock (thickness)*	*Structure*
Grandin Bk.	-	py, aspy, cpy (str.)	gnt-sericite schist (few 10's of cm exposed)	
Iron Cap Adits	-	py, aspy, cpy (diss., str.)	chl-mica schist (Faribault Bk. metabasite)	S-SW directed shearing along N-NE dipping planes
Marleau	2.5 ppm 10.5 ppm	py, aspy, cpy, ga (diss., str.)	silicified sericite schist (2 m)	S-directed shearing along WNW-dipping planes
Mountain Top Adits	1.9 ppm	py, aspy, cpy, bo, cc (diss., str.)	silicified sericite schist (10-20 m)	
Grandin Cliffs	110 ppb	py, cpy (diss., str.) (mal, azu, chry)		S-directed shearing along gently N-dipping planes
Core Shack	4.8 ppm 5.4 ppm	py, aspy, cpy, sph (diss., str., vein)	greenish sericite schist (2-3 m)	
Galena Mine	4.8 ppm(*)	py, aspy, sph, ga (diss., str.)	gnt-sericite schist (few m)	S-directed shearing along gently N-dipping planes
Fisset Bk.		py, aspy, cpy, po, sph (diss., str., vein)	sericite schist (ca. 10 m)	S-W directed shearing
Boarsback (3a)	120 ppb	py, cpy (diss.)	gnt-bio-ky-schist (2-3 m)	SSW-directed shearing along moderately ENE-dipping planes
Fishing Cove River (3b)	17 ppb	as above	as above	

Mode of occurrence: diss. - disseminated, str. - thin foliation parallel layers, vein - discordant vein
Minerals: py - pyrite, aspy - arsenopyrite, cpy - chalcopyrite, ga - galena, bo - bornite, cc - chalcosite
mal - malachite, azu - azurite, po - pyrrhotite, chry - chrysocola
(*) from Currie (1987)

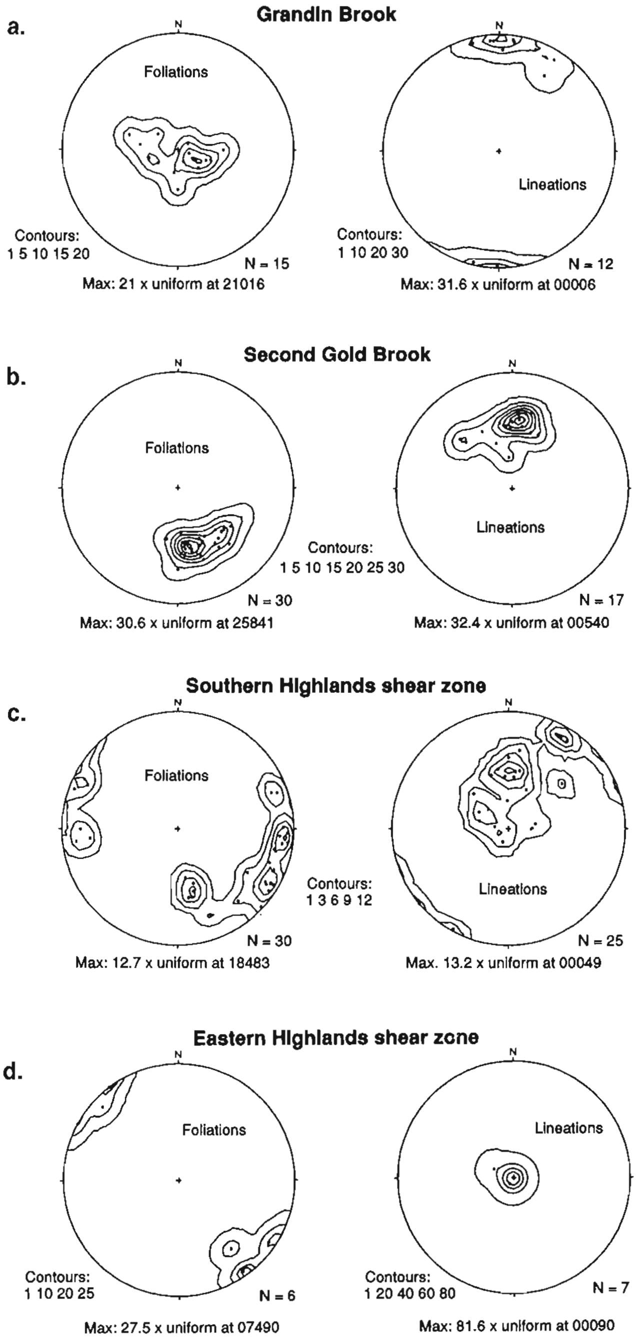

Figure 4. Structural orientation data.
(a) Mineralized lithologies along Grandin Brook (Fig. 3).
(b) Second Gold Brook area; data from a ca. 1 km NW-SE transect perpendicular to the regional strike.
(c) Southern Highlands shear zone; data mainly obtained in the southern half of the SHSZ (Fig. 5).
(d) Eastern Highlands shear zone; data from mylonite zone in NE corner of study area.
All diagrams are lower hemisphere; contour levels are in % per 1% area and contouring follows the weighted contouring method of Robin and Jowett (1986).

On outcrop and sample scale, the zonations observed at each of the occurrences can be attributed to both primary and secondary processes. The mineralogical variation from centre to margin of massive sulphide veins may be caused by temporal and spatial variations in the composition of mineralizing fluids (e.g. Sangster, 1980; Franklin et al., 1981; Large, 1983). However, in view of the intense deformation observed in host rocks, it is likely that synkinematic remobilization and recrystallization overprinted any primary zonation and caused the present distribution of sulphides. Figure 6a-c shows typical sulphide assemblages and microstructures.

In the NE part of the INCO property, an auriferous sericitized zone occurs in mylonites parallel to the SW extension of the EHSZ (*see* Fig. 2, 5). Assays up to 1.5 g over 12 m have been obtained (Slauenwhite, pers. comm, 1990). The alteration zone has been delimited in three dimensions by drilling, and its overall lens shape is discordant (20-30°) to the hosting shear zone.

The NE-striking mylonite zone is poorly exposed in this area, but available outcrops suggest a minimum width of 100 m. Subvertical mylonitic foliations and lineations are variably developed in a package of felsic sericite schists and chlorite schists (Fig. 4d), and are overprinted by later vertical folds. Kinematic indicators (C + S fabrics, asymmetric folds, shear bands) in mylonites show that early movements in the shear zone were NW-side-up. Subsequent folds are generally vertical and subparallel to mylonitic foliations, however, locally they fold the subvertical mylonitic stretching lineation.

The auriferous sericitized zone consists of felsic schists, locally containing dense swarms of 1-10 mm thick quartz veins, which are late- to post-kinematic with respect to mylonitization. Fine grained disseminated pyrite was found in chlorite schists and in the auriferous zone itself.

The timing of the sericite alteration, and thus gold mineralization, is clearly important for mineral exploration, but it is not well constrained at present. If the sericitization and mineralization are associated with quartz vein emplacement, a late- to post-mylonite, but prefolding age is probable. If, on the other hand, the sericite alteration predates the quartz veins, the mylonitization only represents a minimum age of gold mineralization. Considering the extent of the EHSZ, this problem clearly requires further consideration.

In the southern part of the Southern Highlands shear zone (Fig. 5), fine grained white sericite schists occur sporadically in the mylonite zone ('s' in Fig. 5). The mylonitized schists are concordant, and range in width from a few to tens of metres. Locally, the schist contains fine grained pyrite which is disseminated or occurs in concordant thin veins. So far, assayed sericite schists have been non-auriferous.

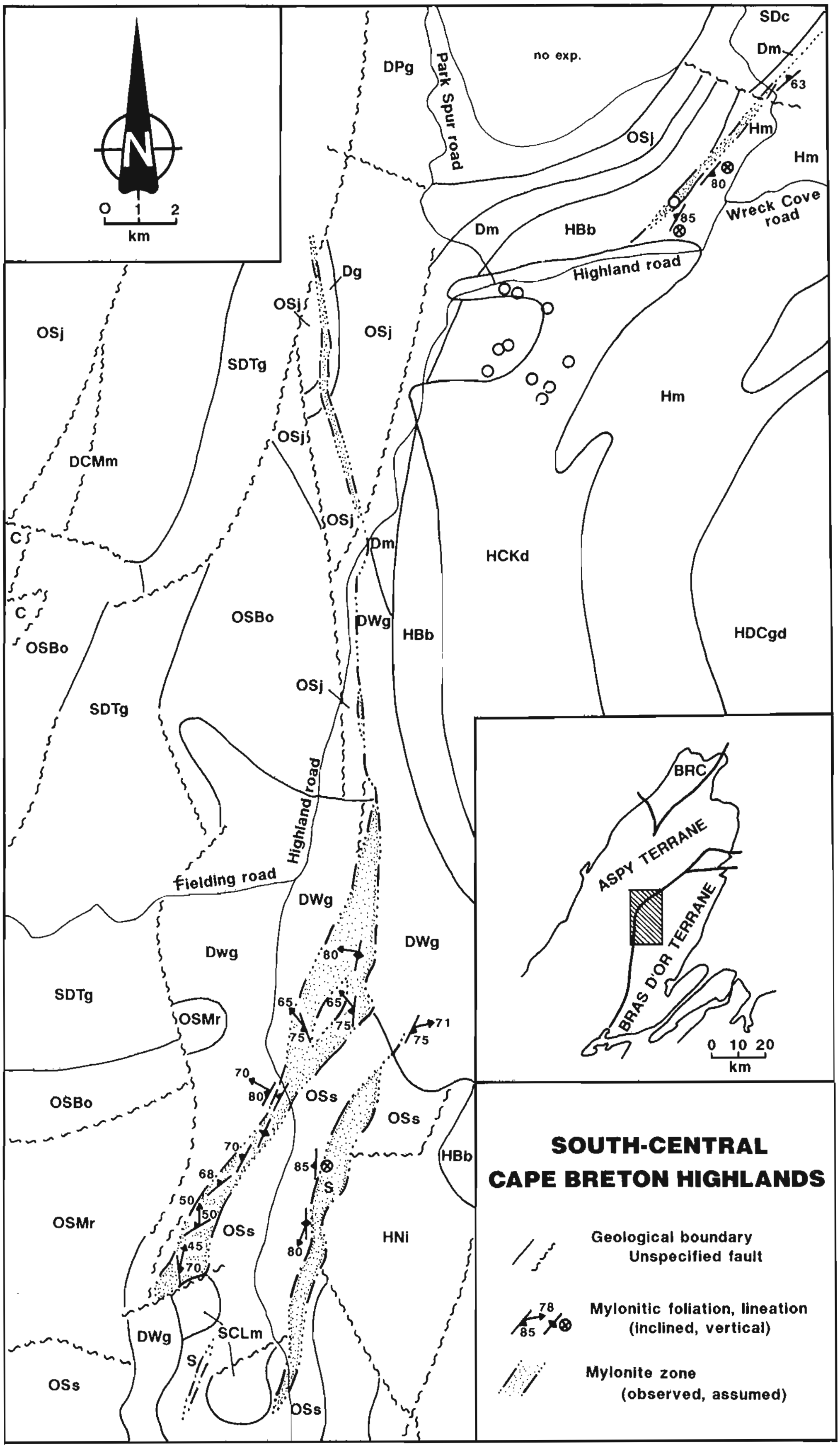
SOUTH-CENTRAL
CAPE BRETON HIGHLANDS
Geological boundary
Unspecified fault
Mylonitic foliation, lineation
(inclined, vertical)
Mylonite zone
(observed, assumed)
ASPY TERRANE
BRAS D'OR TERRANE
BRC
0 10 20
km
0 1 2
km
N
Park Spur road
Highland road
Wreck Cove road
Fielding road
no exp.
DPg
SDc
Dm
Hm
OSj
HBb
Dg
SDTg
DCMm
OSBo
HCKd
HDCgd
DWg
Dwg
OSMr
OSs
HNi
SCLm
C

Table 2. Vein type mineralizations in Aspy and Bras d'Or terranes

Occurrence	*Au-value*	*Sulfides*	*Host rock (thickness)*	*Structure*
Gray Glen Bk. (2)	1.4 ppm	py, cpy, sph (diss., str.)	chl-mica schist (5-40 cm)	Dextral shear in steep N-S zone, moderately NNW plunging lineation.
Jerome Bk. (3c)	250 ppb	py, cpy (Cu-stain common)	mafic dykes in granite	Discordant calcite +/- quartz veins in gabbroic dykes in Cheticamp granite
Second Gold Bk. (7)	VG locally	py, aspy, cpy, sph, ga, bi	greenschist facies metasediments	Qtz-veins 2 cm-2 m thick, syn-post tectonic. SSE-directed thrusting along NNW-dipping planes
Franey Bk. (1)	>10 ppm	py	greenschists	Veins <=10 cm, strongly folded, occurrence hosted by the EHSZ.
Price Point (8a)	trace Au* 6.02 oz/t Ag*	py, cpy, sph	dacitic to andesitic tuffs and flows	Mineralized veins a few cm wide, occur as shear veins and extension gash veins. NNW-trending steep brittle fault with E-side up movement.
Murray Point (8a)	0.19% Cu, 0.27% Mo over 17 m* 0.04% Cu, 0.003% Mo over 180 m (5 DDH)*	py, cpy, mo, (hem)	granodiorite, leucogranite	1-2 cm veins in stockwerk with 5-20 cm sericite alteration zones.
Elders Bk. (8a)	0.47 oz/t* 0.58 oz/t*	py, sph, cpy	andesitic mafic schist	Mineralized vein is 10-20 cm thick and ESE-dipping. Reverse WNW-directed movements along vein.
Goose Cove Bk. (8b)		py, cpy, mo	porphyritic granite	Steep E-dipping NNW-striking faultzone, 1 m alteration zone, sinistral movement with small W-side up component.
INCO (9)	VG locally	py, (hem)	mafic gneisses	Shear veins and a few extension gash veins, up to 1 m wide and 70 m long. Emplaced during N-NW directed thrusting along gently S-SSW dipping planes.

Abbreviations: see Table 1 (bi - bismuthinite).
* from Macdonald & Barr (1985)

Figure 5. Map of the central part of the southern Cape Breton Highlands, showing the location of the Southern Highlands shear zone and generalized geology (from Barr et al., in press). Inset shows location of map (BRC — Blair River Complex). Abbreviations (upper case one or two letter prefixes indicate age of unit (C — Carboniferous, D — Devonian, S — Silurian, O — Ordivician, H — Hadrynian, HC — Hadrynian-Cambrian)): DCMm — Margaree pluton (341 ± 17 Ma (Rb-Sr); O'Beirne-Ryan et al., 1986), Dm — Mylonite, chlorite schist in shear zones (equivalent to EHSZ), DWg — West Branch North River granite (399 ± 4.6 Ma (Rb-Sr); O'Beirne-Ryan and Jamieson, 1986), Dg — Undifferentiated unfoliated granitoids, DPg — Park Spur granite, SCLm — Leonard MacLeod Brook Complex, (granitoids occurring as intrusion breccia in mafic volcaniclastics), SDTg — Taylors Barren Pluton (419 ± 17 Ma (Rb-Sr); Gaudette et al., 1985), SDc — Chéticamp Lake gneiss (396 ± 2 Ma (U-Pb, zircon); Dunning et al., 1990), OSj — undivided Jumping Brook metamorphic suite, OSBo — Belle Côte Road gneiss (433 ± 20 Ma (U-Pb, zircon; Jamieson et al., 1986), OSMr — Middle River metamorphic suite, OSs — Sarach Bk. metamorphic suite (433^{+7}_{-4} Ma, U-Pb (zircon); Dunning et al., 1990), HCKd — Kathy Road Dioritic Suite (560 ± 2 Ma U-Pb (zircon); Dunning et al., 1990), HBb — Bateman Brook gneiss, HM — McMillan Flowage Formation, HNi — North Branch Baddeck River Leucotonalite (614^{+38}_{-4} Ma, U-Pb (zircon); Dunning et al., 1990). 's' — mylonitized sericite schist, 'o' — major quartz veins in the INCO property.

Vein type mineralizations

Mineralization associated with synkinematic emplacement of quartz ± calcite veins occur in both Aspy and Bras d'Or terranes. Table 2 summarizes the characteristics, and Figures 2 and 7 show locations.

The veins at St. Anns (8a, 8b in Table 2) occur in broadly N trending fault zones, and while there is a clear structural control, there are also lithological controls on the types of mineralizations. Veins occurring in granitoids generally carry chalcopyrite and molybdenite, whereas veins in dacitic/andesitic tuffs and flows contain sulphide assemblages characterized by sphalerite and galena. In an attempt to date the mineralizing event, Macdonald and Barr (1985) obtained a 517 ± 18 Ma K-Ar age of white mica from alteration zones. Recent Pb-isotope studies (Sangster et al., 1990) support this age, and show isotopic evidence for an additional vein forming event.

Auriferous quartz veins at the INCO property (9 in Fig. 2, circles in Fig. 5) are hosted by a heterogeneous suite of intrusions of late Proterozoic to early Cambrian age. One of these, the Kathy Road diorite, has been dated at 560 ± 2 Ma (U-Pb, zircon) and 521 ± 2 Ma (U-Pb, titanite; both from Dunning et al., 1990).

The veins are composed of coarse, locally prismatic, milky white quartz, and may contain up to 2% euhedral pyrite grains and a marginal hematized alteration zone. These hematized zones yield the best gold values (Slauenwhite, pers. comm., 1990) and carry sporadic visible gold.

Kinematic indicators (C+S fabrics, shear bands) observed in host rocks show that reverse movements took place along all veins with E-W to ENE-WSW orientations. Veins oriented between NW-SE and NE-SW display a major component of strike-slip movements. In only

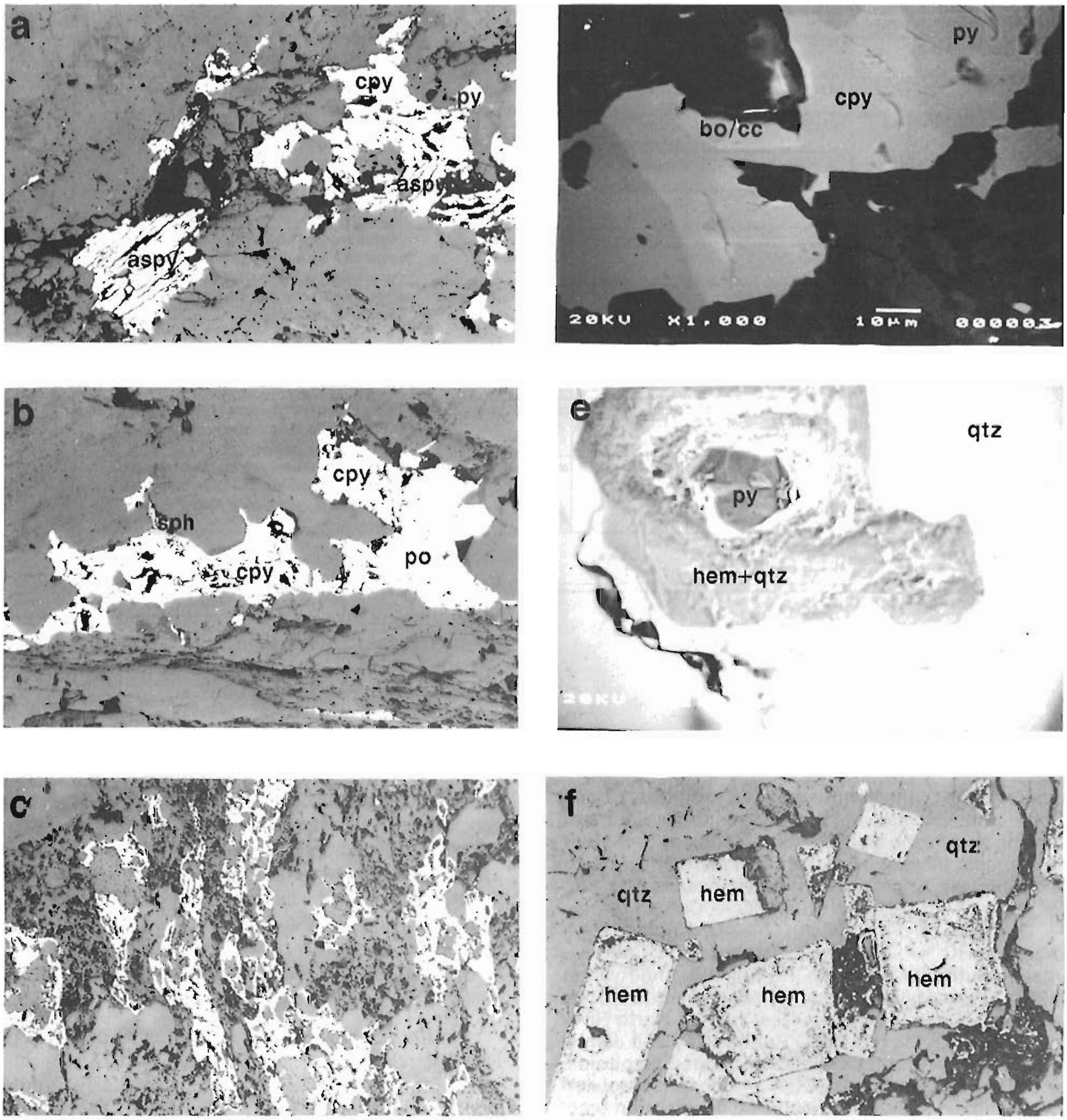

Figure 6. Sulphide assemblages and microstructures.
(a) Mountain Top Adits: Chalcopyrite (cpy) and pyrite (py) with fractured arsenopyrite (aspy) in silicified sericite schist. Long dimension of photo is 1.2 mm. Reflected light.
(b) Fisset Bk.: Thin continuous foliation parallel layer of pyrrhotite (po), chalcopyrite (cpy) and sphalerite (sph) occurring between two thick layers of pyrite (outside field of view) in a quartz-chlorite-sericite matrix. Long dimension of photo is 0.95 mm. Reflected light.
(c) Fisset Bk.: Layers of syngenetic pyrite (white) mobilized into S and C planes during deformation and metamorphism of host rock. Grey matrix is quartz, muscovite, chlorite. Rimward dissolution features in pyrites are common. Long dimension of photo is 1.2 mm. Reflected light.
(d) Second Gold Bk.: Pyrite (py), chalcopyrite (cpy) and bornite/chalcocite (bo/cc) in quartz-calcite-muscovite matrix (dark grey and black). Back scatter electron image.
(e) INCO quartz veins: Pyrite grain (square outline still visible) partially replaced by hematite and quartz (hem + qtz) in a matrix of pure quartz (qtz). Unaltered pyrite (py) in core of alteration. Back scatter electron image.
(f) INCO quartz veins: Pyrite grains (square outline) completely replaced by hematite (hem) in quartz (qtz) matrix. Long dimension of photo is 2.3 mm. Reflected light.

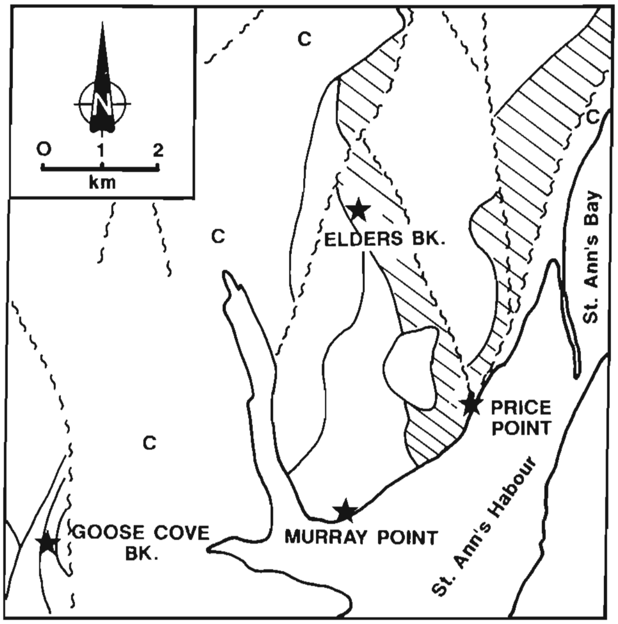

Figure 7. Sketch map of the St. Anns area, showing the location of occurrences (geology from Macdonald and Barr, 1985). Legend: C — Undivided Carboniferous (Horton and Windsor groups), ornamented — Dacitic/andesitic tuffs/flows of the Price Point formation, unormamented — undivided Late Proterozoic to Cambrian granitoids. Stars — location of mineral occurrences. See Figure 2 for location.

a few instances was evidence of down-dip extension observed. Internally, the quartz veins only display brittle deformation with no substantial movement.

It is suggested that the veins were emplaced at relatively shallow levels, synchronously with faulting and shearing in a broadly NNW-SSE oriented contractional regime, and that the observed variation in vein orientation is partly caused by lithological, and thus rheological, variations in the country rocks. The majority of the veins are *shear veins*, whereas the few horizontal veins may be *extension gash veins*.

Hematization, during which pyrite was replaced by hematite ± quartz, was clearly the alteration process with which Au-precipitation was associated (Fig. 6e-f). We propose that hematization was caused by mixture of circulating fluids with meteoric water, thus causing an increase in pH and promoting Au-precipitation. The lack of basemetals and usual gold-pathfinders (As, Hg) support this suggestion, although further studies (e.g. fluid inclusions) are needed to assess the model.

MYLONITES/SHEARZONES IN SOUTHERN HIGHLANDS

The Eastern Highlands shear zone (EHSZ) separates the Aspy and Bras d'Or terranes in the eastern Cape Breton Highlands (Barr et al., 1987a; Barr and Raeside, 1989; Lin, 1990; Barr et al., in press) (Fig. 2).

During the present study, a major mylonite zone was mapped in the southern Cape Breton Highlands (Fig. 5); for convenience, the name Southern Highlands shear zone (SHSZ) will be used below, however, this does not imply that this zone is a regional terrane boundary like the EHSZ.

The SHSZ (Fig. 5) is composed of felsic to intermediate mylonites and phyllonites, presumably developed in pyroclastics of the Sarach Brook metamorphic suite (433^{+7}_{-4} Ma, U-Pb (zircon) in rhyolite, Dunning et al., 1990) and adjacent granitoids (Fig. 5). Mafic mylonites are also observed locally. The width of the zone is clearly related to the nature of the country rock; in felsic pyroclastic rocks in the southern part the zone reaches 1.5 km, whereas it becomes narrower and more diffuse in intrusives and mafic volcanics (Fig. 5).

The mylonitic foliation in the SHSZ is vertical to steeply W-dipping and shows, especially in the southern part, quite variable orientation (Fig. 4c, 5). The orientation of mineral and aggregate stretching lineations varies from moderately N-plunging to vertical. Kinematic indicators (C + S fabrics, shear bands, asymmetric fish) in mylonites all show ESE-side-up movements. Throughout the area, mylonitic fabrics are overprinted by tight vertical folds (centimetre-scale) showing both dextral and sinistral asymmetries. Late brittle fracturing, characterized by open space extension gashes, centimetre-scale kinking and local cataclasis in brittle zones, is ubiquitous.

Coupled with the ESE-up sense of movement, some of the orientation data seen in Figures 4c and 5, would suggest that shearing and mylonitization in the SHSZ occurred in an extensional regime. It is, however, considered more likely that the original tectonic regime was contractional, and that the present orientation of the zone is due to subsequent deformation. The large spread seen in Figure 4c is compatible with this structural overprint.

In Figure 5, the EHSZ is represented by unit Dm. Studies of kinematic indicators in a mylonite zone immediately east of this unit show NW-up movements along steep lineations (Fig. 4d), similar to results obtained by Lin (1990) farther northeast. This is in sharp contrast to the SSE-side up movements in the SHSZ and raises the question whether these two mylonite zones are related, or whether they represent two different stages (possibly intra-terrane) of mylonitic overprint on a major terrane boundary.

In the northern part of the area, the SHSZ swings to a N to NNW trend, and does not appear to join the EHSZ (Fig. 5) along mylonitized zones. Foliated felsic and mafic gneisses (unit Dm in Fig. 5) are shown by Barr et al. (in press) as the SW-continuation of the EHSZ. We do not dispute the presence of sheared rocks in the Dm-unit, but we suggest that the mylonite zones in Figure 5 are not co-extensive. This is supported by the differing trends and kinematics of mylonites in the EHSZ and SHSZ.

Docking of the Aspy and Bras d'Or terranes began in the mid-Silurian and the terranes were fully juxtaposed by late Devonian (Dunning et al., 1990). The EHSZ and SHSZ are major structural features that likely represent terrane boundaries, however the regional kinematics of amalgamation are not well constrained at present.

SUMMARY AND DISCUSSION

Gold-bearing mineralization in the Cape Breton Highlands displays distinct regional differences and similarities. Figure 8 shows the generalized structural setting of each occurrence, and further features are reviewed and discussed below.

Gold mineralization in the Aspy terrane ranges from stratabound sulphide occurrences (Chéticamp area) to mesothermal quartz-vein occurrences (Second Gold Brook). The timing of mineralization also varies across the terrane. In the Chéticamp area, mineralization is probably syn-depositional (ca. 430 Ma, Jamieson et al., 1986; 1990) and (re)mobilized during later metamorphism and deformation of their hosts (400 Ma; Jamieson et al., 1990; Reynolds et al., 1989), contrasting sharply with the syn- to late-kinematic quartz-veins at Second Gold Brook.

In the Bras d'Or terrane, mesothermal quartz-veins at INCO's property were emplaced during brittle-ductile NNE-directed thrusting, whereas fault controlled occurrences at St. Anns clearly formed under conditions of brittle deformation. INCO quartz veins are hosted by foliated mafic gneisses of the 560 Ma Kathy Road diorite suite (Dunning et al., 1990), and have been suggested to be related to Siluro-Devonian movements on the EHSZ (Sangster et al., 1990), however, no radiometric age dates exist. The St. Anns veins are hosted by Late Proterozoic to Cambrian granitoids and volcanics, and K-Ar ages of discordant alteration zones have yielded 517 Ma (Macdonald and Barr, 1985). Recent Pb-isotope studies (Sangster et al., 1990) support this age, and suggest the presence of an additional vein forming event. It is notable, that the 517 Ma age, should it be confirmed by more precise methods, predates Aspy-Bras d'Or terrane amalgamation by ca. 100 Ma.

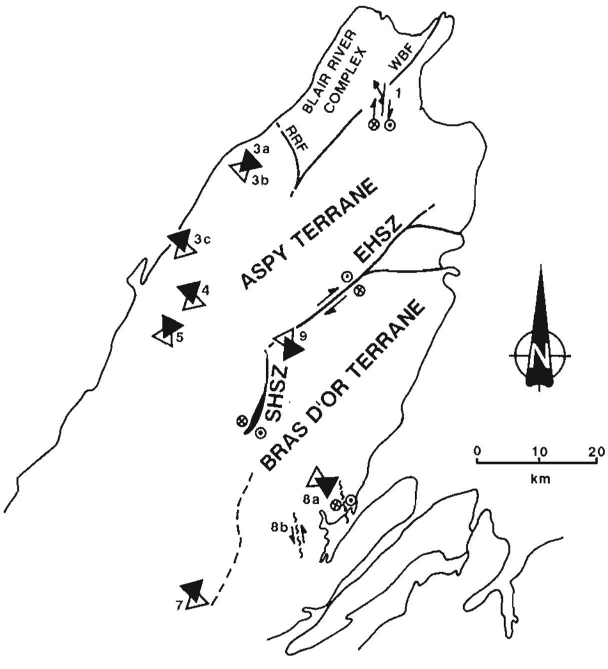

Figure 8. Summary of structural information from lithologies hosting mineralizations. Abbreviations: WBF — Wilkie Brook fault zone, RRF — Red River fault zone, (E/S)HSZ — (Eastern/Southern) Highlands shear zone, numbers correspond to occurrences mentioned in text. Arrowheads (black over white) represents direction of reverse thrust/shear zones. Subvertical components are indicated by circles with dots (up) and x (down).

A NW to SE transect across the Cape Breton Highlands traces the gradual change from ductile, amphibolite facies conditions affecting and mobilizing mineralization in the NW, over greenschist facies brittle-ductile regimes at INCO's property, to brittle low greenschist facies fault controlled occurrences in the St. Anns area.

There is also a systematic variation in the geometry of the contractional regime responsible for the host structures. In the western Aspy terrane, the dominant contraction occurred along NNE-SSW to N-S oriented stress fields, whereas NW-SE oriented contraction appears more likely farther east. This apparent counterclockwise rotation of the stress field is not considered significant as the mineralization may have formed over a 100 Ma period.

The results of this work clearly emphasize the need for a good structural and chronological framework when dealing with mineral deposits. While important progress has been made in establishing a regional chronology in Cape Breton Island (e.g. Dunning et al., 1990) and in the understanding of the EHSZ (Lin, 1990), the problems of kinematics and extent of tectonic features on a terrane scale warrants further investigations.

An understanding of the structural setting and history of mineral deposits is necessary in order to understand and predict the extent of mineralized zones; future work focusing on these problems will benefit both mineral exploration and academic interests.

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L'amas sulfuré de Champagne : un gîte exhalatif dans les argilites ordoviciennes du Groupe de Magog, Appalaches du Québec.

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Bossé, J., Paradis, S. et Gauthier, M., L'amas sulfuré de Champagne : un gîte exhalatif dans les argilites ordoviciennes du Groupe de Magog, Appalaches du Québec; dans Recherches en cours, Partie D, Commission géologique du Canada, Étude 91-1D, p. 129-134, 1991.

Résumé

L'amas sulfuré polymétallique de Champagne est encaissé dans les argilites noires du Groupe de Magog, d'âge ordovicien moyen, reposant dans le synclinorium de Saint-Victor. Il s'agit, en fait, de plusieurs lentilles stratiformes constituées d'agrégats massifs, laminés ou bréchiques de pyrite, de pyrrhotine, de sphalérite, de chalcopyrite et de galène. Une zonation géochimique et minérale verticale est observée à l'échelle de l'amas sulfuré. Ainsi, de la base au sommet, on note le passage d'un assemblage de pyrrhotine-chalcopyrite à un assemblage de sphalérite-pyrite. On y constate également un enrichissement graduel en or, en argent et en arsenic.

L'amas sulfuré de Champagne présente de nombreuses similarités avec les gîtes sous-marins exhalatifs encaissés dans les sédiments (SEDEX).

Abstract

The Champagne polymetallic massive sulfide deposit is hosted by Middle Ordovician black argillites of the Magog Group of the Saint-Victor Synclinorium. The deposit consists of several conformable lenses which are formed of massive, banded and brecciated pyrite, pyrrhotite, sphalerite, chalcopyrite, and galena. A chemical and mineralogical zonation is observed from the base to the top of the sulfide lenses: pyrrhotite-chalcopyrite gives way to sphalerite-pyrite with a progressive enrichment in Au, Ag and As.

The Champagne deposit has most of the characteristics of the sediment-hosted submarine exhalative deposits (SEDEX).

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INTRODUCTION

Le potentiel aurifère des Appalaches du Québec est connu depuis 1823, date de la première découverte d'or alluvionnaire en Beauce (Douglas, 1864; Gauthier et al., 1989). Des recherches entreprises afin de découvrir la source de ces placers aurifères, amenèrent, par la suite, la découverte de nombreux indices minéralisés dans les roches du socle. Plusieurs de ces indices sont contenus dans les roches du Groupe de Magog, lesquelles sont les roches hôtes de l'amas sulfuré de Champagne. Le gîte de Champagne a été trouvé en 1952, à la suite de la découverte d'un bloc erratique minéralisé. L'année suivante, Panet Metals Corporation Limited effectue des levés géophysiques et plus de 3 000 mètres de forage dans la région du gîte de Champagne. À la suite de ces travaux, les réserves de ce gîte sont évaluées à 290 000 tonnes à des teneurs de 2,4 g/t Au, 19,7 g/t Ag, 0,4 % Cu, 2,7 % Zn et 0,5 % Pb (Bergman, 1954). Entre 1957 et 1985, peu de travaux ont été entrepris. En 1985, Golden Hopes Mines Limited acquiert les droits miniers et entreprend des travaux d'exploration sous la direction de R.E. Schaaf et Associés.

Le gîte de Champagne est situé à 105 km au sud-est de la ville de Québec (fig. 1). Il s'agit d'un corps stratiforme de sulfures massifs polymétalliques encaissé dans les roches sédimentaires du Groupe de Magog. L'amas sulfuré de Champagne se compare aux gîtes sous-marins exhalatifs encaissés dans les sédiments (SEDEX), tels que Rammelsberg et Megen en Allemagne ou ceux du bassin de Selwyn au Yukon (Gustafson et Williams, 1981; Large, 1983; Abbott et Turner, 1990).

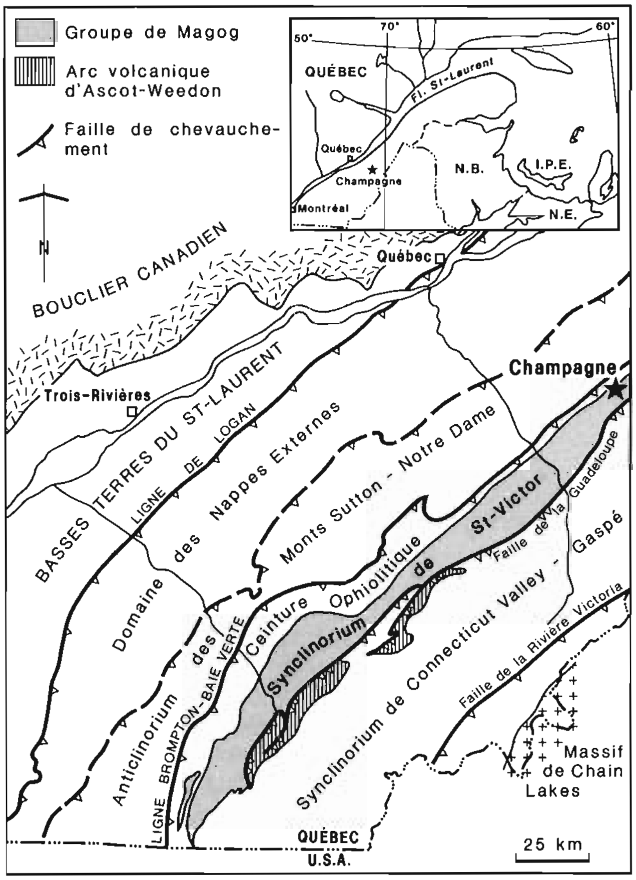

Figure 1. Carte de localisation du gîte de Champagne et du synclinorium de Saint-Victor dans les Appalaches du Québec.

CONTEXTE GÉOLOGIQUE RÉGIONAL

Le gîte de Champagne est encaissé dans les argilites noires de la base de la Formation de Beauceville appartenant au Groupe de Magog. Le Groupe de Magog forme les assises du synclinorium de Saint-Victor qui s'étend sur plus de 255 km, depuis la frontière du Vermont, au sud-ouest, jusqu'à celle du Maine au nord-est (Slivitzky et St-Julien, 1987; fig. 1). Ces roches sont interprétées comme des dépôts de bassin avant-arc euxinique d'âge ordovicien moyen (Llandeilien-Caradocien; Laurent, 1987; Cousineau, 1988). Ce bassin avant-arc s'est développé entre les complexes ophiolitiques obductés à l'ouest et l'arc volcanique d'Ascot-Weedon au sud-est (Desbiens, 1988; Godue, 1988). De l'Ordovicien tardif (orogenèse taconique) jusqu'au Dévonien moyen (orogenèse acadienne), les unités du Groupe de Magog ont été progressivement déformées pour constituer le synclinorium de Saint-Victor. La principale phase de plissement affectant les roches du Groupe de Magog est associée à l'orogenèse acadienne (Bernard, 1987; Cousineau, 1988; Tremblay et St-Julien, 1990).

Cousineau (1988) subdivise le Groupe de Magog en quatre unités lithostratigraphiques qui sont, de la base au sommet (fig. 2) : 1) la Formation de Frontière qui se compose d'une alternance de grès volcanogènes verts et d'argilites noires à verdâtres; 2) la Formation d'Etchemin constituée de mudslates et de roches volcanoclastiques verdâtres; 3) la Formation de Beauceville composée d'argilites noires interstratifiées avec des roches volcanoclastiques; et 4) la Formation de Saint-Victor, une séquence à turbidites classiques au sein de laquelle sont interstratifiées des roches volcanoclastiques. Des filons-couches et dykes de gabbro-diorite, dans lesquels des indices aurifères filoniens sont présents, recoupent les sédiments du Groupe de Magog.

La nature des contacts géologiques entre les différentes formations du Groupe de Magog est normale (J.F. Burzynski, communication personnelle, 1990). Selon Cousineau (1988), ils auraient d'abord été concordants pour être, par la suite, découpés par des failles de chevauchement vers le nord-ouest.

CONTEXTE GÉOLOGIQUE DU GÎTE DE CHAMPAGNE

Roches hôtes

Les argilites noires situées à la base de la Formation de Beauceville sont les roches hôtes du gîte stratiforme de Champagne ainsi que de disséminations ($\leq$ 5 %) syndiagénétiques de pyrites aurifères, argentifères, cuprifères, zincifères, plombifères et arsénicales (Gauthier et al., 1989). Ces argilites noires sont riches en matières organiques et sont localement fossilifères (graptolites d'âge

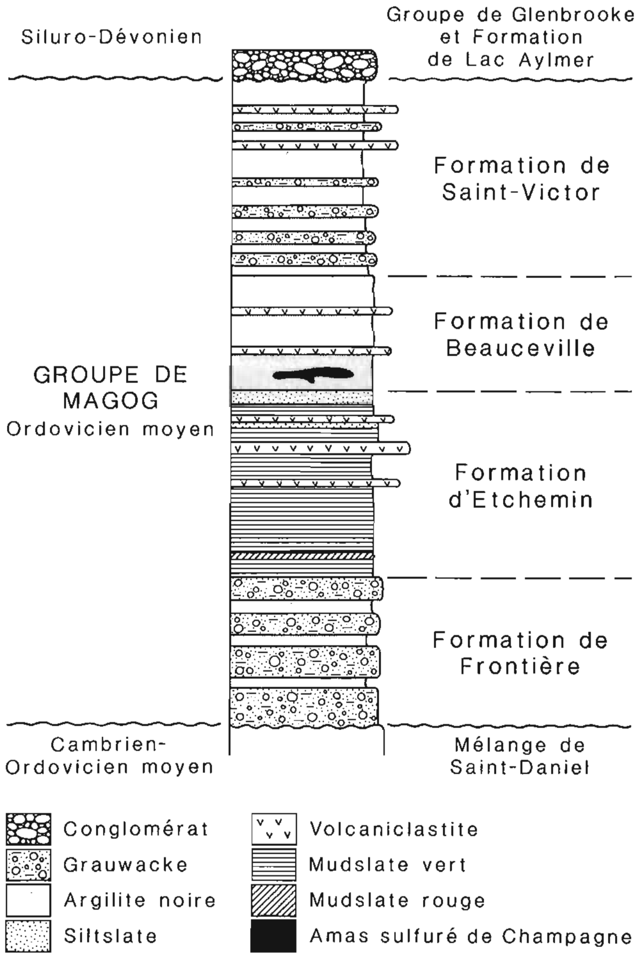

Figure 2. Colonne stratigraphique schématisée du Groupe de Magog. Modifiée de Cousineau (1988).

ordovicien moyen; Berry 1962). Elles se présentent en lits de 2 mm à 5 cm et forment des séquences continues sur plusieurs mètres d'épaisseur. Ces roches sont constituées de quartz, de micas, de chlorite, de graphite et de grains fins de pyrite automorphe et framboïdale formant des interlits d'épaisseur millimétrique. Les argilites noires sont généralement dépourvues d'altération, à l'exception d'une silicification localisée en bordure des filons-couches de gabbro-diorite. Les argilites noires de la Formation de Beauceville représentent un milieu de sédimentation calme, sapropélique, où l'apport en éléments terrigènes est faible comparativement aux autres formations du Groupe de Magog (Cousineau, 1988). D'importantes séquences d'argilites noires ont été observées dans des bassins océaniques profonds (300-3 000 m) où l'apport terrigène est faible et où des conditions anoxiques prévalent (Leggett, 1978, 1980; Watkins et Flory, 1986).

Des brèches polygéniques minéralisées en chalcopyrite et en pyrrhotine ainsi que des grès noirs à fragments volcaniques sont situés à la base de la Formation de Beauceville. Le faciès grossier des brèches est non organisé et consiste en fragments de mudslate et de clayslate siliceux, laminés et massifs (1 mm à 1 m de diamètre) inclus dans une matrice silicifiée. Celle-ci est constituée de fragments de roches volcaniques et de mudslates ainsi que de cristaux de feldspath et de quartz, qui baignent dans une pâte riche en chlorite, quartz, feldspath et micas. Le faciès fin, finement laminé, présente un granoclassement normal et est riche en fragments de roches volcaniques et de mudslates ainsi que de cristaux de feldspath. La nature et l'origine de ces brèches sont incertaines. Il s'agit possiblement de brèches de talus ou de coulées de débris peu remaniées.

L'amas sulfuré de Champagne occupe le flanc sud-est d'un anticlinal de deuxième ordre (fig. 3), situé dans le flanc nord-ouest du synclinorium de Saint-Victor plongeant faiblement vers le sud-ouest (R.E. Schaaf et Associés, communication personnelle, 1990).

Minéralisation

Le gîte de Champagne regroupe plusieurs lentilles tabulaires de sulfures massifs polymétalliques s'étendant sur plus de 1,2 km selon une direction N040° (fig. 4). La puissance des amas varie entre 2 et 6 m (Godue, 1988; Gauthier et al., 1989).

La minéralisation consiste en pyrite, pyrrhotine, sphalérite, chalcopyrite et galène. L'arsénopyrite est localement observée. Ces sulfures se présentent en agrégats massifs, bréchiques ou laminés. Le faciès bréchique est principalement situé dans la partie inférieure de l'amas, là où la minéralisation passe à un faciès chaotique avant

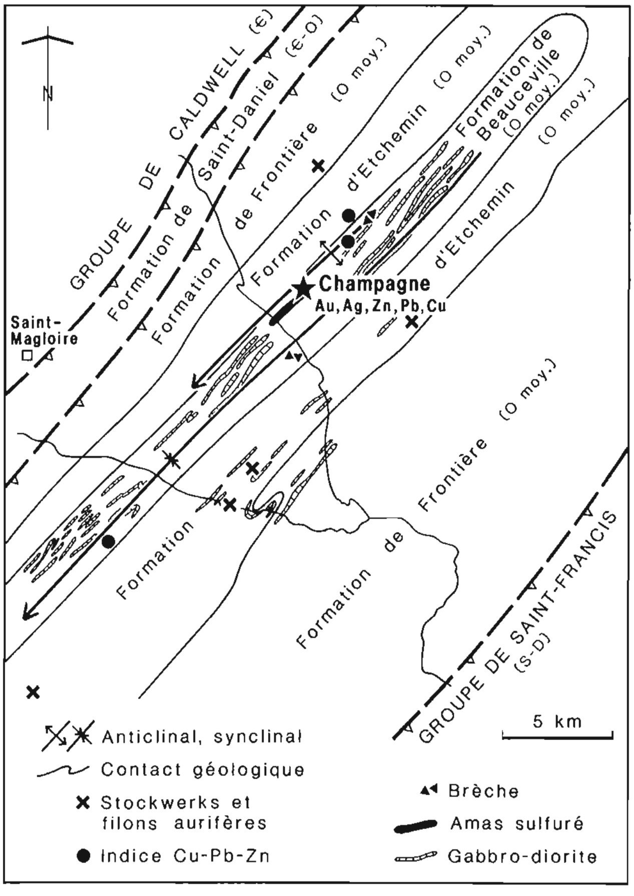

Figure 3. Carte géologique simplifiée de la région du gîte de Champagne dans les Appalaches du Québec.

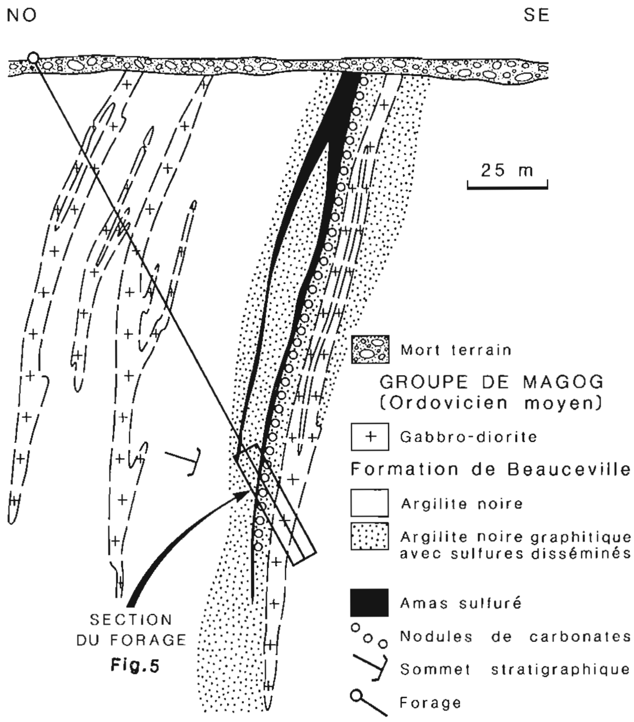

Figure 4. Coupe géologique NO-SE de l'amas sulfuré de Champagne parallèle à la direction d'un trou de forage. Modifiée de Gauthier et al. (1987).

d'atteindre la zone de griffon. Le faciès massif forme le coeur de l'amas tandis que le faciès laminé est prédominant au sommet de l'amas. Les sulfures sont équigranulaires et à grain fin. Les textures primaires sont fréquemment oblitérées par la foliation régionale. Ainsi, les sulfures sont communément remobilisés le long des plans de la schistosité régionale.

Une zonation géochimique et minérale verticale est observée à l'échelle de l'amas sulfuré. Jébrak et Gauthier (sous presse; fig. 5) définissent quatre zones de la base au sommet de l'amas : 1) un « stockwerk » constitué de veinules de quartz et de carbonates, et minéralisé en pyrrhotine, en pyrite et en chalcopyrite (2-3 %); 2) un agrégat massif de chalcopyrite et de pyrrhotine, pauvre en As, Zn et Cd; 3) un agrégat laminé de pyrite et de sphalérite, riche en Au, Ag et Co; et 4) une zone à nodules de carbonates (pseudomorphes de sulfates ?), riche en Ba et en éléments métalliques. Par ailleurs, on constate un enrichissement graduel en Au, As et Ag de la base vers le sommet de l'amas sulfuré.

MODÈLE GÉNÉTIQUE

L'amas sulfuré de Champagne possède plusieurs caractéristiques propres aux gîtes sous-marins exhalatifs encaissés dans les sédiments (SEDEX). Les caractéristiques communes sont : 1) la morphologie stratiforme de l'amas; 2) la zonation minérale verticale c'est-à-dire, d'un agrégat de pyrrhotine-chalcopyrite à un assemblage de sphalérite-pyrite; 3) le niveau à nodules de carbonates (pseudomorphes de sulfates ?) ainsi qu'un enrichissement en Ba au sommet de l'amas; et 4) la présence d'un probable stockwerk hydrothermal et d'une zone minéralisée bréchique à la base de l'amas sulfuré.

Cependant, le contexte géotectonique des gîtes de type SEDEX décrits par Large (1980, 1983) diffère de celui de l'amas sulfuré de Champagne. En effet, le gîte de Champagne est situé au sein de dépôts de bassin avant-arc développé entre des unités de mélange et d'ophiolites démembrées à l'ouest et un arc volcanique à l'est, tandis que les gîtes de type SEDEX décrits par Large (1980, 1983) se sont principalement formés dans des bassins intracratoniques, généralement des aulacogènes. Cette différence d'environnement géotectonique pourra se refléter dans le type et la proportion de métaux présents dans ces gîtes. En effet, la séquence de roches sédimentaires et volcaniques du Groupe de Magog résulterait en partie de l'érosion d'un arc volcanique caractérisé par un volcanisme felsique (Cousineau, 1988) et ultramafique (Labbé, sous presse) et en partie de l'érosion de complexes ophiolitiques. De leur côté, Godue (1988) et Desbiens (1988) suggèrent une source mafique-ultramafique pour les argilites noires et les mudslates du Groupe de Magog. Cet apport de roches mafiques et ultramafiques pourrait provenir de l'érosion des complexes ophiolitiques situés à l'ouest ou de l'arc volcanique situé à l'est du bassin avant-arc. Ces roches étant relativement riches en éléments tels que Au, Ag, As et Co, les gîtes de type SEDEX déposés dans un environnement semblable, comme celui de Champagne, seront naturellement enrichis en de tels éléments (fig. 5) par rapport à ceux de milieux cratoniques plus riches en plomb.

Les gîtes de type SEDEX sont généralement associés à des structures syngénétiques ayant canalisé des fluides hydrothermaux. Bien que plusieurs structures potentielles

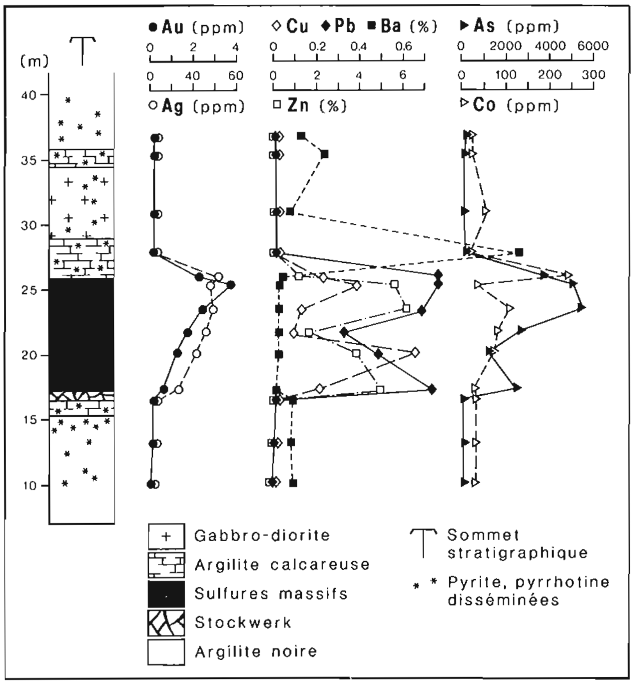

Figure 5. Coupe géologique et profils géochimiques de la section du trou de forage apparaissant à la figure 4. Modifiée de Jébrak et Gauthier (sous presse).

existent, telles que les brèches mentionnées précédemment, les minéralisations syngénétiques du Groupe de Magog n'ont pas de contrôle structural reconnu avec certitude.

SOMMAIRE ET CONCLUSION

L'amas sulfuré de Champagne ainsi que les sulfures disséminés dans les argilites noires de la Formation de Beauceville sont considérés comme des minéralisations exhalatives en milieu essentiellement sédimentaire (SEDEX). Le gîte de Champagne possède plusieurs caractéristiques propres aux gîtes de type SEDEX classiques (par ex. Rammelsberg et Megen en Allemagne, Tom et Jason au Yukon), telles qu'une zonation minérale de pyrrhotite-chalcopyrite à pyrite-sphalérite, des paragenèses minérales qui passent de concordantes et stratiformes au sommet de l'amas à chaotiques puis discordantes à la base, et un horizon à nodules de carbonates (pseudomorphes de sulfates ?) au sommet de l'amas. Cependant, le contexte géotectonique des gîtes de type SEDEX tels que ceux décrits par Large (1980, 1983) diffère de celui de Champagne; ce dernier s'étant formé dans un bassin avant-arc euxinique reposant sur un substratum ophiolitique. Ainsi, l'amas sulfuré de Champagne et les disséminations de sulfures stratiformes sont riches en éléments métalliques, tels que Au, Ag, As et Co.

En résumé, l'amas sulfuré de Champagne possède plusieurs des caractéristiques énumérées par Large (1980, 1983) pour les gîtes de type SEDEX. Ce parallèle fait ressortir le potentiel, jusqu'ici insoupçonné, du Groupe de Magog. L'importance de ce type de gisement et le contenu particulièrement aurifère des gîtes situés dans un environnement géotectonique aussi inusité, devrait commander une importante campagne de prospection. Il devient donc évident qu'une importante classe de gîtes de sulfures massifs est représenté dans les Appalaches du Québec et que le potentiel des sédiments du Groupe de Magog pour les gîtes de type SEDEX, comme celui de Champagne, est excellent.

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Observations on the structural control and tectonic setting of gold mineralization in the Cape Ray fault zone, southwestern Newfoundland.[1]

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Abstract

Field investigation of the Cape Ray fault in the area of the Cape Ray gold deposit and in a coastal section near Cape Ray village suggests that the post-Late Devonian movement of the fault is compatible with a transpressive dextral regime. Two increments of ductile strain have been recorded, the first characterized by reverse-oblique shearing and the second by strike-slip movement. These were followed by late brittle faulting associated with barren silicification.

Geometric analysis of the structures hosting the mesothermal gold veins at the Windowglass Hill and Big Pond showings suggests that mineralization is syn- to late-ductile shearing and genetically and spatially related to Acadian movement within the Cape Ray fault zone. The mineralization is compatible with the main ductile, post-Late Devonian deformation event. Both showings are located in brittle units and in subsidiary structures oblique to the main shear direction.

Résumé

Des travaux de terrain effectués le long de la zone de faille de Cape Ray dans la région du gisement aurifère de Cape Ray et le long d'une section bordant la côte semblent indiquer que le mouvement de la faille, postérieur au Dévonien tardif, est compatible avec un régime de transpression dextre. Deux stades de déformation ductile sont présents, un premier, caractérisé par un cisaillement inverse-oblique et un deuxième, caracterisé par du coulissage. Cette déformation a été suivie par des failles cassantes tardives et une silicification stérile associée.

L'analyse géométrique des structures hôtes des veines aurifères mésothermales des indices de Windowglass Hill et de Big Pond laisse supposer que l'événement minéralisateur est synchrone à tardi-cisaillement ductile. Cet événement est génétiquement et spatialement relié au mouvement acadien le long de la zone de faille de Cape Ray. La minéralisation est compatible avec le principal épisode de déformation ductile postérieur au Dévonien tardif. Les deux indices se situent dans des unités cassantes et dans des structures secondaires obliques à la direction principale de cisaillement.

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INTRODUCTION

The widespread spatial association between gold deposits and major fault zones is well documented, but the cause-effect relationship is not well understood. In order to better understand the gold mineralizing processes and their relationship to fault zones, a long-term project on structure and gold deposits in the Canadian Appalachians has been initiated (Dubé, 1990). Detailed structural mapping has been conducted on gold deposits in western Newfoundland with particular emphasis on the area around the Cape Ray gold deposit, where several mesothermal vein-type deposits are spatially associated with the Cape Ray fault zone (CRFZ). This paper summarizes the results of fieldwork done in this area during the summer of 1990.

Brown (1976, 1977) interpreted the Cape Ray fault as a Taconian suture, representing the trace of the Iapetus ocean, whereas Wilton (1983, 1984) considered it to be a major Acadian sinistral ductile shear zone. Chorlton (1983) recognized two distinct faulting events within the Cape Ray fault system: pre- to syn-Late Devonian sinistral wrench faulting and post-Late Devonian reverse-oblique dextral shear. Piasecki (1989) reaffirmed that the Cape Ray fault has the character of a major suture. He also recognized two movements along the fault zone: a pre-Late Devonian north-northwest overthrusting event followed by a post-Late Devonian overthrusting to the west-northwest.

A fundamental goal of this long term study in the Cape Ray area is to define the timing of the gold mineralizing event and its relationship to the structural history and geometry of the fault zone. For the Cape Ray deposit, Wilton and Strong (1986) have proposed a granite-related model in which gold mineralization slightly predates deformation and is related to an exsolved hydrothermal fluid originating from the synvolcanic Windowglass Hill granite. Tuach (1986) and Tuach et al. (1988) suggested that the mineralized lodes occur in a late brittle splay of the Cape Ray fault and that the Windowglass Hill granite and accompanying mineralization postdated the main movements on the fault zone.

This paper presents some key observations about the structural framework of the Cape Ray deposit area and associated gold occurrences using the well-exposed Windowglass Hill and Big Pond showings (Fig. 1). Based on this study, we propose that this gold mineralization is syn- to late-shearing and therefore genetically and spatially related to the fault zone.

GEOLOGICAL SETTING

The Cape Ray deposit is located within the Cape Ray fault zone in the southwestern part of Newfoundland (Fig. 1). The Cape Ray fault is a 100 km long and 1 km wide shear zone (Brown, 1977; Chorlton, 1983; Wilton, 1983). It separates two distinct terranes: the Cape Ray granite and isolated remnants of the Long Range ophiolitic complex to the northwest, and the Port-Aux-Basques gneissic complex and associated granite to the southeast (Wilton, 1983) (Fig. 1). The Devonian Windsor Point Group (WPG), located between these two terranes, consists of strongly deformed volcanic and sedimentary rocks intruded by the synvolcanic Windowglass Hill granite (Wilton, 1983). The northwestern and southeastern terranes have been respectively intruded by late tectonic post-Devonian granites: the Strawberry and Isle aux Morts Brook granites.

GEOLOGY OF THE CAPE RAY FAULT ZONE IN THE AREA OF THE CAPE RAY DEPOSIT

In the deposit area, the northwestern limit of the fault zone is marked by a well-developed mylonite zone within both the Windsor Point Group and the Cape Ray granite (Fig. 2). To the southeast, the fault zone is defined by a reverse-oblique ductile shear zone located at or near the contact between the Port aux Basques (PAB) gneiss and the Windsor Point Group. In between, the rocks are strongly deformed and most of the strain has been accommodated by folding. In the following sections, these shear-dominated and fold-dominated domains will be referred to as high strain zone (mylonites) and medium strain zone (fold-dominated) (Fig. 3).

Two styles of deformation can be observed in the area: a prominent ductile deformation and a local later brittle event. They will be discussed separately.

As was recognized by Wilton (1983, 1984), several deformation events can be identified in the Cape Ray deposit area and, on the basis of crosscutting relationships, foliations, stretching lineations, and folds are designated S_1, L_1, F_1, S_2, etc. However, the various fabrics described here are probably the result of a single incremental major deformation event. The orientation and distribution of the different structural elements and associated kinematics indicators are shown in Figure 2.

Ductile deformation

Medium strain domain (fold-dominated domain)

More than one planar fabric can be observed in the medium strain domain of the Cape Ray fault zone. The dominant fabric is defined either as a schistosity, cleavage, or pebble flattening. S_1 and S_2, trending northeast-southwest and dipping southeast generally form a composite S_{1-2} foliation, but S_1 is clearly discernible in the hinges of F_2 folds. The S_1 foliation is a penetrative bedding-parallel schistosity associated with a stretching lineation plunging steeply to moderately east or southeast (Fig. 2 and 4). No F_1 folds have been observed although Wilton (1983) reported that the S_1 foliation is axial planar to rare isoclinal folds. The S_2 foliation is axial planar to asymmetric F_2 folds which have redistributed the S_0 and S_1 foliation along a great circle having a pole coincident with measured F_2 fold axes (see Fig. 2, stereonet H). F_2 hinges are subcoaxial to the stretching lineation and plunge east-southeast. This parallelism is attributed to transposition of the hinges along the tectonic transport direction. The F_2 folds are millimetre to centimetre scale tight to isoclinal or chevron-type folds. Locally, some folds are more open. The structural vergence of F_2 folds usually suggests a reverse dextral movement. F_2 sheath folds are present in some areas, a reflection of the large strain sustained by these rocks.

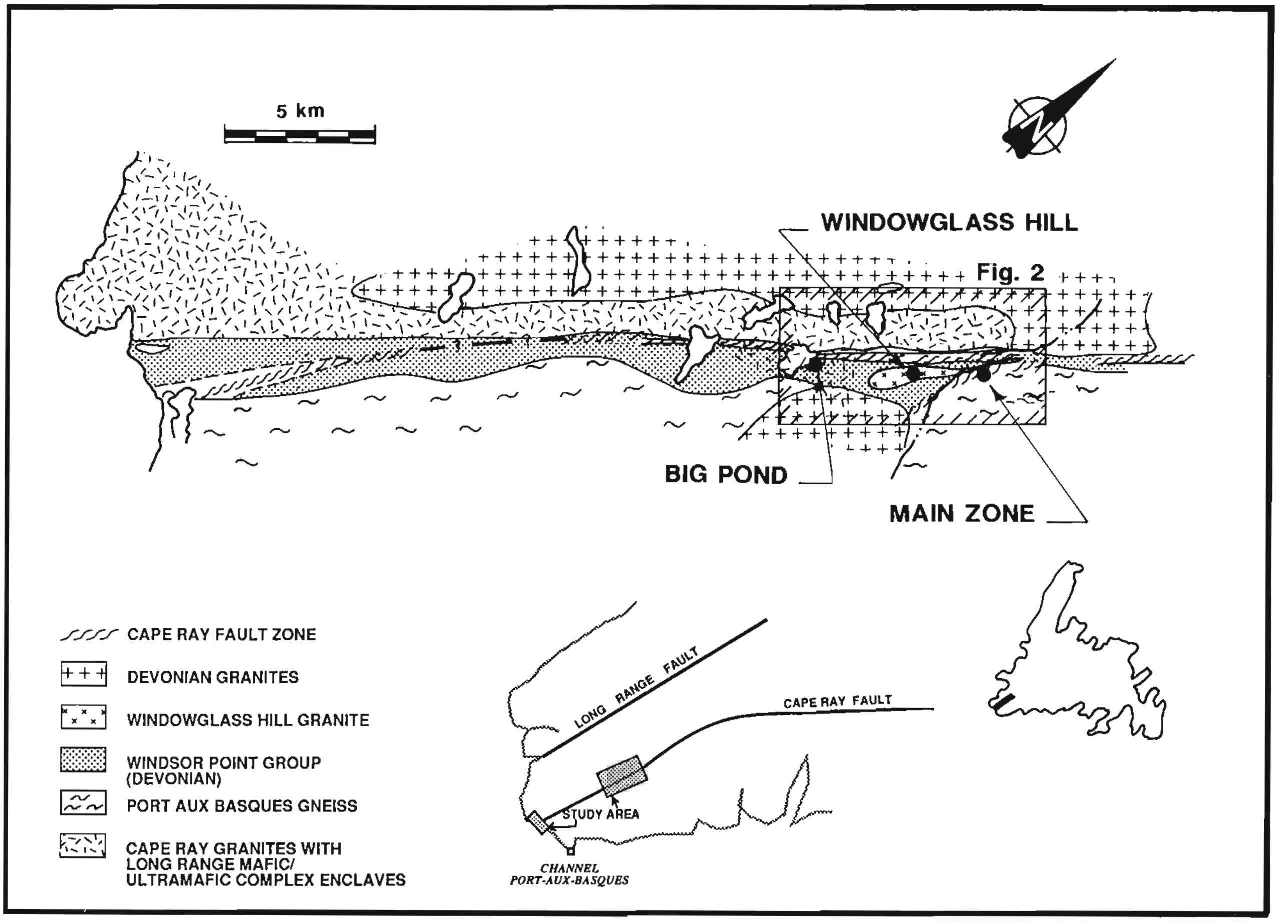


Figure 1. Simplified geological map of the Cape Ray fault zone showing the location of the studied gold showings (modified from Tuach et al. (1988); after Wilton (1983) and Tuach (1986)).

Kinematic indicators such as the vergence of F_2 folds and porphyroclast rotations suggest a reverse oblique movement with a dextral component along a northwest-southeast tectonic vector as indicated by stretching lineations. D_2 structures are locally folded by open, centimetre to metre scale F_3 folds. Poles of S_2 foliations are distributed along a great circle having a southwest pole corresponding to measured F_3 fold hinges (Fig. 2, stereonet C). The axial plane of F_3 folds is approximately coplanar with the overall northeast trend of the fault. More work is required to establish the exact relationship between F_3 fold and earlier structures.

High strain domain

Two mylonite zones have been recognized in the study area (Fig. 3). To the northwest, there is a well-developed mylonite zone which, for the most part, is located at the contact between the Windsor Point Group and Cape Ray granite. To the southeast, another ductile shear zone, known as the "50 shear", is located at or near the contact between the Port aux Basques gneiss and the Windsor Point Group. In the central part of the study area (Fig. 2), the latter is subparallel to and occurs approximately 300 m south of the structure known as the "main zone" shear which hosts most of the ore. To the east, in the deposit area, these two structures merge or are very close to one another.

The "50 shear" zone

The overall orientation of the 50 shear is (042°/75°), slightly oblique to the general trend of the Cape Ray fault (Fig. 2). The shear zone is more than 3 km long and approximately 100 m wide. It is well exposed in the Isle Aux Morts River (IAMR), in several brooks, and in the underground ramp of the mineralized "41 zone" (Dubé, 1990) where it occurs in the hangingwall of the mineralized lodes. It is made up of mylonitic chlorite and chlorite/sericite schists produced either by retrogression of the Port aux Basques gneiss or shearing of the Windsor Point Group. Mylonitic quartzofeldspathic rocks are also present within this high strain zone. All the mylonitic schists contain discontinuous foliation-parallel quartz shear-veins. Due to the intensity of deformation, it is commonly difficult to recognize the contact between the Port aux Basques gneiss and the Windsor Point Group units.

The internal structure of the "50 shear" is characterized by a S-C-type mylonitic foliation (Lister and Snokes, 1984) and/or a compositional layering subparallel to the shear zone. Noncoaxial deformation in the schist is suggested by the presence of a second schistosity with a

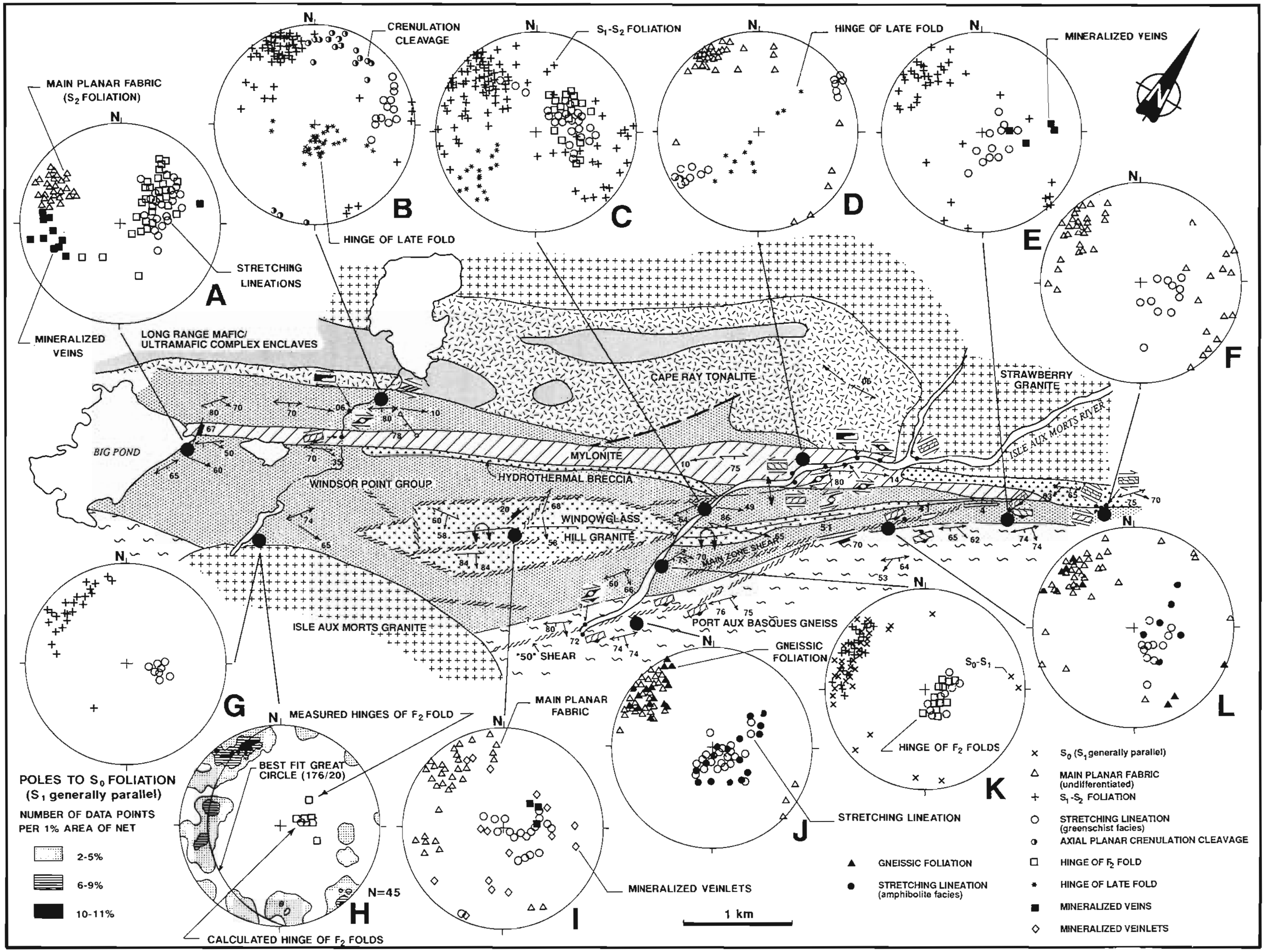
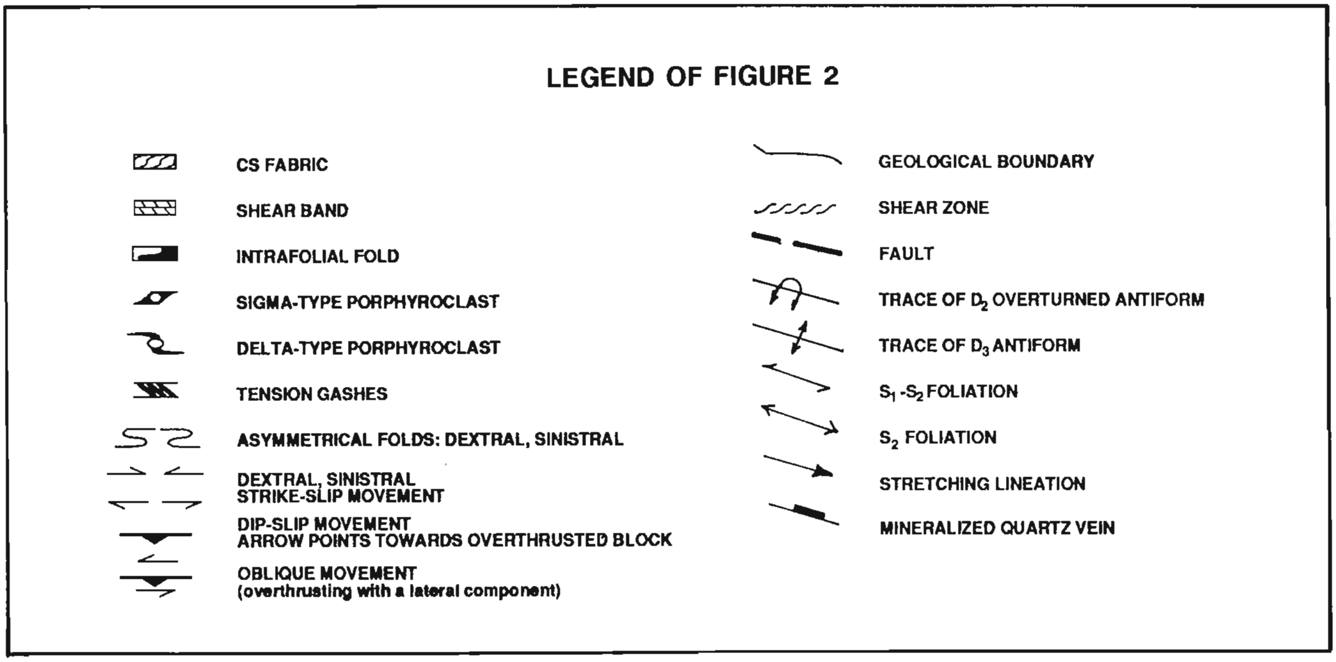


Figure 2. Geological map of the area between Big Pond and the Cape Ray deposit area (Main zone) illustrating the main fabric with observed stretching lineations, kinematic indicators, and a compilation of the main structural elements on equal area projections (lower hemisphere). Modified from the compilation map of Dolphin Explorations, after Wilton (1984) and work by the company.



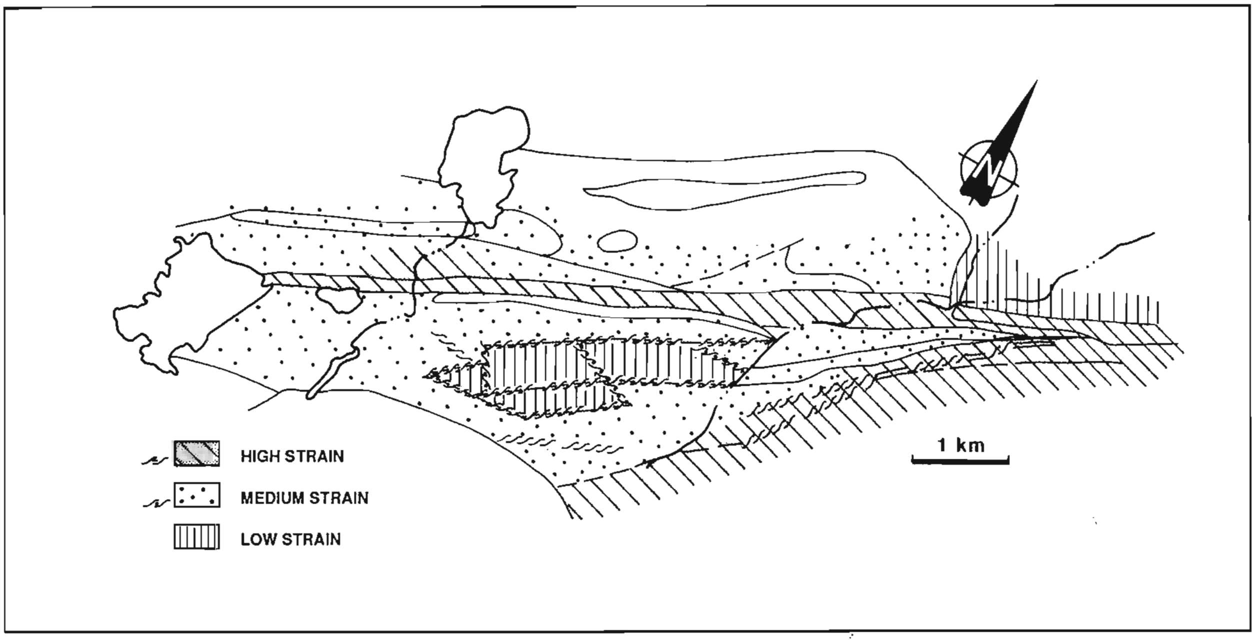


Figure 3. Map showing the high, medium, and low strain domains.

reverse or steeper dip than the main schistosity suggestive of "C and S" type fabrics. In plan view, the relationship between the S and C planes indicates either dextral or sinistral movement. The intersection between S and C planes is at a high angle with steeply southeast- or southwest-plunging stretching lineations (Fig. 2, stereonet J). Shallow dipping shear bands oriented at 062/17 are locally present. Isoclinal folds, with hinges trending 080/60° have their axial planes subparallel to the mylonitic foliation. They show a northwestward vergence in section. All kinematic indicators suggest that the "50 shear" is a reverse-oblique fault zone with a dominant vertical movement. In plan view, either an apparent sinistral or dextral lateral component can be seen and is consistent with the variation in orientation of the steeply plunging stretching lineation.

Due to lack of exposure, the relationship between the "50 shear" and the "main zone" shear that hosts the Cape Ray gold deposit is not clear. The "main zone" corresponds to the trace of a Windsor Point Group graphitic schist unit located in the footwall of the "A" vein and is defined by an electromagnetic conductor farther west. Underground work (Dubé, 1990) indicates that the hangingwall of the main mineralized vein in the "41" zone

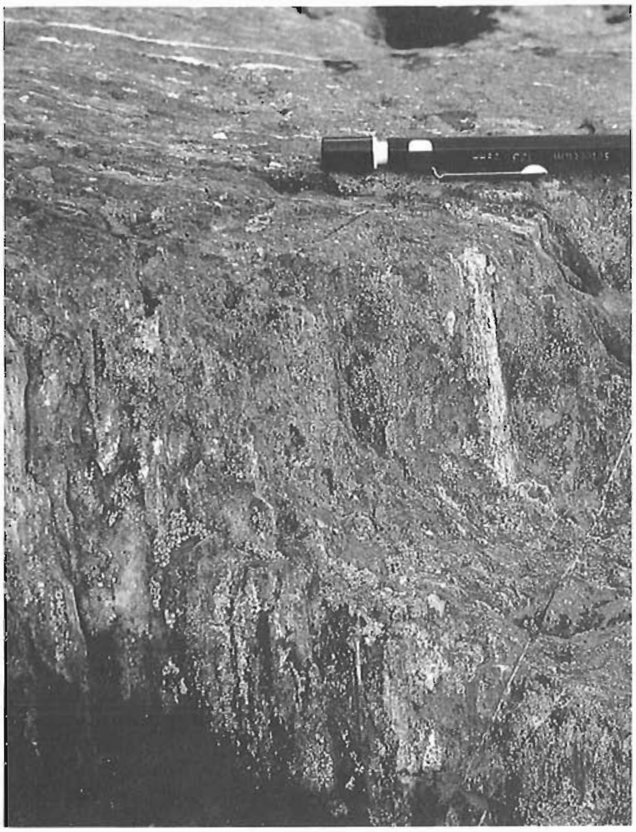


Figure 4. Longitudinal view along foliation planes showing the steeply plunging stretching lineation in microconglomerate of the Windsor Point Group in the Isle aux Morts River area.

is characterized by an oblique-reverse movement, followed by late brittle strike-slip movement which deformed the auriferous "A" vein. The "50 shear" and the "main zone" shear are almost subparallel and have probably recorded the same reverse-oblique ductile deformation event. In the deposit area, the two structures merge and could in fact represent the same structure which has branched out farther west due to the brittle nature and related strong anisotropy of the graphitic schist unit.

The "mylonite zone"

This zone occurs near the northwestern boundary of the Cape Ray fault (Fig. 3). The mylonite strikes 064° and dips 78° to the south. Its average width is 250 m. The internal fabric is dominated by a penetrative mylonitic foliation and/or a millimetre scale compositional layering. In the Cape Ray granite, the mylonitic fabric is associated with a grain size reduction. Shallow plunging to subhorizontal stretching and mineral lineations are present on the mylonitic fabric (Fig. 2, stereonet D). Striations with rare sinistral steps are also observed locally. Synshearing isoclinal folds with axial planes subparallel to the main foliation are common; their hinges are steep and show a dextral vergence. Late millimetre to centimetre scale folds associated with northwest-southeast-trending crenulation cleavages are also developed within the mylonite zone. They fold the mylonitic fabric and show a dextral vergence. Kinematics indicators such as porphyroclast rotation (Fig. 5), subvertical en echelon extension veins, synshearing fold vergence, and C-S-type fabric within chlorite-rich layers indicates both dextral and sinistral apparent movement along the mylonite zone.

Therefore, the "mylonite zone" is predominantly characterized by oblique to subhorizontal movement indicators which strongly contrast with the oblique-reverse movement recorded within the "50 shear" mylonite and the medium strain domain.

Two conjugate sets of well-developed, steeply dipping crenulation cleavages and/or kinks are locally present within rocks of the high strain domain. A first set trending N111°/78° is related to millimetre to centimetre scale S-shaped folds with associated, south-southeast steeply plunging crenulation lineations. Another set trends northeast (N020° to N055°). Associated Z-shaped folds plunge steeply south-southwest to southwest and locally curve into the main fabric. These structures are postmylonitic and are probably related to late increments of the ductile deformation.

Brittle deformation

Evidence of brittle deformation is well exposed along the mylonite zone in the vicinity of the Strawberry granite in what has been locally termed hydrothermal breccia. This zone is essentially characterized by the absence of any consistent planar fabric and by a strong low temperature silicification. Different facies of these brittle fault rocks (cataclasite) can be observed. The Strawberry granite, which is the protolith to many of these, has been dated by Wilton (1983) at 362 + 27 Ma. (Late Devonian to Early Carboniferous) using Rb-Sr whole-rock isochrons. It is mainly composed, in the immediate vicinity of the mylonite zone, of potassium feldspar phenocrysts up to 3 cm set in a finer grained pink-orange matrix of potassium feldspar and flattened quartz defining a foliation

Figure 5. Plan view of the mylonite derived from the Cape Ray granite, located at the northwest boundary of the Cape Ray fault zone. The porphyroclast rotation suggests a sinistral movement; lower edge of photograph = 10 cm.

striking 070°/85°. This foliation is locally crosscut and sinistrally offset by a spaced cleavage defined by chloritic planes. Towards the fault zone, the Strawberry granite passes into a fault breccia made up of rounded quartz and less abundant potassium feldspar porphyroclasts and orange-pink fragments set in a light green powdery chloritic matrix. The orientation of foliation within the larger granitic fragments is different from one fragment to another. Another facies is a rock made up of millimetre size quartz porphyroclasts within a silicious purple aphanitic matrix. To the southeast, these rocks grade into to a pale grey massive cherty rock. In several localities, a late stockwork of quartz and/or chalcedony veinlets is injected within these cataclasites. Chalcedony also forms the cement of a breccia mostly composed of angular to sub angular fragments of the grey silicious material and of some mafic fine grained rocks. Within the Windsor Point Group, brittle deformation is indicated by fault breccias composed of monolithic, angular to sub-angular fragments of strongly foliated rocks with a weak proportion of matrix. The orientation of the foliation is highly variable from fragment to fragment. Clear indications of movement along this brittle fault are relatively rare. Occasional shallow southwest-plunging striations along fault planes indicate a strike-slip movement.

According to Wilton (1983) and Chorlton (1983), the Strawberry granite is a late to post-tectonic intrusion. Although some plastic deformation did occur, indicated by the flattened quartz ribbons, no intense mylonitic foliation was seen in the Strawberry granite suggesting that it was emplaced after peak ductile deformation, which is therefore earlier than Late Devonian to Early Carboniferous (age of the granite). Later sinistral offset of the foliation along the late spaced cleavage indicates that the intrusion has been affected by late movements along the Cape Ray fault zone which could well correlate with the development of the brittle fault rocks.

The coastal section

A well exposed section of the Cape Ray fault zone outcrops along the coast near Cape Ray village (Fig. 1). This area has previously been studied by Wilton (1983) and by Piasecki (1989). We used it as a typical cross-section of the Cape Ray fault in order to understand the structural framework of the area of the Cape Ray deposit.

Two high strain zones outcrop along the coast: 1) a well-exposed mylonite zone in the southwestern part, and 2) a poorly-exposed mylonite zone in the central part of the area (Fig. 6). The southwestern mylonite is located at the contact between the Port aux Basques gneiss and the Windsor Point Group. As for the "50 shear", this zone is partly a retrograde shear within the gneiss. As shown by Piasecki (1989), prograde stretching lineations in the Port aux Basques gneiss plunge moderately to the southwest whereas stretching lineations produced during the post-Late Devonian retrograde shearing of the Port aux Basques gneiss plunge to the east-northeast at a relatively high angle to the older one (Fig. 6, stereonet B).

The well-exposed mylonite at the contact between the Port aux Basques gneiss and the Windsor Point Group strikes on average 054° and dips 72° south. Approximatively 750 m wide, it is mainly composed of beige to pink quartzo-feldspathic mylonites which could originate from a "granitic" protolith. The main fabric is a well-developed mylonitic foliation. Non-coaxial deformation is indicated by "C-S" fabrics showing a reverse-dextral movement. Stretching lineations plunge moderately to steeply to the east (Fig. 6, stereonet C). The mylonitic fabric is axial-planar to centimetre scale isoclinal folds with relatively steep fold axes (195°/55°) and a dextral-reverse structural vergence.

Northwest of the Port aux Basques gneiss and Windsor Point Group contact (Fig. 6), the quartzo-feldspathic mylonites pass into a zone of sericite-chlorite S-C mylonites (Lister and Snokes, 1984) showing well developed C-S synthetic shear bands (Fig. 7). The fabric shows a penetrative "C" foliation oriented approximately 062°/60° and a slightly oblique "S" foliation (045°/75°) is commonly preserved. The intersection between these two foliations is almost perpendicular to a strongly developed stretching lineation (062°/52°) present on the penetrative "C" foliation indicating that they can be used as evidence to a reverse dextral movement. Both asymmetric and sheath-like syn-shearing F_2 folds with an axial plane parallel to the main foliation and fold hinges parallel to the stretching lineation show a reverse and dextral vergence. Many sub-horizontal joints (165°/25°) are perpendicular to the stretching lineation.

Medium strain zone

To the northwest of the Port aux Basques gneiss and Windsor Point Group contact, the mylonite zone grades into a fold-dominated (medium strain) domain of sedimentary rocks, gabbro and mafic pillow lavas of the Windsor Point Group (Fig. 6). A strong lineation plunging approximately 50° east-northeast is defined by stretching of pebbles and amphiboles in the conglomerate and gabbro of the Windsor Point Group (Fig. 6, stereonet A).

Two sets of late axial plane cleavages and local kinks, similar to those described in the high strain zone near the orebody, are developed. Centimetre to metre scale sinistral verging folds are related to a well developed set trending approximately 090°/80°, whereas dextral verging folds, plunging moderately to the southwest (190°/38°) have a second axial plane cleavage oriented at 025°/80°. Striations that plunge shallowly to the southwest with local sinistral steps are locally present on the main fabric. Finally, local subvertical en echelon extension veins oriented at 300°/54° suggest a dextral motion.

Based on the observations in the coastal section, it appears that the post-Late Devonian deformation along the Cape Ray fault zone is compatible with a reverse-dextral shear. As demonstrated by Piasecki (1989), prior to that shearing event, the Cape Ray fault zone has recorded pre-late Devonian reverse-sinistral faulting between the Port aux Basques gneiss and the Cape Ray granite. The Windsor Point Group was deposited after the earlier deformation event.

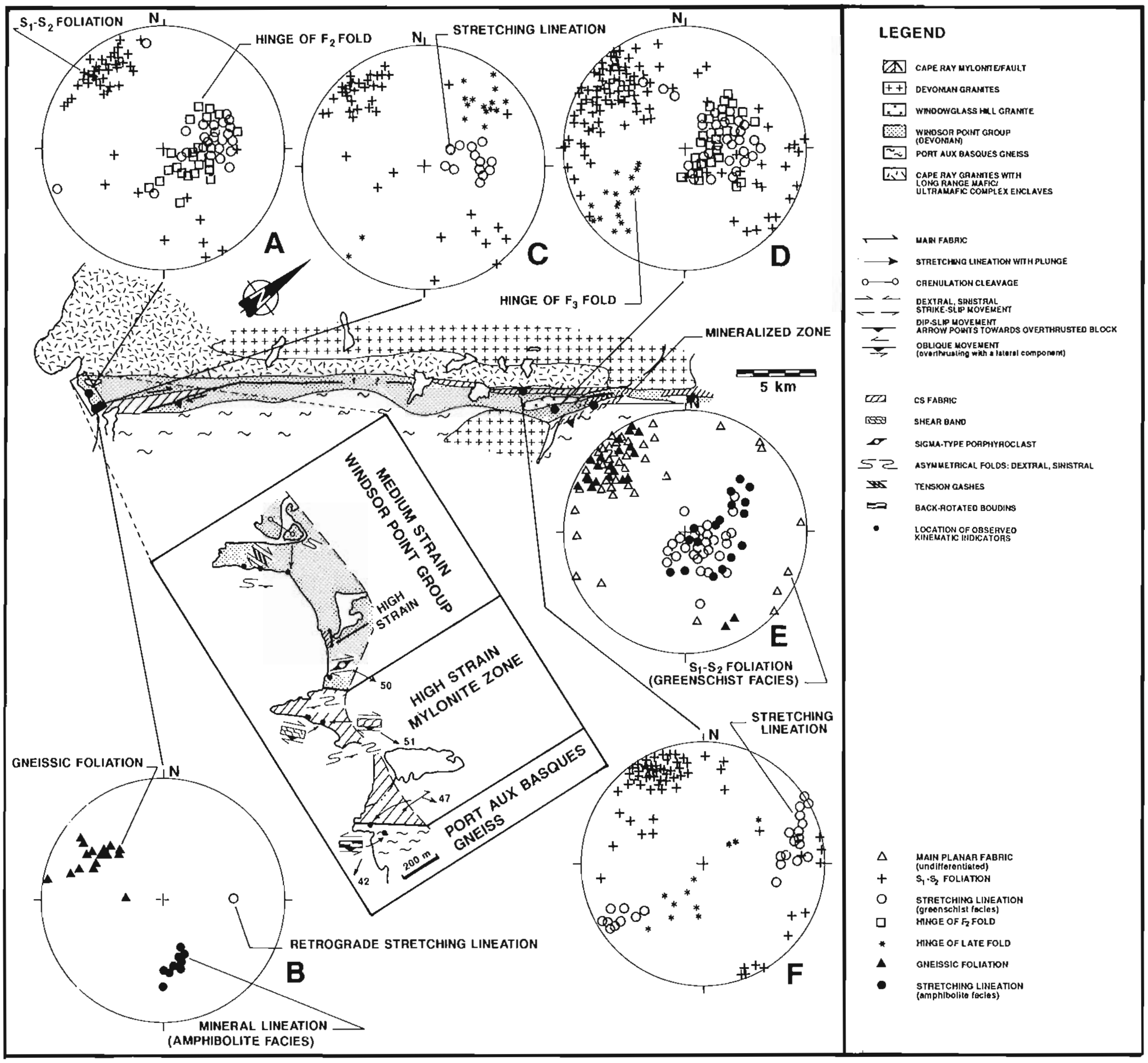


Figure 6. Compilation of the structural data from the Cape Ray deposit area and the coastal section (equal area projections on the lower hemisphere). Map modified from Tuach et al. (1988); after Wilton (1983) and Tuach (1986).

The coastal section across the fault is similar to that across the "50 shear" in the deposit area, whereas the high strain zone, characterized by sub-horizontal movement in the northwestern part of the deposit area seems to be absent in the coastal section (Fig 6, stereonets B and F).

GOLD MINERALIZATION

The gold occurs mainly in galena and chalcopyrite-rich quartz veins hosted by shear zones at or near the structural top of a graphitic unit of the Windsor Point Group. The main deposit consists of three mineralized zones (4, 41 and 51) occupying the same structure known as the "main zone". Other significant auriferous quartz veins occur in the Windowglass Hill granite (Wilton and Strong, 1986) and in the recently discovered Big Pond showing hosted by the Windsor Point Group. This part of the paper focuses on the last two significant gold occurrences.

Windowglass Hill

The Windowglass Hill granite is a key unit to establish the timing of the gold mineralizing event in the Cape Ray area. Due to its high competency contrast with surrounding rocks, this syn-volcanic granite (Wilton, 1983) acted as a resistant core which seems to have been relatively protected from late tectonic events, and thus has preserved the fundamental relationships on the timing and the structural control of the mineralization. Detailed mapping of the Windowglass Hill showings indicates that the main mineralized quartz veins are pyrite-rich (± chalcopyrite ± galena) sub-horizontal extension veins

Figure 7. Section view of shear bands in sericite-chlorite schists of the coastal section, showing evidence of a reverse movement.

(195°/20°) enclosed within discrete ductile-brittle shears oriented approximately 350°/70°. These mineralized veins are almost perpendicular to a well developed stretching lineation (104°/75°) associated with the hosting shear (Fig. 2, stereonet I), as previously mentioned by Piasecki (1989). Stretching lineations are sub-parallel to those characterizing the reverse-oblique ductile shearing and measured, in the Cape Ray deposit area, in the medium strain domain and in the "50 shear" of the high strain domain, as well as along the coastal section. This observation suggests that the main mineralized veins in Windowglass Hill are coeval with the reverse-oblique shearing event. Similar mineralized extension veinlets showing the same relationship with stretching lineations have been observed in the H Brook showing (Fig. 2, stereonet E). Other, mineralized shear-veins, some oriented sub-parallel to the main northeast-southwest S_1-S_2 fabric and some oblique to it (280°/53°), as well as the associated mineral lineations (050°/50°), are compatible with this interpretation.

Big Pond

Detailed mapping at the Big Pond showing reveals that the main mineralized quartz vein is pyrite-rich (+chalcopyrite, galena, arsenopyrite) and oriented approximately at 350°/70° (Fig. 8). The vein is mainly hosted within a graphitic unit of the Windsor Point Group, and contains pyritic veinlets sub-parallel to the wall of the vein and fragments of foliated surrounding rocks. No significant foliation sub-parallel to the vein has been developed in the adjacent schists.

The footwall of the mineralized vein and graphitic host is gabbroic whereas the hangingwall is mainly composed of sericite and/or chlorite schists. All units bear a bedding-parallel penetrative S_1 foliation. S_{0-1} is folded by asymmetric F_2 folds that plunge to the northeast (Fig. 8, stereonet C) and that are showing sinistral vergence in the western part of the area (Fig. 8), and dextral vergence to the southeast. The variation in the asymmetry of these small scale F_2 folds is interpreted as parasitic folds of a synform with an axial plane sub-parallel to S_2. Stretching lineations on both S_{1-2} schistosity planes are sub-parallel to the fold hinges (Fig. 8, stereonet B). All these structural elements are similar to those characterizing the reverse-oblique ductile movement of the high strain domain, as well as those described earlier from the medium strain domain along the Isle aux Morts River.

The mineralized vein is oblique to both S_1 and S_2 foliations (Fig. 8, stereonet A), but it is not clear if the mineralized vein has been folded or not. However, in the southern part of the main trench, the vein does not cut through a graphitic zone in an area where a well defined F_2 fold closure is observed, suggesting that it may have also been folded.

In one area, the mineralization is located within a distinct mineralized breccia made up of centimetre-scale angular, sericitic wall rock fragments with a penetrative fabric. The orientation of the foliation is relatively constant from one fragment to another and is sub-parallel to the lithological contacts and structural relationships with surrounding rocks suggest that the fabric in the fragments is the S_1 schistosity.

The mineralized vein at Big Pond is post-S_1 but could be pre to syn-F_2. The mineralized vein is sub-parallel to many topographic lineaments in the area. It is not sub-parallel to S_2 but rather is oblique to it.

DISCUSSION

The Cape Ray fault zone is a long-lived structural break that records superimposed deformational events (Brown, 1976; Chorlton, 1983, Piasecki, 1989). However, detailed structural work in the deposit area shows that the post-Late Devonian movement along the Cape Ray fault zone is compatible with an overall dextral transpressive regime (Sanderson and Marchini, 1984). As proposed by Robert (1989) for the Archean Cadillac tectonic zone, we suggest that the deformation zone has evolved from a deformation zone with a significant shortening component to a zone dominated by a transcurrent shearing component.

The ductile deformation can be subdivided into two strain increments. The first increment, northwesterly directed shortening, is recorded in the "50 shear" and the medium strain domain. It is characterized by structures indicating dip-slip to oblique-slip movement such as relatively steep stretching lineations, S-C fabric and shear bands present in section view, fold hinges sub-parallel to the stretching lineations and vergence of F_2 folds. The second strain increment is documented in the high strain "mylonite zone" near the northwest boundary of the Cape Ray fault zone, and is characterized by strike slip structures, such as shallow plunging stretching lineations.

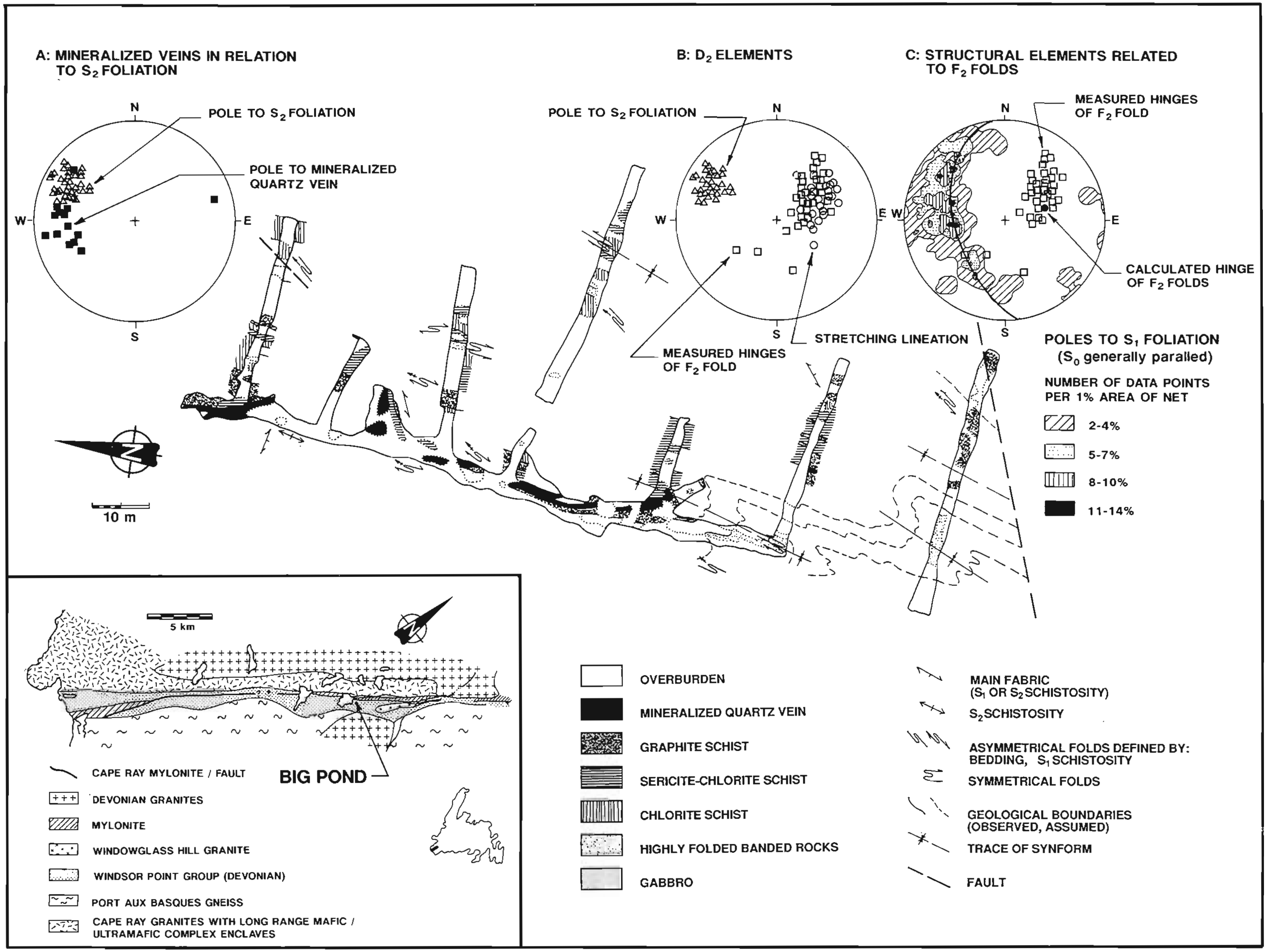


Figure 8. Geology of the Big Pond showing with a compilation of structural elements on equal area stereographic projections (lower hemisphere).

Other indications of the possible dextral transpressive regime, is the obliquity of the S_2 foliation and F_2 fold axes to the overall orientation of the mylonite zone and the fault itself (Fig. 2). Also, the main shear, the "50 shear" and other topographic lineaments, could be interpreted as oblique splays to the main fault zone direction. The orientations of these splays fit better with a transcurrent movement model than with a dip-slip reverse faulting model.

In the coastal section, the exposure of the fault zone is not as complete as along Isle aux Morts River. Stretching lineations in the high strain mylonites consistently plunge moderately steeply and do not suggest a superimposed strike slip movement as along Isle aux Morts River. This indicates that evidence for a dextral transpressive regime is not present along the entire length of the Cape Ray fault. As it is a reactivated fault zone, different deformation regimes may result from the previous geometry of the fault zone. The flexure of the Cape Ray fault zone to the northeast of the studied area (Fig. 1), may be responsible for the variation of the deformation regime and the development of strike slip structures in the Isle aux Morts River study area.

It is also possible that the two apparent strain increments were produced by two unrelated phases of deformation. The first, of Acadian age, characterized by ductile dip-slip to oblique-slip and the second, still ductile, but characterized by strike-slip structures. However, because of the coaxiality between elements of the dip-slip and strike-slip structures, we favour the view that the observed geometry is related to one progressive deformation event.

Based on the observations presented in this study, we suggest that the gold mineralization at the Windowglass Hill and Big Pond showings was deposited in a syn- to late-ductile shearing event and as such, is genetically and spatially related to an Acadian reactivation of the Cape Ray fault zone. The mineralization in those two areas is coeval with the main ductile, post-Late Devonian deformation. Future work will be oriented more specifically on mineralization located in the "main zone" area in order to confirm or refute that model.

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A preliminary report on the relationship between mineralization and carbonate diagenesis in the Gays River Formation, Nova Scotia

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Savard, M., A preliminary report on the relationship between mineralization and carbonate diagenesis in the Gays River Formation, Nova Scotia; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 147-156, 1991.

Abstract

Preliminary petrographic investigations of the Gays River Formation established distinct diagenetic successions in the Antigonish and Mahone Bay areas and in the Musquodoboit and Shubenacadie sub-basins. In the latter, the succession is marine cementation, pervasive microdolomitic replacement, partial dissolution, dolomite cementation, fracturing-brecciation-mineralization, sulphate precipitation, calcite cementation and stylolitization. In cathodoluminescence, the replacement dolomites are uniformly dull, the dolomite cement is nonluminescent whereas the late calcite comprises a zoned pattern and a luminescent one. In Antigonish and Mahone Bay areas, dolomitization had minor effects but leaching was intense. Marine cementation, fracturing, brecciation and coarse calcite were observed in all the studied areas. Genetic processes suggested for dolomites associated with the Gays River and six other deposits are briefly compared. This revives the unresolved fundamental problem concerning the role of dolomitization in Mississippi Valley Type deposit genesis.

Résumé

Des travaux pétrographiques préliminaires permettent de distinguer deux séquences diagénétiques dans la Formation de Gays River, l'une commune aux régions de Mahone Bay et Antigonish et l'autre propre aux sous-bassins de Musquodoboit et Shubenacadie. Dans ces derniers se succèdent cimentation marine, remplacement microdolomitique pénétratif, dissolution partielle, cimentation dolomitique, fracturation-bréchification-minéralisation, précipitation de sulfates, cimentation calcitique et stylolitisation. En cathodoluminescence, les dolomies de remplacement sont uniformément ternes, le ciment dolomitique est nonluminescent et la calcite tardive comprend un patron zoné et un accroissement luminescent. Dans les régions de Mahone Bay et d'Antigonish, la dolomitisation a eu un effet mineur mais le lessivage a été intense. Les cimentation marine, fracturation, bréchification et calcite tardive apparaissent dans toutes les régions étudiées. Les processus génétiques proposés pour les dolomies des dépôts de Gays River et de six autres régions connues sont comparés succinctement. Cet exercice rappelle que l'évaluation du rôle joué par la dolomitisation dans la genèse des gisements de type vallée du Mississippi reste un problème fondamental irrésolu.

INTRODUCTION

As all reported Mississippi Valley Type (MVT) lead-zinc deposits are hosted exclusively by dolostones, dolomitized successions constitute a valuable metallotect for exploration geologists. It therefore is essential to establish whether or not the dolomite is related to the ore-forming solutions. It seems critical as well to study the relationship between mineralization and diagenetic processes such as cementation, dissolution and stylolitization. Migration of ore-bearing solutions must be placed within the entire diagenetic scenario, not only locally next to the deposits but, ideally, on a regional scale. This represents the "diagenetic approach", differing from the common approach to MVT metallogeny which emphasizes determination of source, transportation and deposition of base metals.

The diagenetic approach recently became popular and is now applied to specific problems of many known carbonate-hosted deposits (for a brief summary: Mazzulo and Gregg, 1989). Among others, the research by Gregg (1985, 1988) shows the relevance of regional diagenetic studies. For example, Gregg proposed that the off-reef facies of the Bonneterre Formation, host rocks to the southeast Missouri lead-zinc district, were dolomitized by ore forming fluids, implying that the area affected by mineralizing processes is considerably larger than the deposit itself.

The Gays River lead-zinc deposit is hosted by dolostones of the Carboniferous Windsor Group, on mainland Nova Scotia. Many studies have dealt with investigations of base metal sulphides or local diagenetic studies of gangue minerals. Akende and Zentilli (1984) and Ravenhurst et al. (1989) have published geochemical data on Gays River carbonates. They concentrated on the attributes of the sulphides but, they also investigated the oxygen, carbon and strontium isotopes and fluid inclusions of a few dolomites and calcites sampled at the deposit. Hitherto, there are no reported regional diagenetic studies of the carbonates.

Through a diagenetic approach and on a regional scale, the aim of this project is to identify the different fluids that affected the Gays River Formation. The work will consist first, in the identification of the different dolomite and calcite phases and their relative timing, and then, in the characterization of every generation with geochemical tracers ($\delta^{18}O$, $\delta^{13}C$, $^{87}Sr/^{86}Sr$, Mg, Sr, Fe and Mn).

This paper is a preliminary petrographic report that contains a description of the calcite and dolomite phases, and shows their chronological relationship to mineralization. It also compares the main diagenetic features of the Gays River deposit with those of six other MVT deposits.

GEOLOGICAL SETTING

The Windsor Group was deposited in several areas of Nova Scotia; among others, these are the Antigonish, Minas, Shubenacadie and Musquodoboit sub-basins, and Mahone Bay area, all parts of the larger Maritimes Basin (Roliff, 1962; Fig. 1). During the Late Devonian and following the Taconian and Acadian orogenies, the basin developed by rifting of the Cambrian-Ordovician sedimentary sequence (Meguma Group). Early Carboniferous marine transgressions led to the deposition of carbonates and evaporites of the Visean Windsor Group. Maritimes and Palisades disturbances later affected the Windsor Group which then became exposed to meteoric influences by Mid-Cretaceous.

The Pb-Zn deposits of the Gays River area occur in the Gays River Formation of Major Transgressive-Regressive Cycle 1 (Giles, 1981b), in the Shubenacadie and Musquodoboit sub-basins. Unconformably overlying Ordovician and Devonian rocks, the formation is composed of carbonates which form a part of the Windsor Group "basal carbonates". In turn, a thick succession of evaporites-carbonates-clastics forming major cycles 1, 2, 3, 4 and 5 covers the Gays River Formation (Fig. 2).

Gays River Formation

Bell (1929) defined the Windsor Group in which he recognized subzones A, B, C, D and E. More recent stratigraphic work led to the definition of smaller units recognizable at a regional scale; in Subzone A, the Gays River and Macumber formations are popularly referred to as "Basal Windsor Carbonates".

Massive dolostones with well preserved textures of the precursor limestones, at the base of the Windsor Group, have been defined as the Gays River Formation by Giles et al. (1979). The type section is core GR-256, drilled in the vicinity of the Gays River mine (Fig. 1). At the very base of the Gays River Formation, the discordant Meguma-Windsor contact is angular and marked by a spectacular carbonate conglomerate containing large boulders of slate or sandstone. Occasional basal contact with Devonian granitic rocks is also discordant.

The Gays River Formation contains bank and interbank facies. Interbank facies are composed of fine, sparsely fossiliferous and vuggy dolostones. The large banks are locally developed on northeasterly-southwesterly trending basement highs of the Nova Scotia Platform (Boehner et al., 1989). They are mollusc-rich and algal-rich and texturally composed of algal bindstones, coralgal bafflestones and skeletal packstones. Dasycladaceae algae played a primary role in their production (Ryan and Moore, in preparation). Characteristically, the Gays River biota are restricted, the banks lack costate and plicate brachiopods and crinoids, three fossils notably abundant in younger carbonate banks of the Windsor Group, (Ryan and Moore, in preparation).

Fine peloidal laminites dolomitized and, in places, up to 6 metres thick, have been defined as the Macumber Formation by Weeks (1948). The laminites are locally brecciated (Pembroke Breccia). The depositional environment of these fine laminites has been debated by stratigraphers and sedimentologists. Schenk (1967) first interpreted these beds as being ancient analogues of modern supratidal stromatolitic facies whereas Giles and Boehner (1982) simply suggested that they are deep water fine cryptalgal or cyanobacterian laminites. More recently, Schenk (1990) proposed that they are deep water

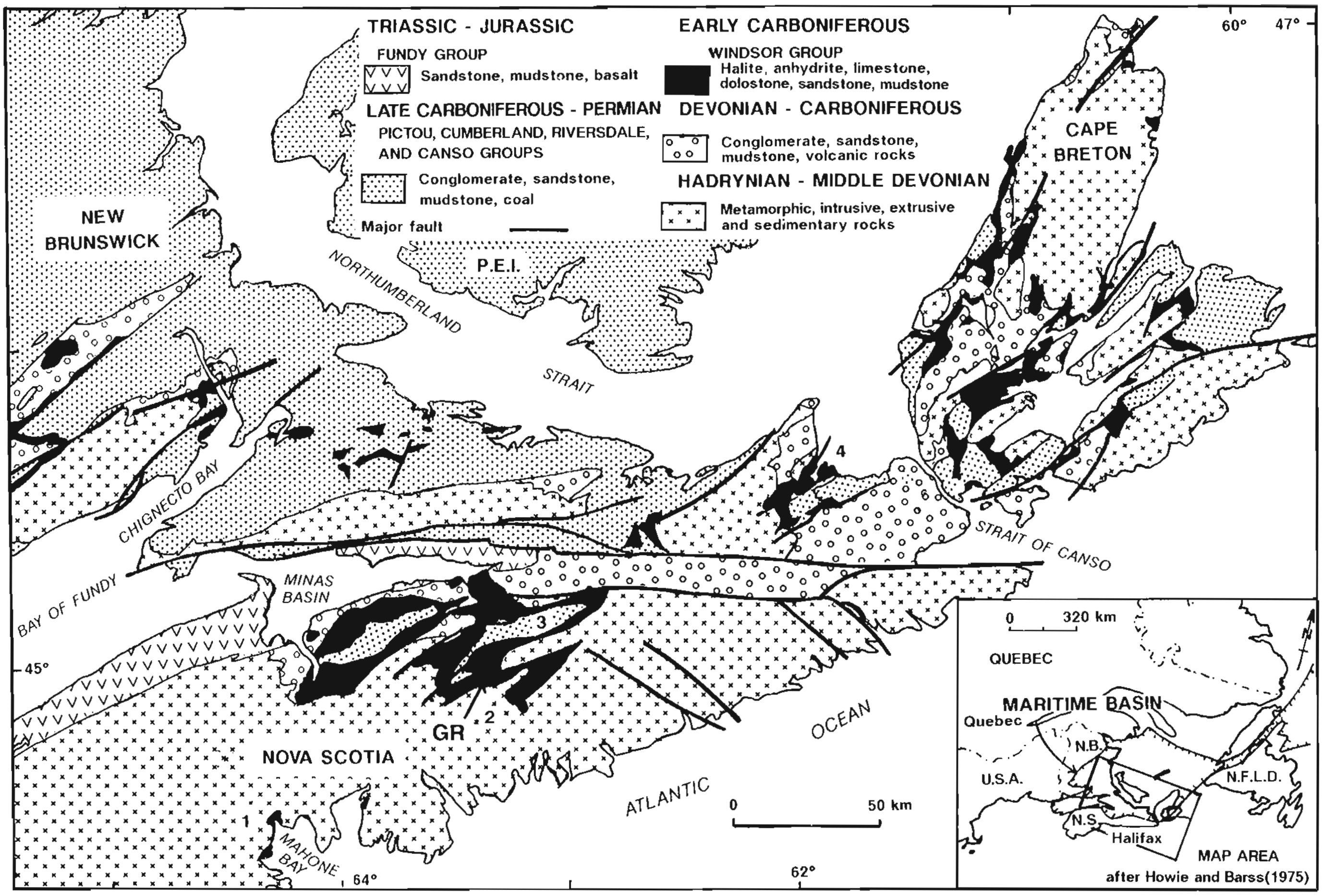

Figure 1. Geology and location map (modified after Boehner et al., 1989). The studied areas are: Mahone Bay (1), Musquodoboit (2), Shubenacadie (3) and Antigonish (4). GR indicates the Gays River deposit. The inset diagram shows the limits of the Maritimes Basin.

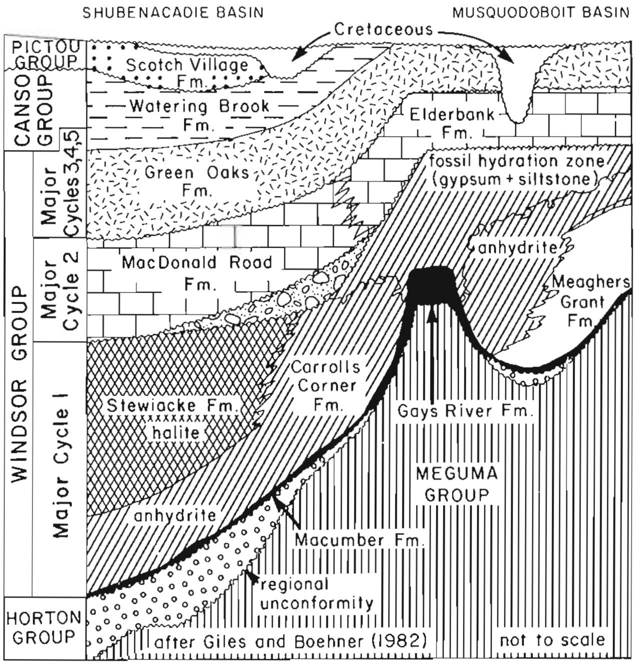

Figure 2. Stratigraphic relations of the Windsor Group in the Musquodoboit and Shubenacadie sub-basins (from Giles and Boehner, 1982).

methane-related structures. So now, although genetic processes are still disputed, most agree on a basinal sedimentary environment.

Macumber and Gays River formations are laterally equivalent. They represent deep water and shallow water units, respectively. Macumber strata rest on conglomerates and sandstones of the Tournaisian Horton Group, whereas the Gays River Formation generally is in discordant contact with pre-Carboniferous strata (Fig. 2). Locally, the relief of the basement reached 700 to 800 m, with the Gays River units perched on the highs and the Macumber units blanketing the lower parts (P. Giles, pers. comm.).

According to Schenk and Hatt (1984) and Boehner (1984), deposition of the Gays River Formation is the product of a rapid marine transgression. The succeeding regression stopped the growth of the banks and gave place to the evaporites.

In the Mahone Bay and the Antigonish areas, limestones with biogenic constituents identical to those of the Gays River area, but only partly dolomitized, were assigned to the Gays River Formation (Giles, 1981a; Boehner, 1980). Distribution of facies in the Musquodoboit and Mahone Bay areas suggests that both regions recorded shallow marine shelf sedimentation but on facing flanks of a Nova Scotia basement arch (Giles, 1981a). In the Antigonish area, sedimentation of the Gays River Formation took place within a sub-basin for which the paleogeography is obscured by thrust faults (Boehner, 1986).

DISTRIBUTION OF SAMPLES

The Gays River Formation in the Mahone Bay area and Antigonish, Shubenacadie and Musquodoboit sub-basins was sampled for its carbonate diagenetic phases (Fig. 1). In the Shubenacadie and Musquodoboit sub-basins, samples were selected from fifty cores and two quarries, in Mahone Bay area, from four cores and two quarries and, in Antigonish sub-basin, from five cores, three outcrops and one quarry.

PETROGRAPHIC TECHNIQUES

To date, 150 thin sections were systematically observed under a light microscope. They were given a fine polish and examined in cathodoluminescence (CL) using a Nuclide Corporation, model EEM2E instrument. The operating conditions were a voltage of 10 kV, an intensity of 0.5 mA, and a beam focused at about 6 mm. Mixed Alizarin Red S and potassium ferrocyanide solutions were used to distinguish calcite from dolomite and to quantify the iron content of seventy five hand specimen. For fine calcite-dolomite intergrowths, the luminoscope greatly helped to decipher mineralogy. Scanning Electron Microscope (SEM) petrography of microdolomites has been performed on broken surfaces of ten selected samples.

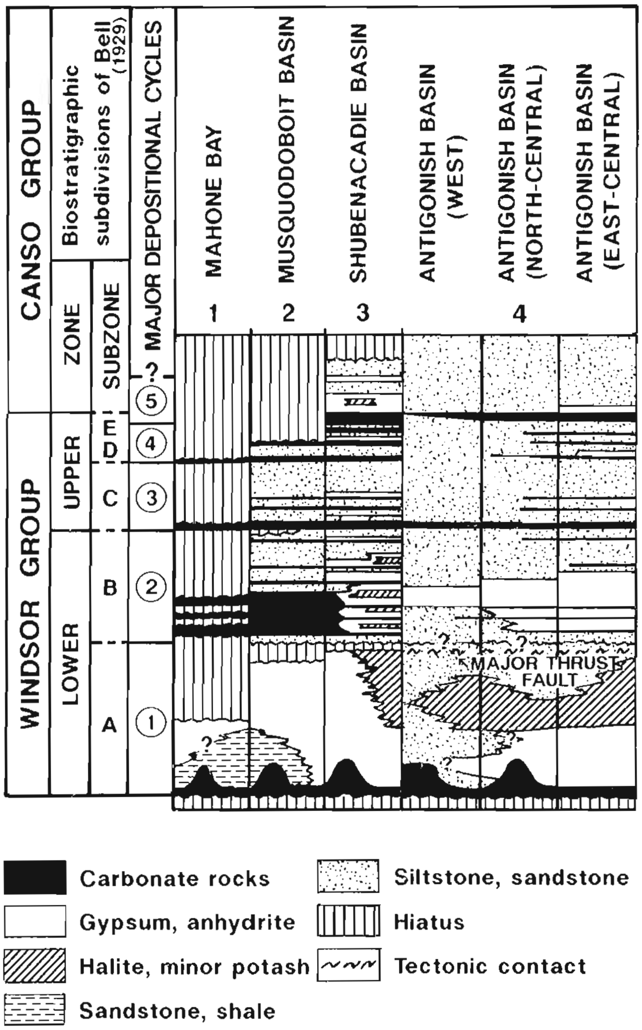

Figure 3. Lithostratigraphy and major cycles of the Windsor Group in the studied areas (modified from Boehner, 1986). See Figure 1 for locations.

DIAGENETIC SUCCESSIONS

For the studied areas, two regional diagenetic successions have been recognized. In the Musquodoboit and Shubenacadie sub-basins, pervasive dolomitization and lead-zinc mineralization were the dominant processes, whereas in the Mahone Bay and Antigonish areas, intense leaching characterizes the succession.

Musquodoboit and Shubenacadie sub-basins

In the Musquodoboit and Shubenacadie sub-basins, the Gays River Formation is exclusively composed of dolomite with minor amounts of sphalerite, sulphates and calcite. These phases succeeded each other in the following fashion: (1) fibrous isopachous early marine cements, (2) replacive microdolomites closely followed by partial dissolution and corrosion, (3) pore filling euhedral rhombohedric dolomite, (4) sphalerite, contemporaneous with fracturing and brecciation, (5) sulphates, and (6) xenomorphic calcite prior to stylolitization.

In the largest fenestral and interorganism pores, inclusion-rich, early marine cements (EMC) were the first diagenetic products to occur. Although entirely dolomitized, their primary fine fibrous and isopachous texture is partly preserved and perhaps belongs to a High Magnesium Calcite (HMC) precursor.

In hand specimens, staining tests show that dolomitization did not affect textures of the marine components; bryozoans, algal molds and cements can still be identified. However, on micron-scale, with SEM, a mosaic of anhedral, subhedral to euhedral rhombohedric microdolomites obliterated all component ultrastructures (Fig. 4, 5). In cathodoluminescence, these replacement microrhombs form a uniformly dull mass. It is clear that dolomitization occurred very early, closely following marine cementation and preceding any sign of shallow burial cementation. Partial dissolution closely followed this generation of dolomite (Fig. 5). In turn, the succeeding pore filling dolomite, coarser and perfectly rhombohedric, invariably postdated dissolution but predated precipitation of sphalerite (Fig. 6). In cathodoluminescence, this generation of dolomite is generally nonluminescent, rarely dull.

Fracturing, brecciation and precipitation of sphalerite (and minor galena) appear as contemporaneous events which affected dolomitized marine components (Fig. 7). Sphalerite occurs as yellow to brown translucent fine crystals that are nonluminescent or green in cathodoluminescence. Crystals often show geopetal structures, occurring as lining on the bottom of cavities or on the top of micro-breccias.

Calcite in inclusion-free, large (up to one centimetre) xenomorphic crystals is the last cement to precipitate (Fig. 7). It fills micro-fractures and encompasses brecciated fragments, sphalerite and rare occurrences of sulphate (barite). The calcite can be iron-free or iron-rich. In cathodoluminescence, two distinct zones are ubiquitous: a first zone containing alternating dull to luminescent fine layers and a subsequent zone made of uniformly luminescent overgrowths (Fig. 8). Stylolites invariably postdate this last generation of cement.

Locally, pore-filling dolomite alternating with ankerite postdates dolomitized and corroded early marine cements. The ankerite is in turn closely followed by fracturing and migration of hydrocarbons as indicated by droplet-like bitumen occurring in micro-fractures and centre pores (eastern Musquodoboit sub-basin).

Primary porosity of the bank limestones was visually estimated to be 25% and has remained essentially unchanged by replacement dolomitization. Precipitation of marine calcite and dolomite cements occluded 5 and 1 % of porosity, respectively, whereas pre-ore dissolution increased it by a maximum of 2%. Prior to sphalerite and galena precipitation, porosity was about 21% and has been reduced to about 4% by the precipitation of xenomorphic calcite.

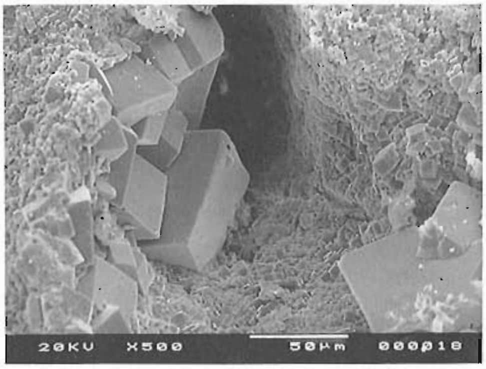

Figure 4. Euhedral replacive and pore-filling microdolomite around a fenestral cavity. From Shubenacadie and Musquodoboit sub-basins. SEM.

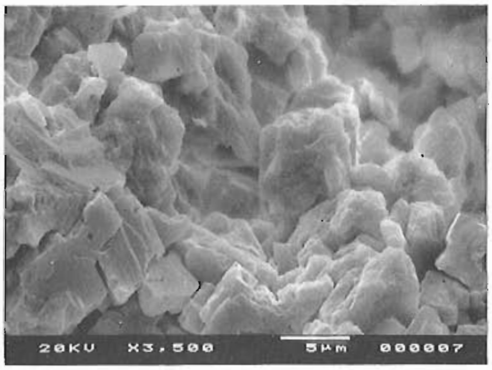

Figure 5. Corroded subhedral replacement dolomite. From Shubenacadie and Musquodoboit Sub-basins. SEM.

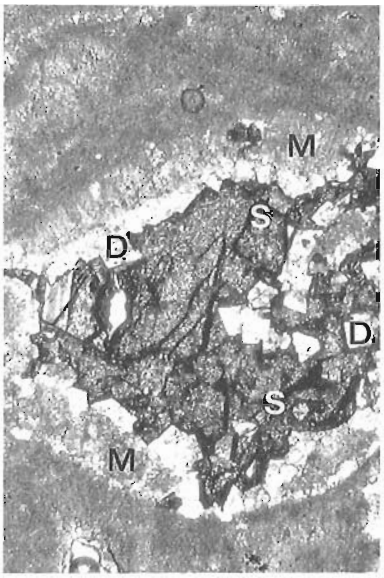

Figure 6. Dolomitized early marine cement (M) partly dissolved and covered by euhedral pore-filling dolomite (D). Centre pore is filled with translucent sphalerite (S). From Shubenacadie and Musquodoboit sub-basins. Plane-polarized light. Length of photograph is 1.7 mm.

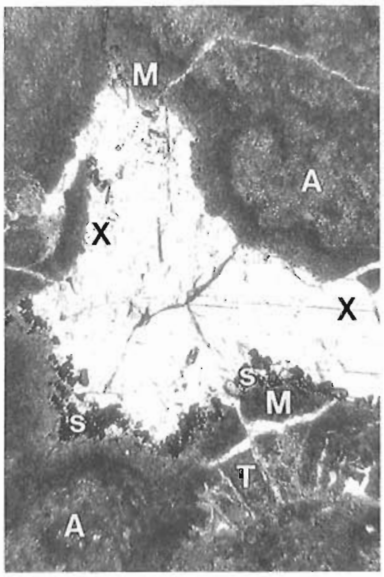

Figure 7. Growth cavity in an algal bindstone. Recrystallized algae (A). Dolomitized mud (T) and early marine cement (M) form the brecciated fragments. Sphalerite (S) is either displaced by or fallen over micro-fractures. Xenomorphic calcite (X) envelopes brecciated fragments. From Shubenacadie and Musquodoboit sub-basins. Plane-polarized light. Length of photograph is 3.5 mm.

Mahone Bay Area and Antigonish sub-basin

The Mahone Bay and Antigonish banks are composed of calcite with minor dolomite and traces of sulphates. Giles (1981a) reported traces of galena in dolomitic rocks, but sphalerite and galena were not found in the samples examined to date in this study. The diagenetic phases occurred successively as follows: (1) early isopachous and splayed fibrous marine cement, (2) leaching and partial dissolution of algae and early cements, (3) fracturing, (4) sulphates, (5) xenomorphic calcite and (6) stylolites.

The marine cements are iron-free, inclusion-rich and, although their crystalline shape is not recognizable, they still show an isopachous fabric. Generally, they are blotchy or moderately luminescent in cathodoluminescence, as is their marine substratum (mud and biogenic components). They are undeniably recrystallized. Only locally were these cements spared from thorough recrystallization, as shown in cathodoluminescence by still visible nonluminescent fibres arranged in splays or isopachous layers (Fig. 9). Due to dolomitization, such primary cathodoluminescence features are unknown in the Musquodoboit-Shubenacadie area.

The dissolution event that affected the early cements as well as the other marine components proceeded either through nonselective solution (Fig. 10) or intense leaching of marine components (Fig. 11). Micro-fracturing and local brecciation, followed by xenomorphic calcite cementation, and finally, stylolitization, also affected the carbonate sequences. The late xenomorphic calcite surrounds brecciated fragments, fills micro-fractures and centre pores. It is inclusion-free and iron-free to iron-rich. In cathodoluminescence, it shows a pattern of alternating dull to luminescent layers followed by a uniformly, more frequent, luminescent pattern (Fig. 9, 11).

Very rare occurrences of iron-rich dolomites are found locally in the Antigonish Sub-basin. The dolomites are euhedral and postdates the marine cementation. In Mahone Bay area, replacive microdolomite occurs locally

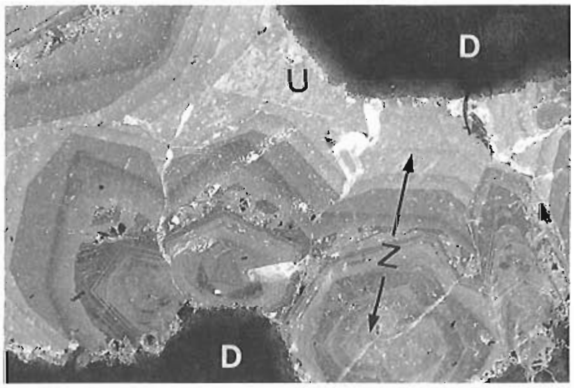

Figure 8. Uniformly dull microdolomite (D) after marine cement and mud. Coarse crystalline calcite with zoned (Z) and uniform (U) cathodoluminescence pattern. From Shubenacadie and Musquodoboit sub-basins. Cathodoluminescence. Width of photograph is 2.75 mm.

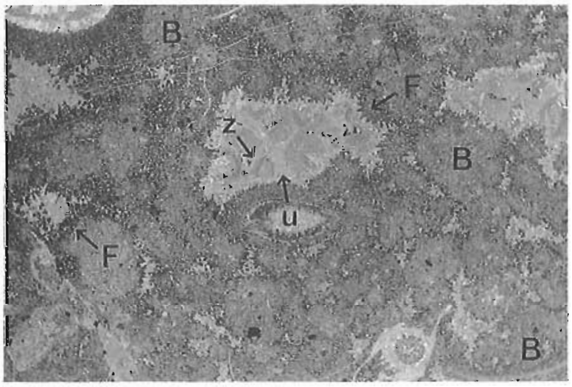

Figure 9. Luminescent-dull blotchy marine components (B) covered by nonluminescent fibrous marine cement (F). Pore centres are occluded by coarse luminescent to dull zoned calcite (Z) and uniformly luminescent calcite (U). From Mahone Bay area. Cathodoluminescence. Width of photograph is 2.75 mm.

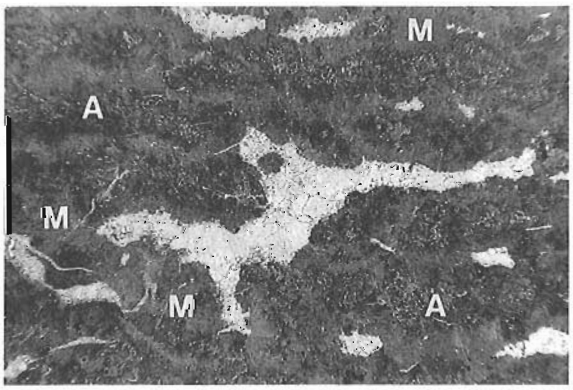

Figure 10. Secondary cavity produced by nonselective dissolution. Cavity walls cut through early marine cement (M) and recrystallized algae (A). Coarse calcite fills the pores. From Mahone Bay area. Plane polarized light. Width of photomicrograph is 2.5 mm.

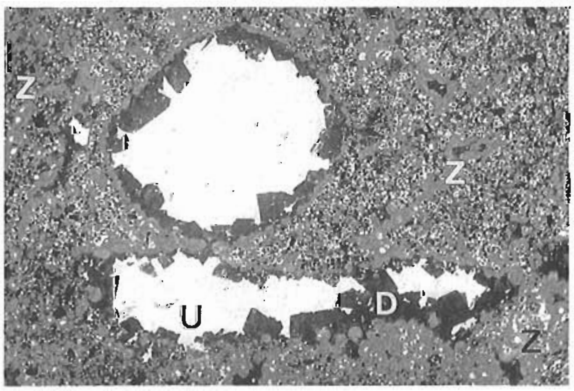

Figure 11. Replacement microrhombohedral dolomite with zoned cathodoluminescence pattern (Z). The succeeding pore filling dolomite is uniformly nonluminescent (D). Centre pore is filled with uniformly luminescent calcite (U). Bottom mold results from biogenic component leaching. From Mahone Bay area. Cathodoluminescence. Width of photograph is 2 mm.

Table 1. Sequences of carbonate phases obtained for the Gays River Formation. Adjectives in parentheses refer to carbonate CL pattern (NL = nonluminescent; LUM = luminescent). Normal marine cements are excluded from the sequences.

SCALE OF THE STUDY	CARBONATE PHASES	RELATIVE TIMING	SUGGESTED ORIGIN	AUTHORS
Local, mine area	Replacive microdolom.	Early	Evaporitic sea waters	MacLeod, 1984
	Replacive microdolom.	Early	"	
	Dolomite cement	Early	"	
	Coarse calcite cement	Post-ore	Meteoric water (vadose)	
Local, mine area	Replacive microdolom.	Very early, pre-ore	Mixing of saline and meteoric water	Schenk and Hatt 1984
	Coarse calcite cement	Post-ore	Meteoric water	
Local, mine area	Dolomite and dedolom.	Pre-ore	Not discussed	Akende and Zentilli, 1984
	Calcite cement	Ore-stage	Hydrothermal fluids	
	Calcite cement	Post-ore	Cooling hydrothermal fluids	
Regional scale	Replacive microdolom. (dull)	Early, pre-ore	Work in progress	This study
	Dolomite cement (NL)	Pre-brecciation, pre-ore	"	
	Coarse calcite cement (zoned)	Post-ore	"	
	Overgrowth calcite (LUM)	"	"	

where it follows marine cementation and is postdated by pore filling dolomite and late calcite. The replacive dolomite is iron-free, inclusion-rich and euhedral. In cathodoluminescence, it is zoned with luminescent cores and dull overgrowths whereas the pore-filling dolomite is uniformly nonluminescent (Fig. 11).

Apart from pervasive early dolomitization, which is restricted to central Nova Scotia, numerous diagenetic features are common to carbonates from Antigonish, Mahone Bay and Musquodoboit-Shubenacadie areas: early marine cements (perhaps HMC), partial dissolution, fracturing and brecciation, late calcite (zoned) and late calcite (uniformly luminescent). However, at this stage of the study, it is too early to speculate either on the diagenetic processes that were involved or on their regional extent. Further geochemical research will investigate whether the responsible processes correlate on a basin scale.

COMPARISON WITH LOCAL DIAGENETIC STUDIES

In the immediate area around the lead and zinc deposits, on petrographic grounds, MacLeod (1984) established that replacive micritic-microsparry, sucrosic and large rhombohedral dolomite affected the bank limestones (probably produced in conjunction with the deposition of Windsor evaporites), before mineralization. The latter occurred early in the diagenetic history of the carbonates and preceded meteoric late calcite and chemical compaction.

Schenk and Hatt (1984) presented a diagenetic sequence in which the Gays River Reef had undergone dolomitization and leaching under meteoric conditions. They considered the dolomitization event as an early one resulting from mixing of saline and meteoric waters; the late stage of calcite they regarded as a meteoric product. Akende and Zentilli (1984) found a similar sequence in which they added a syn-ore calcite that precipitated, as the latest coarse calcite, from ore-forming fluids.

In Table 1, the sequences of carbonate phases inferred for the mine area by the above researchers are compiled with the Musquodoboit and Shubenacadie regional sequence of this study. In the four sequences, the replacive dolomitization is an early phenomenon. However, there are disagreements on the number of calcitic phases and on the suggested origin for the dolomites and calcites.

As mentioned in the previous section, pre-ore dissolution provoked slight enhancement of porosity on a regional scale (about 2%). However, this solution stage does not correlate with the meteoric dissolution and leaching described by Schenk and Hatt (1984). They placed a dissolution event before marine cementation and another one after mineralization. The only dissolution event recognized in the present study followed marine cementation.

Micro-fractures and micro-breccias observed during this study occurred at the same relative chronological position as the intense brecciation observed at the deposit itself (Akende and Zentilli, 1984). Further investigations are required to decide whether the micro-fractures represent a regional manifestation of the intense brecciation.

GAYS RIVER AND OTHER MVT DEPOSITS — DISCUSSION

For six known MVT deposits, a compilation of the main carbonate diagenetic products and their timing relative to ore is shown in Table 2. The majority of these studies have been started recently and are ongoing. The compilation underlines the diversity and complexity of epigenetic processes associated with MVT deposits. Late coarse calcite is commonly post-ore whereas dolomite is either pre-ore or syn-ore. Because of the exclusive association of MVT deposits with dolomite, it is relevant to briefly review the suggested origin of dolomites.

For the Polaris and Viburnum Trend deposits, early dolomite is considered unrelated to ore forming fluids and, consequently, its origin is not discussed. At Olkusz,

Table 2. Diagenetic carbonate associated with six Mississippi Valley type deposits (normal marine cements are excluded from the sequences). Carbonate phases are described according to the author's terminology. For the Polaris deposit, the early replacive, replacive and pore filling sparry carbonate phases are dolomites. Calcite is also present in the diagenetic sequence and postdates the pore filling dolomite.

DEPOSITS	CARBONATE PHASES	RELATIVE TIMING	SUGGESTED ORIGIN	AUTHORS
Daniel's Harbour (Newfoundland)	Microdolomite (NL) Clustered dol. crystals (zoned) Replacive dolomite Replacive saddle dolomite Fault-related, replacive dolo.	Syn-sedimentary Syn-stylolitization Pre-ore Post-ore Syn and post-uplift	Not discussed " " " "	Lane, 1990
Olkusz (Poland)	Replacive dolo. Replacive ankerite Calcite cement	Pre-brecci. and ore " " Post-ore	Oxidant mineralizing solutions Reducing mineralizing solutions Oxidant mineralizing solutions	Sass-Gustkievicz 1990
Pine Point (NWT)	Fine replacive dolomite Pendant calcite Blocky calcite Replacive, medium dolomite Replacive coarse dolomite Pore filling saddle dolomite Coarse white calcite Late coarse calcite	Marine, very early Early Early Early burial Pre-ore Pre, syn and post-ore Post-ore, late " "	Evaporitic marine water Meteoric water Meteoric water Uncertain Burial, from basin brines "	Qing and Mountjoy, 1990
Polaris (Canadian Arctic)	Early replacive Replacive and pore-filling sparry	Pre-ore Post-ore	Not discussed Burial, controlled by carbon dioxide variation and sulfate reduction	Randell and Anderson, 1990
Tri State (Kansas, Missouri and Oklahoma)	Replac. coarse grey dolomite Saddle pink dol. cement Coarse calcite	Pre-ore " Post-ore	Brines (modified meteoric) " Cooling ore-fluids	Hagni, 1976
Viburnum Trend (Missouri)	Replacive dolomite (dull to NL) Dolomite cement (dull to lum.) Dolomite cement (dull to NL) Dolo. cement (zoned lum-NL) Saddle dolomite (dull to NL) Coarse calcite (lum. to zoned)	Very early Pre-ore Post-dissolution 1 syn-ore ", Syn-dissolution 2 ", post-dissolution Post-ore	Not discussed Basinal fluids " " " Not discussed	Voss, Hagni and Gregg, 1989

it is interpreted to be a product of mineralizing fluids which are presented as single-sourced diagenetic solutions undergoing continuous changes. At Pine Point and Tri-State, early dolomite is shown to be unrelated to ore forming fluids. For Pine Point, later coarse dolomites are shown to be related to ore-forming fluids, but, it seems that there is no ore-fluid-induced dolomite around the Tri-State district deposit. Likewise, according to MacLeod (1984), Schenk and Hatt (1984), and, Akende and Zentilli (1984), dolomitization around the Gays River deposit was totally unrelated to mineralization.

The succinct review implies that MVT dolomitization and mineralization do not have an obvious link. The peculiar diagenetic profile for the Tri-State and Gays River deposits perhaps suggests that these two deposits differ somewhat from the five other MVT deposits. Furthermore, their dolomite-ore mineral reference raises the fundamental question concerning the role of dolomitization in MVT deposit genesis, a problem that Anderson and Macqueen (1982) have already pointed out. More attention should be given to processes of dolomitization as well as to diagenetic differences between deposits as, eventually, it could play a role in the development of mineral deposit models.

SUMMARY

Preliminary regional study of the Gays River Formation established that the Musquodoboit and Shubenacadie sub-basins and the Mahone Bay and Antigonish areas were affected by distinct diagenetic successions. In the Musquodoboit-Shubenacadie area, early pervasive dolomitization affected bryozoan-algal banks of the Gays River Formation, without changes of the overall porosity. Marine cement, replacement dolomite, partial dissolution and pore-filling dolomite preceded mineralization which was followed by two generations of coarse calcite, zoned and uniform in cathodoluminescence. In the Mahone Bay and Antigonish areas, dolomite is minor, traces of galena occur in dolomitic rocks, but, sphalerite is unknown. Partial dissolution and leaching were the main diagenetic processes affecting the banks. Fracturing, brecciation and coarse calcite (zoned and uniform in cathodoluminescence) are diagenetic attributes that might correlate with those of the Shubenacadie-Musquodoboit area.

The diagenetic succession observed at the basin scale partly corresponds to what has been found around the deposit. Fracturing and brecciation could well be a regional manifestation of the intense brecciation that affected the carbonates at the mine itself.

Compilation of diagenetic studies of the deposit from Gays River and six other known areas raises the fundamental problem concerning the evaluation of the role of dolomitization in the genesis of MVT deposits. The connection between the Gays River deposit and the presence of dolomite has yet to be established.

For the second step of this project, the above petrographic results will guide the geochemical micro-sampling systematics. The aims of the research will be to evaluate the link between dolomitization and mineralization, to assess the stratigraphic and regional extent of the diagenetic events and to draw a paragenetic scenario into which the mineralizing event should be placed.

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Multiparameter surveys offshore Nova Scotia in 1988 and 1989

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Abstract

A four year program of gravity and magnetic surveys is underway on an opportunity basis on the Nova Scotia Continental Shelf. Preliminary analysis of discrepancies on track crossovers indicates an accuracy of better than 2 m in depth, 1.5 mGal in gravity and 20 nT in magnetic field. Modern instrumentation and high density of ship's tracks will produce a new set of maps suitable for interpretation using image processing and three dimensional inversion techniques.

Résumé

On procède actuellement, aux moments propices, à la réalisation d'un programme de levés gravimétriques et magnétiques d'une durée de quatre ans sur la plate-forme continentale Néo-Écossaise. L'analyse préliminiare d'écarts aux points de croisement de tracés indique une précision supérieure à 2 m pour la profondeur, à 1,5 mGal pour les mesures gravimétriques et à 20 nT pour le champ magnétique. Des instruments modernes et la densité des parcours de navires permettront de produire, à l'aide de méthodes de traitement d'images et d'inversion tridimensionnelle, un nouveau jeu de cartes se prêtant bien à l'interprétation.

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INTRODUCTION

Multiparameter Surveys, a joint project of the Geological Survey of Canada (GSC) and the Canadian Hydrographic Service (CHS) (Macnab, 1983a; Macnab et al., 1985; Verhoef et al., 1987), were resumed in 1988 after an interruption of several years. As in the past, the surveys in 1988 and 1989 were conducted on board CSS BAFFIN, length 86.9 m, displacement 4987 tonnes. For navigation control we used Loran-C in Rho-Rho mode (Wells and Grant, 1981) supplemented by GPS during it's 'visibility window'. The positional accuracy is estimated to be between 50 and 100 m.

Because of other survey commitments, CSS BAFFIN was available only late in the season: 26 October to 30 November 1988 and 16 October to 12 November 1989. In 1988 the weather conditions were not good for most of November and some survey time was lost while the ship was hove to in gale force winds. The weather was exceptionally good in 1989 but the survey had to be terminated prematurely because of labour difficulties.

The survey specifications called for east-west survey lines at 2 km line separation. North-south control lines were run at 8 km line spacing providing a 2 x 8 km grid of intersection for quality control. The bathymetry in the surveyed area (approximately 14 000 km^2) is shown in Figure 1.

EQUIPMENT INSTALLATION AND OPERATION

Operation of KSS-30 Seagravimeter

A seagravimeter (BSW KSS-30 Ser. No. 12) was installed in the gravity lab, aft of the galley on the Main deck. This is the standard installation position used for all Multidisciplinary Surveys since 1966.

In 1988, the seagravimeter operated with an internal filter setting at "Sea State 3" (roughly corresponding to a filter time constant of 3 minutes). We used the same setting for the first 8 days of the 1989 cruise (Days 289 to 297 inclusive) but changed the setting to "Sea State 2" (2 minutes time constant) for the remainder of the cruise. This change in filter setting was designed to provide data for investigation of instrumental delays.

In 1989, two KSS-30 seagravimeters operated side by side. The second instrument (Ser. No. 18), now owned and operated by Lamont-Doherty Geological Observatory (LDGO) of Palisides, N.Y., was compared with our instrument on a previous cruise in 1983 (when it was owned by PetroCanada Exploration and used on M.V. Bernier).

We operated the two KSS30 Seagravimeters so that LDGO could test the operational worthiness of their newly acquired instrument, while we wanted to re-evaluate the reliability of measurements with our instrument.

Comparison of two KSS-30 gravimeters

Ideally, the two gravimeters ought to be measuring and recording the same values and the difference between them should be a constant to account for the offset of the initial harbour settings. In practice this difference is not a constant for several reasons: The instruments are operating under slightly different accelerations as they are a couple of metres apart; the instruments may have different drift rates showing a slow change in the constant; the calibration factor of either (or both) instruments may not be correct, showing a change in difference with local absolute gravity (most pronounced when steaming north-south); dynamic characteristics of platforms may be different thus responding differently to large horizontal accelerations. The difference between the readings of the two seagravimeters over a period of 15 days is shown in Figure 2. This figure illustrates all of the above perturbations.

The record in Figure 2. starts on Day 300 when BAFFIN returned to the Bedford Institute of Oceanography for a 12 hour stopover alongside the jetty. The difference between the two meters is -250.2 mGal, constant within 0.05 mGal ('A' in Fig. 2.). When the ship departed the Institute (at 15 h 15) the difference changed by almost one mGal and only once approached the difference at the jetty. This was at point 'B', at the north end of one of the long N-S check lines, close to the latitude of the Bedford Institute. It would thus appear that there is a small error in a calibration factor, but this cannot be resolved from the differential drift which is evident from a slight but persistent positive overall slope of the difference curve over the 15 day period.

Sharp spikes in the graph (as for example at 'C') occur at sharp course alterations at the ends of survey lines. At these points there are large horizontal accelerations and the responses of the two platforms are slightly different, probably due to mechanical differences (friction in bearings, mass distribution, etc) and differences in position of sensors in the gravity lab. The interesting observation is that these spikes are between 2 and 4 mGal while on HUDSON in 1983 the differences on turns were between 10 and 12 mGal (Loncarevic, 1987). The gyro stabilized platforms respond in slightly different ways to the large horizontal accelerations experienced while the ship is altering course. We suspect that BAFFIN altered course more gradually and the horizontal accelerations were smaller than on HUDSON.

At 'D' and 'E', the LDGO Platform was deliberately set off level in roll by 720 arcseconds during the experiment to check the levelling of the instrument. At 'F' the pitch setting was adjusted by 480 arcseconds to check the levelling in that axes. The conclusion of these experiments was that under favorable sea conditions (very calm sea), the levelling of a gravimeter can be tested underway within 10-20 arcseconds by comparison with a second instrument.

At 'G', the filter setting of the LDGO gravimeter was changed from SeaState = 2 to 1, while the AGC instrument remained at 2. This produced an immediate jump of about 1.5 mGal in the difference while the long term response of the filter was settling down to the new level. While the difference showed a tendency (during Day308) to return to previous values, this experiment was not con-

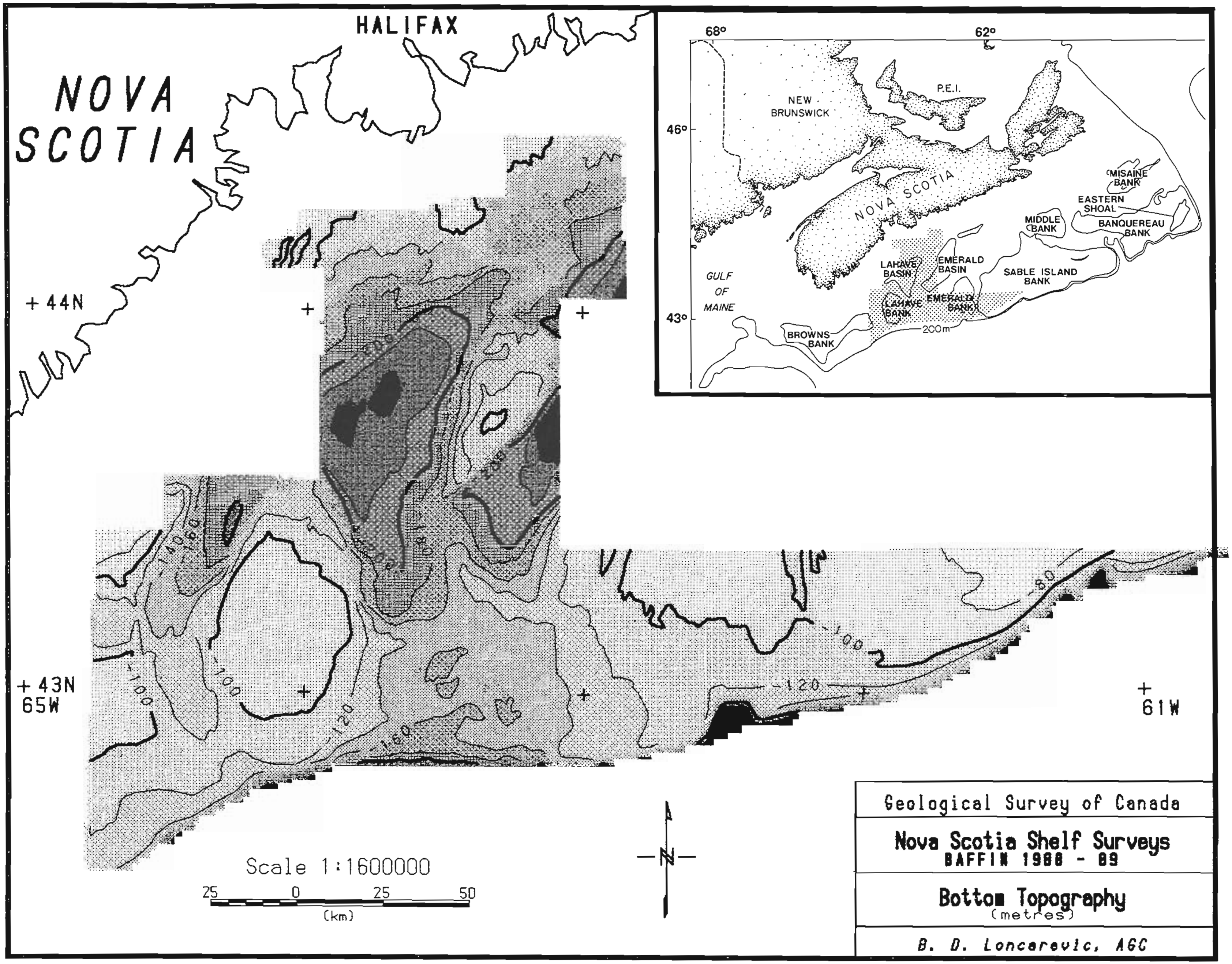


Figure 1. Bottom topography in the area of 1988 and 1989 multiparameter surveys. The survey area of approximately 14 000 km^2 is located directly south and east of Halifax as indicated in the inset.

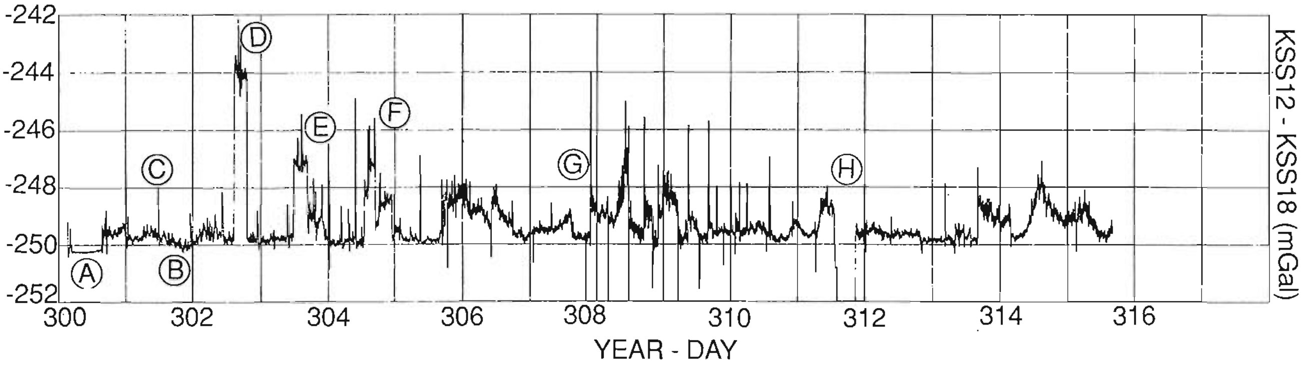


Figure 2. Difference in raw gravity readings of two seagravimeters installed side-by-side on board CSS BAFFIN over a 16 day survey period. Various features of this graph (A-H) are described in the text.

clusive due to a sudden deterioration in weather when the ship had to heave to for 12 hours. At 'H' the LDGO platform was shut down to test the second (spare) gyro.

We can derive two conclusions from the above observations. KSS30 gravimeters Ser. #12 and Ser. #18 measure the same quantity within about 1 mGal when on steady, straight course. New understanding of the dynamics of gravity measurements at sea may be derived from a careful study of the wealth of data collected on this cruise (and on cruise HUDSON 83-017) (Loncarevic, 1987).

Magnetometer

We used a 274m long cable with the towed sensor of the Varian V-75 Proton precession magnetometer. This cable length should be sufficient to eliminate the magnetic disturbance of the field due to the ship but we suspect that there may be a small residual effect not accounted for by Bullard and Mason (1961). It was estimated earlier (Ryall and Hall, 1979) with a shorter tow cable (182m) that the heading correction for BAFFIN is

$$- 6.6 \cos^2 H + 11 \cos H + 18 \text{ nT}$$

where H is the heading of the ship with respect to magnetic north. With a declination of 22°W in the working area, the amplitude of the heading correction on opposite survey tracks is 8.2nT. This should be less with the longer tow cable but the drop-off is gradual and it is possible that a correction of about 5nT remains. This correction has not been applied to the data. The sampling rate was 6 seconds representing about 30m distance over the ground at the nominal ship's speed of 10 knots. This is sufficient to resolve all the ground sources of magnetic anomalies along the track profile but the frequency spectrum of gridded data is limited by the across-track sampling (2 km).

DATA LOGGING

The CIGAL (Computer Integrated Geoscience Acquisition and Logging) System (Loncarevic and Coldwell, 1987) performed well while logging gravity and magnetic data. Every day, shortly after midnight (GMT), the day file was transferred to a shipboard uVAX computer where the data was merged with bathymetry.

SHIPBOARD DATA PROCESSING

The result of data processing on board ship using the SHIPAC system (Macnab, 1983b), are the Day Files (DFL) which are combined into a binary Cruise Master file. These files reside on the uVAX system. The BIGRID program (developed by Geosoft Inc. of Toronto) was used to produce grids with 500 x 500m cell size. These grids were transformed into ASCII files and downloaded to a PC using the Ethernet/RAF network.

On the PC, the ASCII files were converted into GEOSOFT Binary format and then converted again into GEOPAK Binary format for use with the RTI Imaging system (Real Time Imaging software by Aerodat/Geopak of Toronto). The procedure was cumbersome but it was sufficient to prove the usefulness of RTI for quality control.

RTI can produce both colour images of grid contours and/or "shaded relief" (shadowgram) maps for display on a flat screen monitor or for plotting on an HP PaintJet plotter. The shadowgrams emphasize linear features in the data and thus are a powerful tool for detecting systematic errors which depend on the direction of travel.

A common problem for all measurements on offshore surveys is the accuracy of positioning and timing of fixes. It is clear from this cruise that there are various delays in BIONAV which may not be accounted for in data processing. Any delay would put the actual ship position ahead of the fix. On opposite courses this error is doubled, producing a pronounced 'herring bone' pattern when surveying across horizontal gradients. Two problems were identified in navigation: there was a displacement of GPS and LORAN Antennae of about 37 m; and there was a delay in the LORAN receiver of about 20 seconds. Both of these errors are revealed by the inspection of shadowgrams.

In magnetic field data, the shadowgrams identified a serious problem with the diurnal variation. Some attempt was made to use the Bedford Institute of Oceanography Ground Base station to evaluate the corrections, although this approach has limitations and we are doubtful that a satisfactory solution can be found without mooring a buoy magnetometer in the centre of the survey area to monitor diurnal variation. We suspect that there might be a problem with the ships heading correction of the order of 5-10 nT.

Several problems were identified in the gravity data: delays due to internal filtering, tides (earth and sea) and ship's heading (platform tilt). Although the internal filtering problem may produce errors with a magnitude of the order of the previously mentioned differences between the values observed by the two gravity meters, the latter two errors are likely to be smaller. A more critical problem is that of satisfactorily monitoring the motion of the ship and computing an Eotvoes correction to an accuracy consistent with that of the gravity meter.

Inspection of shadowgrams suggests that bathymetry data could be improved by better corrections for the velocity of sound, heave compensation, tilt of the ram transducer and tides.

TRACK CROSSOVERS

The 1988-89 offshore multiparameter surveys were designed on a 2×8 km grid to provide a sufficient number of track crossovers to evaluate the internal consistency and accuracy of the surveys. The means and standard deviation of the absolute values of the crossover discrepancies are shown in Table 1.

We are not happy with the current implementation of the Eotvoes correction algorithm and intend to investigate possible improvements, for example, through better matching of the gravity and navigational filtering. A crude correction for diurnal magnetic variation was applied to the 1988 data but not to the 1989 survey, hence there is a considerable increase in the standard deviation of the latter. Moreover, the data collected during a three day magnetic disturbance in 1989 have not been critically examined and removed from the data set. We expect an improvement in crossovers after final data processing.

Table 1. Means and Standard Deviations of Crossover Differences for Bathymetry, Gravity and Magnetic Measurements

	Bathymetry (metres)	Gravity (mGal)	Magnetics (nanoTesla)
1988:			
Number of Observations	824	759	895
Mean	2.106	1.208	16.404
Standard Deviation	2.559	.990	13.379
1989:			
Number of Observations	411	368	374
Mean	1.314	2.690	21.337
Standard Deviation	2.889	2.425	23.229

Note: The values in the above Table are still considered preliminary and will be further adjusted following the completion of the Nova Scotia Shelf surveys in 1991.

Because of the high density of crossover points, it was possible to grid the data for each parameter on a 2×2 km grid. The results are shown in Figure 3. (a, b, and c). This type of presentation is useful for quality control as it draws attention to the problem areas.

The elongation of contours in all three parts of Figure 3 (L-arrows) indicates systematic errors along survey lines while alignments of high crossover values in the north-south direction (C-arrows) indicate bad control lines. Common areas of high crossovers for different parameters indicate a common cause, usually bad navigation.

The positional accuracy of crossovers is critically dependent on the local horizontal gradients. This is evident in the case of magnetic anomalies in the central area of the survey and in the case of gravity and bathymetry near the continental shelf edge.

BOTTOM TOPOGRAPHY

The topography of the sea bottom in the survey area is dominated by several banks and deep basins sculptured by the Pleistocene ice sheet (see inset in Figure 1 for locations of geographical features). In the southwest is the large and relatively flat LaHave Bank with the LaHave Basin to the north-northeast. This basin is interesting because of its triangular shape. The northwestern edge of the basin parallels the coastline trend while the eastern edge is oriented north-south. Emerald Basin to the east is characterized by a steep faulted western edge.

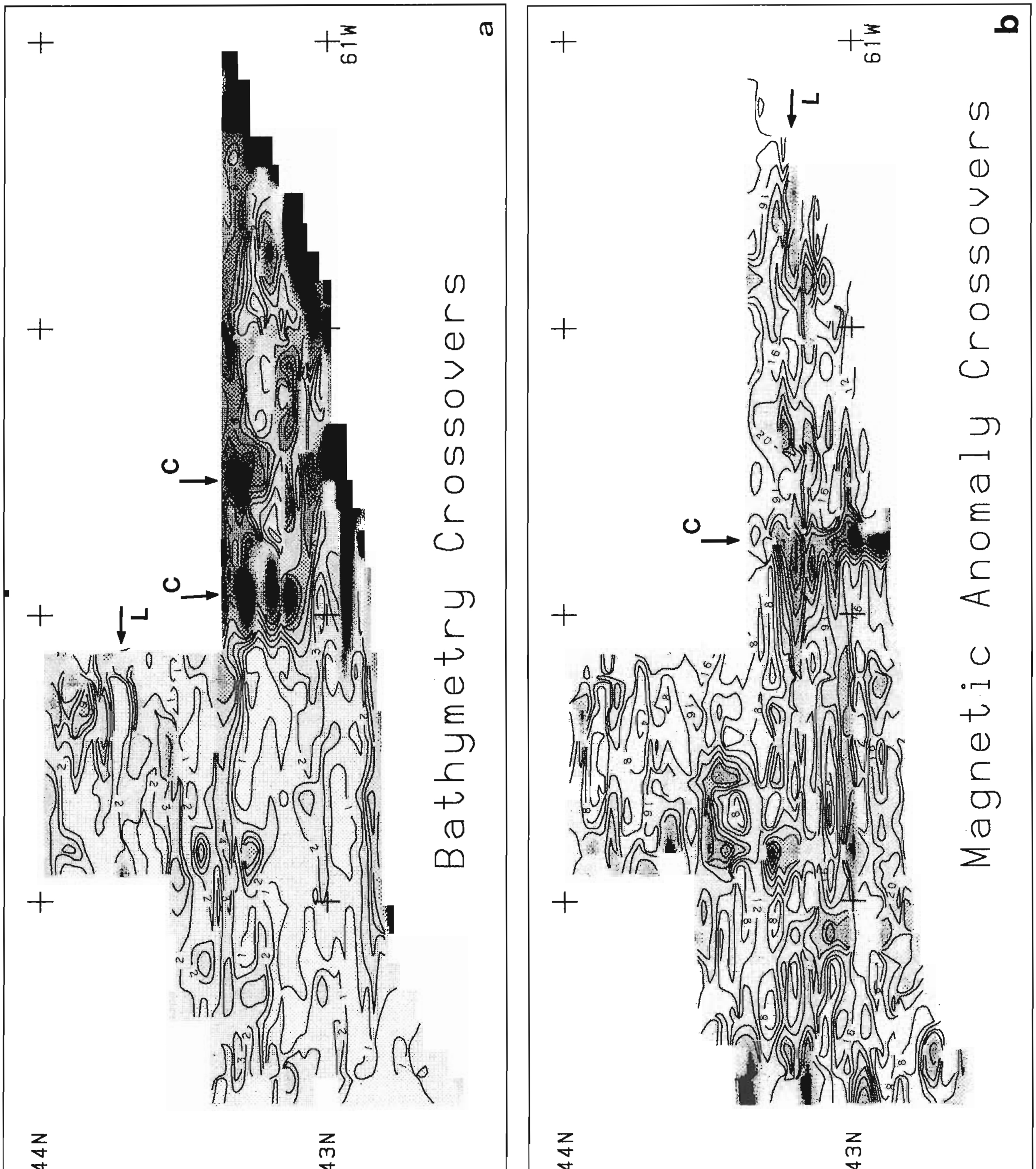
Between the LaHave and Emerald Banks is an area of sculptured topography representing ice lobes or terminal glacial moraines. Farther to the east is the Sable Island bank with a deeply incised drainage pattern.

Strong tidal currents are generated around the edges of the banks affecting the ship's speed and course. Passage of solitons has been documented in the area (Sandstrom and Elliot, 1984). These long period (5 - 10 minutes), small amplitude (10 - 20 cm) wave packets may be a cause of previously unsuspected errors in gravity measurements.

THE GRAVITY FIELD

Preliminary contouring of the free air anomalies is shown in Figure 4. The general trend of the contours and gravity features parallels the coastline and on-land trends in Nova Scotia. This is an important observation since the offshore extent of the Meguma terrane is not known. It appears from the gravity and magnetic (see below) data that Meguma basement may underlie at least part of the sedimentary basin.

The zero contour appears to coincide with the "Hinge Line", the inshore boundary of the Nova Scotia sedimentary basin. The anomalies between the coastline and the zero contour are negative with a large negative anomaly south-east of Halifax probably indicating a granitic intrusion. Farther offshore the anomalies become positive as the edge of the shelf is approached.


Bathymetry Crossovers
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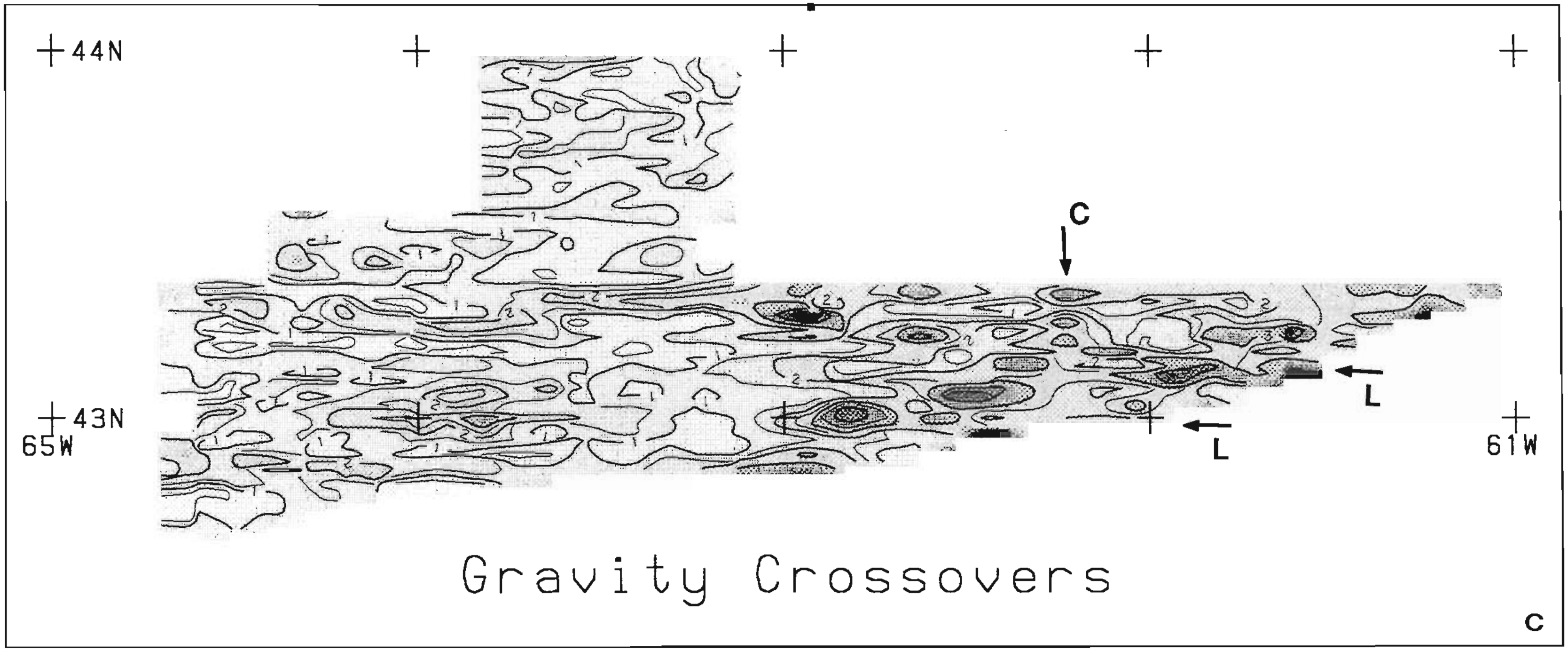

Figure 3. Contoured maps of crossover discrepancies based on a 2 × 2km grid. (a) Bathymetry crossovers contour interval 1m. Within most of the survey area the crossovers are within 2m. Note the high values of crossover discrepancies along the continental shelf edge between 61°W and 63°W where the depth increases rapidly. There are also high crossover discrepancies along the control lines indicated by the C-arrows. (b) Magnetic anomaly crossovers contour interval 4 nT. The high frequency content of this map is due to large horizontal gradients in the magnetic field. Control line indicated by C-arrow was recorded during a severe magnetic disturbance resulting in lager crossover discrepancies. (c) Gravity anomalycrossover contour interval 0.5 mGal. Discrepancies are generally within 2 mGal with the exception of some isolated values. Some of these 'bad' values will be edited out of the data after closer examination.

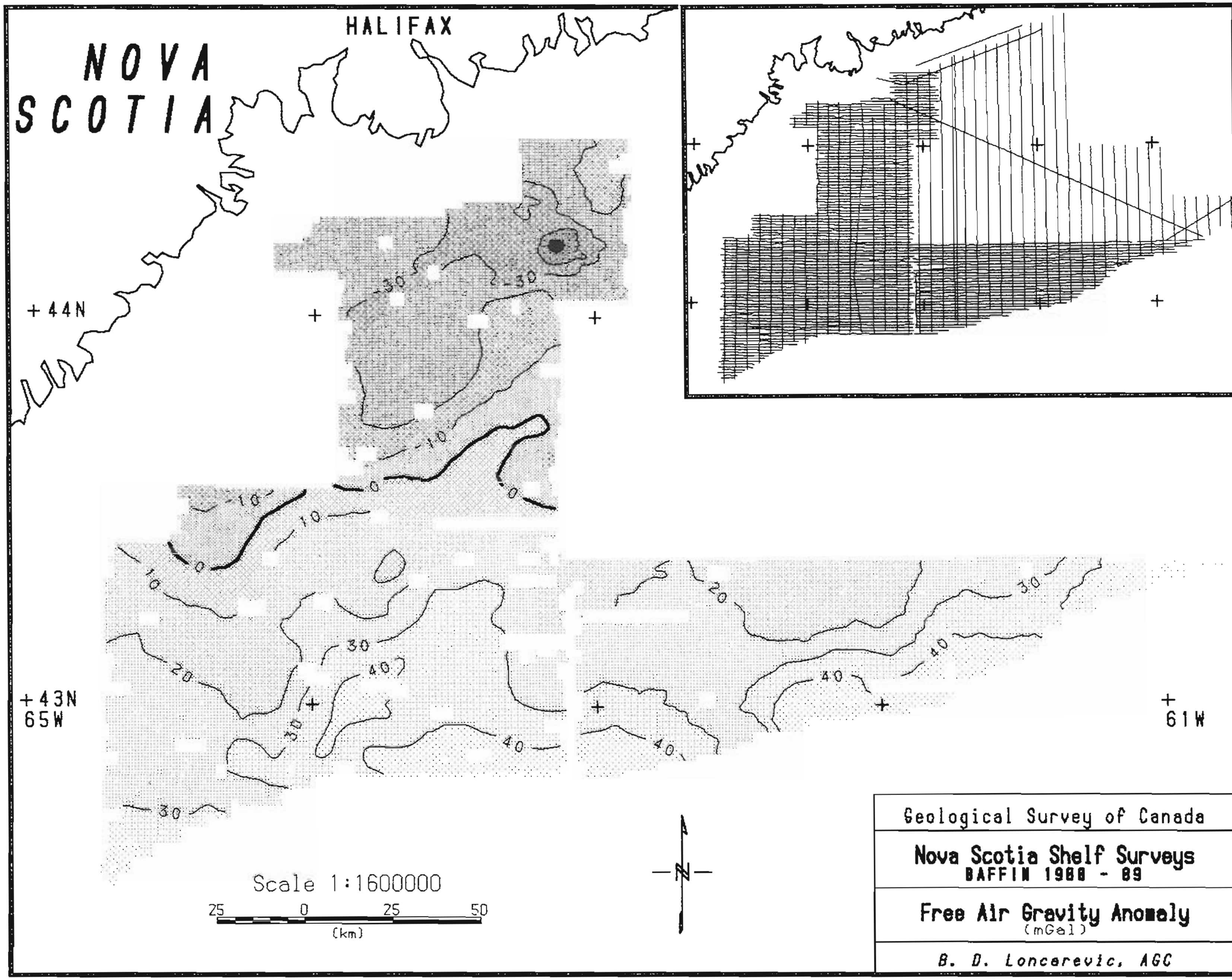

Figure 4. Free air gravity anomalies in the area of the 1988 and 1989 multiparameter surveys. Contour interval is 10 milliGals. Note the large negative anomaly southeast of Halifax. The positive anomalies near the southern edge of the survey are due to the continental shelf edge effect and are caused by the higher density of the oceanic crustal rocks at depth. The ships tracks along which the data was collected are shown in the inset.

THE MAGNETIC FIELD

Daily magnetic variation observations

In both survey years, a land-based station magnetometer was operated at the Bedford Institute of Oceanography non-magnetic hut since it was expected that the current solar cycle may produce substantial variations in the strength of the magnetic field (Hirman et al., 1989). The data were collected for two purposes: i) to identify magnetic storm days during which magnetic measurements at sea are useless; and ii) to correct sea measurements for diurnal magnetic variation. The magnetometer used was an old Barringer station magnetometer interfaced to a CORONA PC using a BCD board. The magnetometer could not be triggered from the computer and was free running at a 10 second polarizing rate. It was sampled by the PC every 30 seconds. Occasionally the PC readout occurred during the polarizing cycle and a bad reading was recorded.

A new program for data logging, MAGLOGST, was written in QuickBasic. The data are stored in dayfiles and a new file is created and opened automatically when the program is loaded by AUTOEXEC. The program uses default values of 30 second for sampling and 10 minute data blocks. The advantage of the new program is better data format and automatic resumption of data logging after a power break.

The mean daily values for the base station are compared to calculated IGRF85 values (Barraclough, 1987) for that location in Figure 5. We see that if the local anomaly value is adjusted to fit the calculated IGRF85 to the 1989 data (by subtracting −410 nT) we are then left with a 40 nT discrepancy in 1988. It is clear that the time-dependent extrapolation of IGRF85 is not very good. While the 1988 and 1989 (and presumably 1990) surveys could be made to fit by adding an arbitrary correction of about 40nT per year, this could only be considered a temporary measure.

When DGRF85 is calculated, sometime in 1990, all magnetic surveys between 1985 and 1990 should be reprocessed.

Magnetic anomalies

Preliminary contouring of the total field anomalies is shown in Figure 6. Although the range of anomalies is from −400 to 1600 nT, most of the field is within +/−200 nT. Negative anomalies occur in stripes paralleling the Nova Scotia coastline and the main trend of Meguma mineralization on land. Farther offshore the anomalies are subdued due to the increasing thickness of the Paleozoic sedimentary cover. The trends continue in the subparallel direction suggesting that the Meguma underlies at least part of the sedimentary basin.

The East Coast Magnetic Anomaly, a prominent feature bordering the North American continent, lies farther offshore and our survey only touched it between 62°W and 63°W. Although not very prominent, the Montagnais impact structure (Jansa and Pe-Piper, 1987) seems to be outlined near the southern edge of the survey, just west of 64°W.

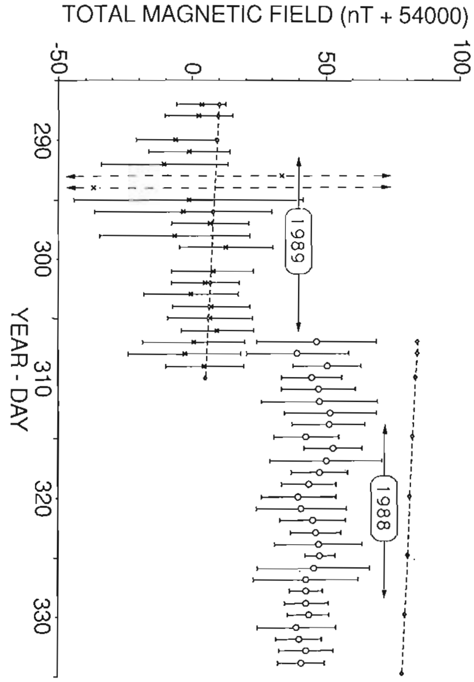

Figure 5. Daily means (circles) and standard deviations (bars) of station magnetometer readings. Note the highly disturbed field on Days 1989/293-295. Near horizontal dashed line represents IGRF85 prediction. Subtracting an arbitrary value (−410nT) makes the prediction fit the 1989 readings but leaves almost 40 nT discrepancy with 1988 readings.

In the northern area of the survey the magnetic sources are close to the surface and the spectrum of the magnetic field contains high frequencies. For this area, the 2km line spacing of our survey is not adequate. Within this area, directly south of Halifax, is an area of subdued magnetic anomalies which may represent a granitic pluton similar to the Peggy's Cove granites and other intrusions on land. The offshore granites were first identified by Hood (1967); we expect that our surveys will delineate their extent more precisely.

DISCUSSION

Acquisition of geophysical data during offshore hydrographic surveys is a cost effective way of conducting gravity and magnetic surveys.

Most of the multiparameter surveys conducted since 1970 were of reconnaissance nature with a track separation between 5 and 20 km. The present survey of the Nova Scotia shelf represents the return to the early standard (1964-1969) of dense coverage compatible with the instrumental resolution and detectability of anomalies. The data from high density surveys can be used with modern image processing and inversion techniques to gain new insight into the structure and geological history of inaccessible regions like the continental shelf.

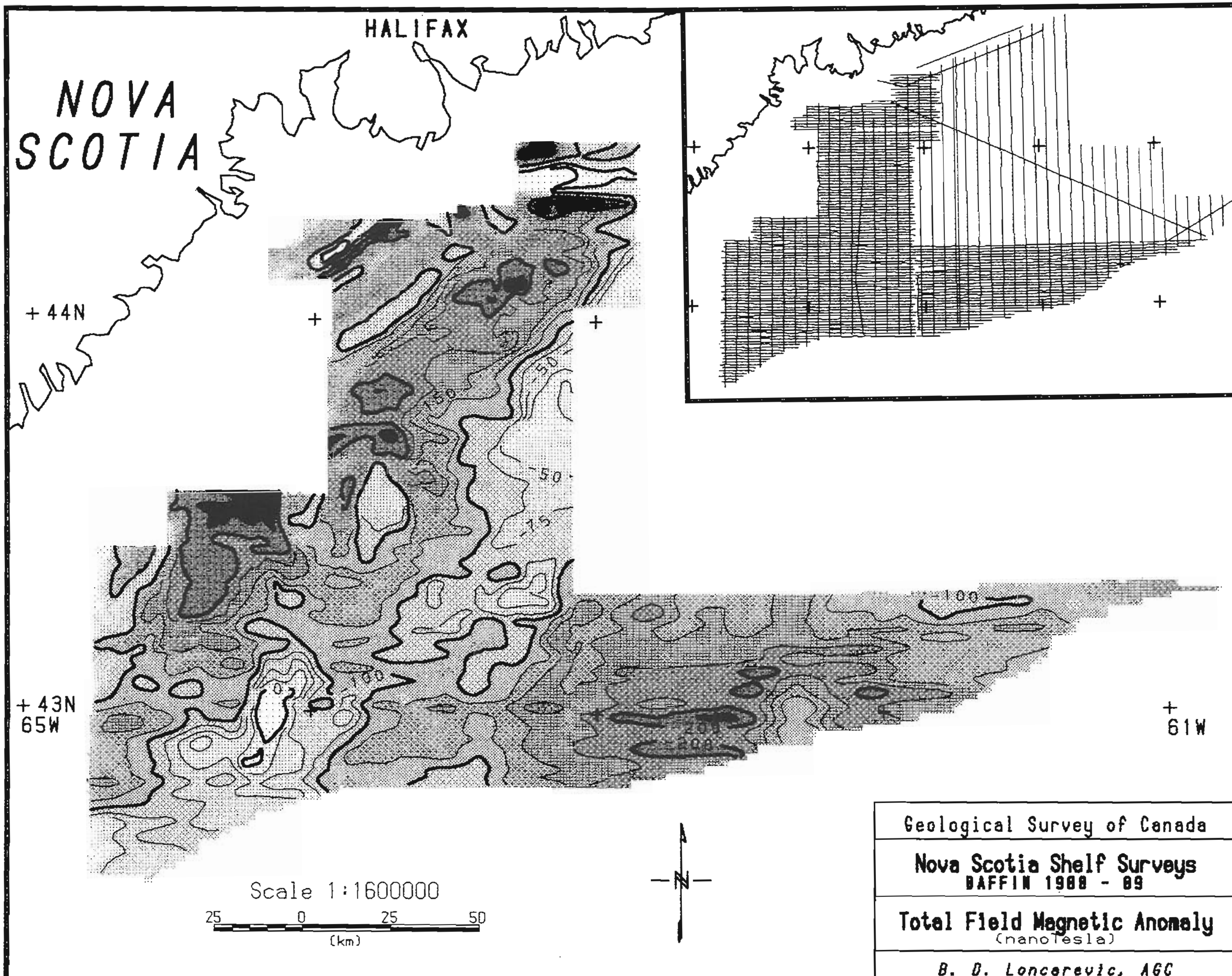

Figure 6. Total magnetic field anomalies after subtraction of the IGRF85. Note the high intensity anomalies in the north-west sector of the survey area where the basement is shallow. The general northeasterly trend of the anomalies parallels the Meguma trends on land.

ACKNOWLEDGMENTS

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Ground magnetic and rock magnetism studies near the Appalachian Dunnage-Gander terrane boundary, northern New Brunswick

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Abstract

Ground magnetic traverses have been surveyed across Ordovician rocks of the Elmtree inlier, the Nine Mile synform, and a fold-thrust package on the south side of the Rocky Brook-Millstream (RBM) fault. Magnetic models for all of these structures consistently indicate the presence of steeply dipping magnetic units extending to minimum depths of 1.5 to 2.5 km. Natural remanent magnetization (NRM) has been measured on oriented drill core samples with the aim of better constraining the magnetic models, which so far have been derived assuming all magnetizations to be induced by the earth's present magnetic field. NRM directions are generally steep, and point downwards. Magnetic fabric demagnetization has isolated a removable overprint in the fold-thrust package south of the Rocky Brook-Millstream fault. At another site, in the Elmtree inlier and much farther from the fault, no overprint was detected. Fabric analysis suggests that the last deformation in the area was dextral displacement along the fault.

Résumé

On a effectué des levés magnétiques terrestres en travers des roches ordoviciennes de la boutonnière d'Elmtree, de la synforme de Nine Mile et d'un ensemble de plis et de chevauchements sur le compartiment sud de la faille de Rocky Brook-Millstream. Les modèles magnétiques de toutes ces structures indiquent de façon cohérente la présence d'unités magnétiques à pendage abrupt s'étendant jusqu'à des profondeurs minimales de 1,5 à 2,5 km. On a mesuré la magnétisation rémanante naturelle (MRN) d'échantillons de carottes prélevés par sondage orienté de façon à restreindre à l'intérieur de paramètres mieux définis les modèles magnétiques, qui jusqu'alors avaient été établis par dérivation, en supposant que toutes les magnétisations sont induites par le champ magnétique actuel de la Terre. Les directions de la MRN sont en général abruptes et pointent vers le bas. La démagnétisation de la fabrique magnétique a permis d'isoler une surimpression amovible dans l'ensemble de plis et de chevauchements situé au sud de la faille de Rocky Brook-Millstream. À un autre endroit, dans la boutonnière d'Elmtree et à une distance beaucoup plus grande de la faille, on n'a pas détecté la présence d'une aucune surimpression. L'analyse de la fabrique révèle que la dernière déformation dans la région a été un déplacement dextre le long de la faille.

INTRODUCTION

In May and June 1990, detailed gravity and ground magnetic traverses were carried out near Bathurst, New Brunswick as a contribution to ongoing structural and tectonic investigations in the area (van Staal, 1986, 1987; van Staal et al., 1988). This report provides a preliminary description and interpretation of the magnetic data, including an appraisal of magnetic properties determined for oriented drill-core samples. The gravity data have been evaluated in Thomas et al. (1991).

Ground magnetic traverses were positioned to cross prominent aeromagnetic anomalies revealed on maps produced from data in the National Aeromagnetic Data Base, maintained by the Geological Survey of Canada. Generally, aeromagnetic data in the data base were recorded at a flight elevation of 305 m. Because the magnetometer is removed from the earth's surface, aeromagnetic surveys provide less detail of the magnetic field than do ground surveys. Consequently, a single anomaly in an aeromagnetic profile may be resolved into two or more anomalies in a ground magnetic profile. Geological units in the Bathurst area have widths in the order of a few hundred metres (van Staal and Langton, 1990). Accordingly, ground magnetic surveys having magnetometer stations spaced as close as 100 to 200 m were considered a viable means of resolving the magnetic signatures of such units, thereby allowing detailed models of the causative geological bodies to be calculated.

Five ground traverses were surveyed across four prominent positive anomalies with the objective of examining the third dimension of the following geological features:

1. The Ordovician Elmtree inlier.
2. A fold-thrust package of Ordovician sedimentary and volcanic rocks immediately south of the Rocky Brook-Millstream fault and northwest of Bathurst.
3. Ordovician volcanic units within the Nine Mile synform.

In addition to the ground surveys, magnetic investigations included in situ measurements of magnetic susceptibility, the collection of rock samples for laboratory measurements of susceptibility and rock density, and the drilling of oriented core samples for laboratory investigations of magnetic properties, including natural remanent magnetization.

GEOLOGICAL SETTING

A simplified geological map of the region based on Davies (1979), and modified by van Staal et al. (1990) and van Staal and Langton (1990) is shown in Figure 1. The region contains the boundary between the Dunnage and Gander tectono-stratigraphic zones of the Appalachian orogen as defined by Williams (1979), which, in New Brunswick, was thought to lie at or near the Rocky Brook-Millstream fault. The Dunnage zone lay to the north of the fault, and the Gander zone to the south. The Dunnage zone, in general terms, was thought to represent the vestiges of the Iapetus Ocean, e.g. island arc assemblages that developed on oceanic crust, whereas the Gander zone was generally interpreted as the eastern continental margin of the Iapetus Ocean, which locally may have been of Andean-type (Williams, 1979). Although initially referred to as zones, these tectono-stratigraphic units are now commonly termed terranes.

Williams and Hatcher (1983) noted that, in New Brunswick and southward, where volcanic rocks similar to those of the Dunnage terrane overlie clastics of the Gander terrane, the distinction between the two terranes can only be made at deeper stratigraphic levels. van Staal (1987) was of a similar opinion, concluding that, because "Gander" rocks represent basement to the "Dunnage" rocks, a tectono-stratigraphic boundary between the two terranes in this area has no real significance. More recently, van Staal et al. (1990) added support to this idea by demonstrating that the Ordovician Fournier Group is represented on both sides of the Rocky Brook-Millstream fault, thereby eliminating the fault as the Dunnage-Gander boundary. Notwithstanding the demise of the fault as the accepted boundary, for the purpose of describing the geology of the area, it provides a convenient line of subdivision. In the following descriptions, therefore, the terms Dunnage and Gander terranes equate with regions north of and south of the fault, respectively.

The Dunnage terrane

The Dunnage terrane contains the Ordovician Elmtree inlier, near the coast of Chaleur Bay, which is surrounded on most sides by Silurian sedimentary rocks of the Chaleur Group, although a large intrusion of Devonian gabbro and diabase flanks it to the west (Davies, 1979). Two groups of rocks were recognized in the inlier (van Staal and Langton, 1990). The Fournier Group is Middle Ordovician (Sullivan et al., 1990) and lies structurally above the older Tetagouche Group. The Tetagouche Group is represented by the sedimentary Elmtree Formation, formed largely of slate and shale. The Fournier Group comprises basalt, gabbro, diabase, minor plagiogranite, greywacke and slate, which are assigned to two formations. An association of pillow lavas, gabbros, amphibolites, trondhjemites, sheeted dykes, greywacke and slate, which defines the Deveraux Formation (van Staal and Langton, 1990), was interpreted as an ophiolitic sequence by Pajari et al. (1977). The Pointe Verte Formation includes sedimentary rocks and alkali basalt.

The Silurian Chaleur Group forms an east-northeast trending belt separating the Elmtree inlier and various Silurian and Devonian rocks to the north from Ordovician rocks of the Miramichi Highlands to the south. The Chaleur Group consists predominantly of sedimentary rock types, including conglomerate, sandstone, siltstone, argillite, greywacke and limestone. Locally it is overlain unconformably by Carboniferous sandstones, conglomerates, siltstones and shales. Davies (1979) showed the belt as being fault-bounded on both sides, the southern fault being part of the Rocky Brook-Millstream fault system.

The Elmtree inlier is intruded by the Devonian Antinouri Lake granite, and the Chaleur Group by the Devonian Nicholas Denys granite.

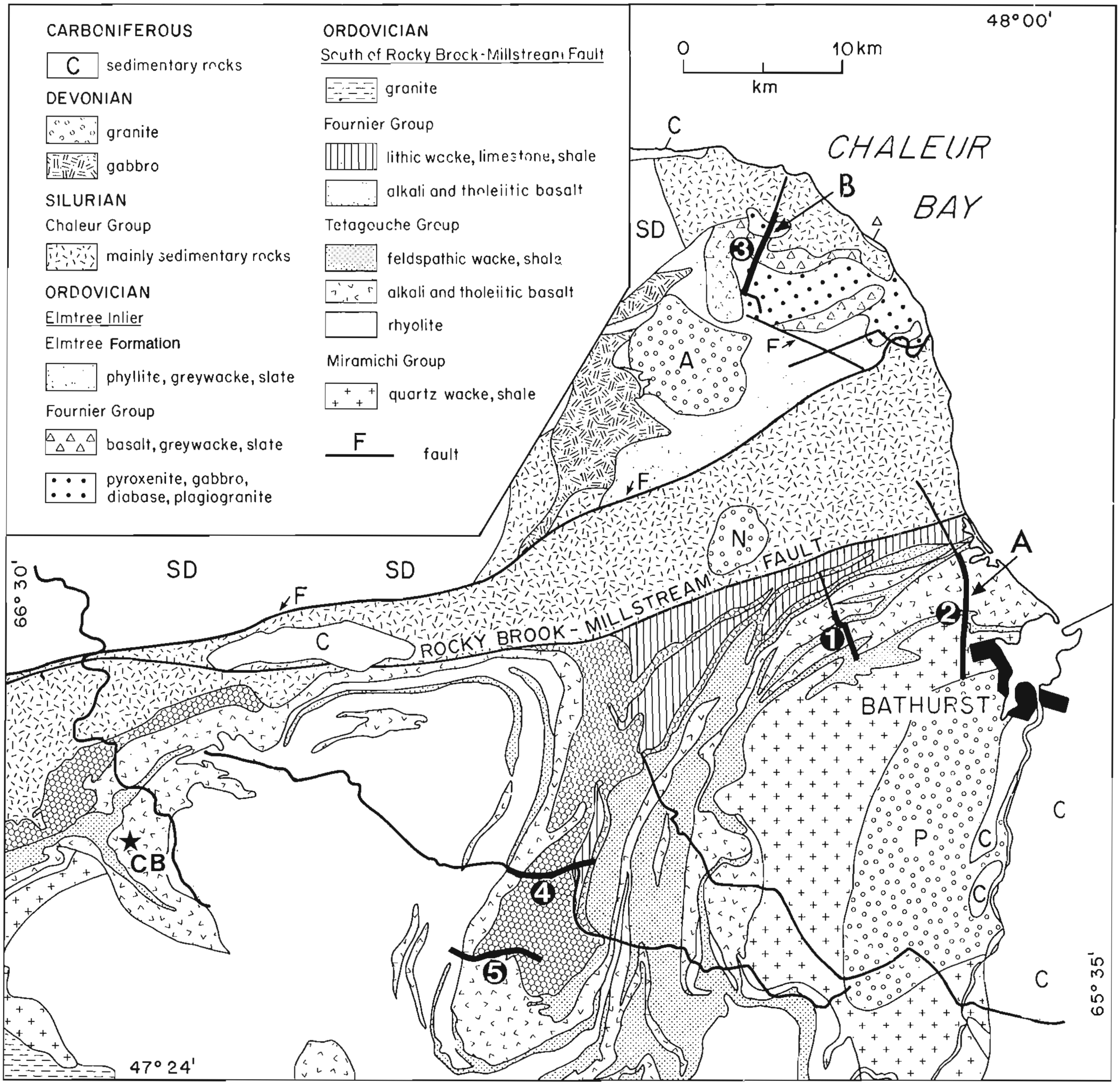

Figure 1. Geological map of region spanning Dunnage-Gander boundary zone near Bathurst, based on Davies (1979) and van Staal and Langton (1990). Devonian granites: A, Antinouri Lake; N, Nicholas Denys; P, Pabineau. SD, undivided Silurian and Devonian sedimentary and volcanic rocks. CB, Camel Back Mountain. Ground magnetic traverses are indicated by the heavy black lines labelled 1 through 5. Narrower solid lines are gravity traverses reported in Thomas et al. (1991). Drill core sites A and B are indicated.

The Gander terrane

In the Bathurst area, the Gander terrane is represented by Cambrian-Ordovician sedimentary rocks of the Miramichi Highlands. These rocks were previously known informally as the Tetagouche Group (e.g. van Staal, 1987; van Staal et al., 1988). A more recent geological analysis and compilation (van Staal and Langton, 1990) subdivides the former Tetagouche succession into three groups. The oldest is the Miramichi Group, which is Arenigian and older. The two younger groups, a redefined Tetagouche Group and the Fournier Group, have similar ages extending from Llanvirnian to Caradocian. The Miramichi Group, comprising quartz wacke and shale, is restricted to the southeast corner of the area where it flanks the Devonian Pabineau granite. It is succeeded to the immediate north and west by relatively narrow belts of sedimentary (feldspathic wacke and shale) and volcanic (tholeiitic and alkali basalt) assemblages of the Tetagouche Group. Farther west, in the central part of the region, lies the relatively broad Nine Mile synform. Its flanks comprise sedimentary and volcanic rocks of the Tetagouche Group, whereas the core is underlain by tholeiitic and alkali basalt and lithic wacke, limestone and shale of the Fournier Group. Westward again is the broad Tetagouche antiform built predominantly of rhyolite and associated epiclastic sedimentary rocks of the Tetagouche Group. The northwestern and western flanks of this antiform comprise mainly basaltic rocks of both the Tetagouche and Fournier groups.

Structurally the region is dominated by the Tetagouche antiform and the Nine Mile synform, which are the latest folds formed during the polyphase structural history. Much narrower folds occurring in the southeast of the region are the Mud Lake, Pabineau and Brunswick antiforms and the Pabineau and Brunswick synforms. The axes of these folds trend variously between north and northeast.

Many of the lithological units on van Staal and Langton's (1990) geological compilation are bounded by thrusts, which, for clarity, are not indicated on the simplified geological map of Figure 1. The down dip direction of the thrusts, notwithstanding various excursions around antiforms and synforms, is towards the north. In the area north of Bathurst, the structure of the Ordovician rocks had been viewed as a north-younging homocline, but recent mapping by van Staal et al. (1988) indicates that stratigraphy is repeated and inverted, characteristics attributed to early D_1-related, east or southeast directed thrusting and folding. This structural package is traceable westward to the region of Camel Back Mountain. Basalts and associated sediments of the upper part of the Tetagouche Group (old definition) are interpreted structurally to overlie silicic volcanics and sediments of the lower part, the contact being an F_1-related thrust (van Staal, 1987; van Staal et al., 1988).

Plate tectonic models

Rast and Stringer (1980) interpreted the Elmtree ophiolitic complex of the Dunnage terrane as remnant oceanic crust of a southeast-dipping subduction zone, and the rocks of the Tetagouche Group (old definition) in the Gander terrane as representing a complementary volcanic arc. Subduction was terminated in late Early Ordovician time through collision of the Iapetian mid-oceanic ridge with the Tetagouche arc.

Van Staal (1987) favoured a model in which both the Fournier and Tetagouche groups evolved in a marginal back-arc oceanic basin developed above a southeast-dipping subduction zone that closed the Iapetus Ocean to the northwest. Northwestward subduction leading to closure of the marginal basin was proposed to explain the occurrence of blue schist metamorphism in the Tetagouche rocks, and southeastward thrusting of these same rocks above silicic volcanics comprising the bulk of the lower part of the Tetagouche Group.

THE AEROMAGNETIC FIELD

The remainder of this paper deals with the earth's magnetic field and rock magnetism. The various parameters discussed are presented in SI (Systeme International) units. A list of parameters and their respective fundamental units as used herein is given below:

Parameter	Unit(s)
Magnetic field strength	nT (nanotesla), mT (millitesla)
Intensity of magnetization	A/m (ampere per metre)
Magnetic susceptibility	dimensionless

A simplified residual aeromagnetic total-field anomaly map is illustrated in Figure 2, in combination with a simplified geological map. The magnetic map is derived by subtracting the theoretical International Geomagnetic Reference Field from the measured total field. Aeromagnetic data were obtained at an elevation of approximately 305 m above ground surface along flight lines spaced about 805 m apart. Most anomaly values in the region fall in the range -250 to $+250$ nT, but values within some of the stronger positive anomalies are much more positive.

An extensive region of positive anomaly in the northwest of the area is associated with Silurian and Devonian sedimentary and volcanic rocks. Individual culminations, not apparent in the simplified magnetic map, trend parallel to the strike of the geological units; ENE in the western part, swinging to NNE in the north-central part of the region. Near the western margin of the area, this positive feature extends southward to embrace sedimentary rocks of the Silurian Chaleur Group and volcanic rocks of the Ordovician Fournier and Tetagouche groups in the region of Camel Back Mountain. The source of the positive response over the Silurian rocks is not apparent, but local positive anomalies over Ordovician terrane are closely associated with volcanic rocks.

The northern part of the Elmtree ophiolitic inlier correlates with a strong, grossly east-west trending positive anomaly, herein called the Elmtree High. It links with a more extensive positive anomaly offshore in Chaleur Bay, which may also signify the presence of ophiolitic rocks.

The Antinouri Lake granite correlates with a weak positive magnetic signature, which is not apparent in Figure 2. This contrasts with the much stronger anomaly associated with the Nicholas Denys granite, which intrudes Silurian sedimentary rocks. The anomaly extends WSW and ENE across the sedimentary rocks indicating subsurface extensions of the granite in the same directions. The southern flank of the anomaly coincides with the Rocky Brook-Millstream fault, suggesting that the fault marks the southern boundary of the granite. The Pabineau granite has a local moderate positive anomaly in the north, which continues as a weak ridge to the SSW. Other parts of its outcrop are characterized by a low, featureless field.

There is a noticeable difference between the characteristics of the magnetic fields over Ordovician rhyolitic rocks in the west and sedimentary and mafic volcanic rocks in the east, south of the Rocky Brook-Millstream fault. The field over rhyolitic rocks is uniformly low; the

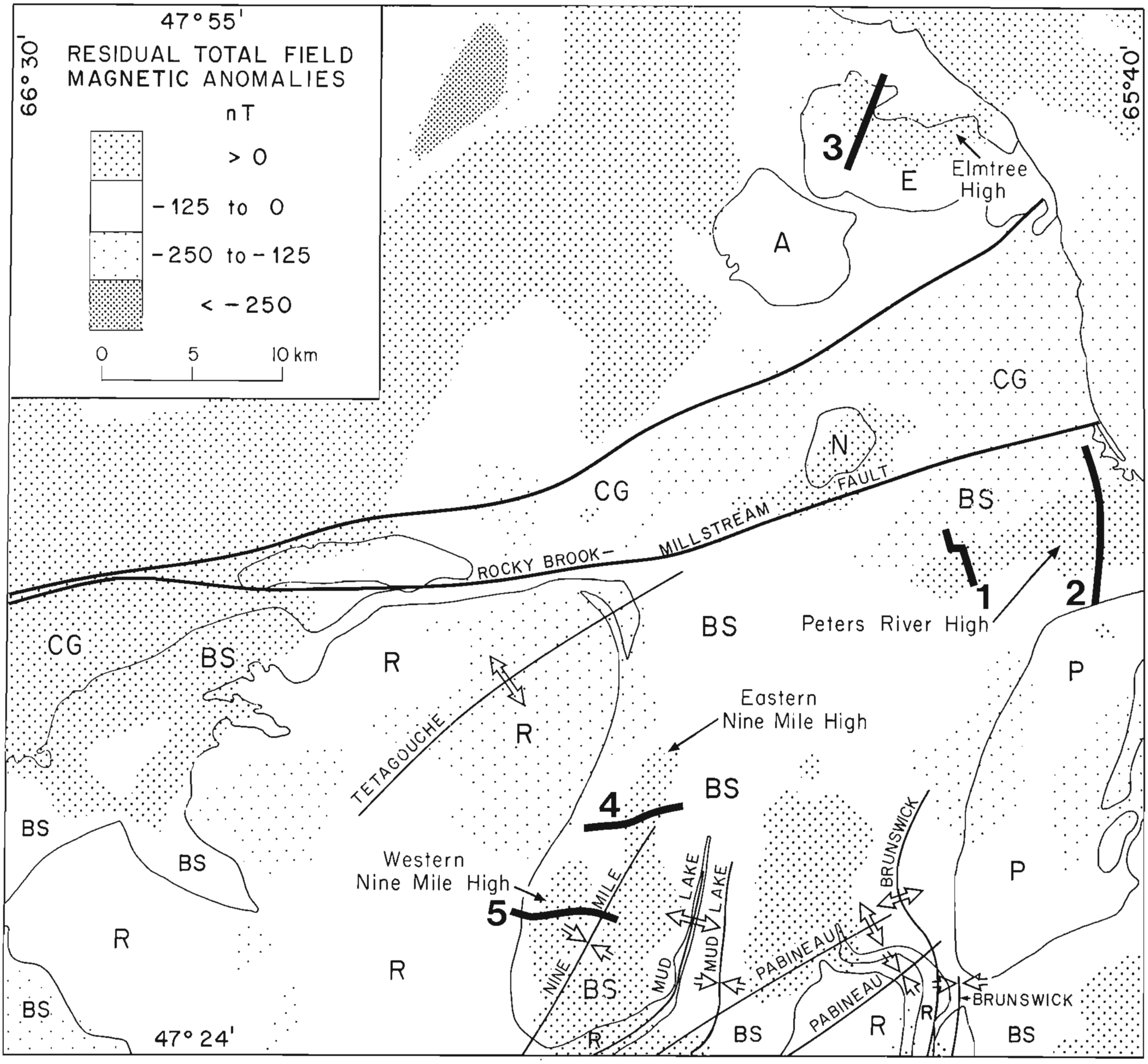

Figure 2. Geological map simplified from Figure 1 with addition of major antiforms and synforms (van Staal, 1986), and with a simplified residual total field magnetic anomaly map superimposed. Ordovician: BS, predominantly sedimentary and basaltic rocks of the Fournier and Tetagouche groups, and sedimentary rocks of the Miramichi Group; R, mainly rhyolitic and associated sedimentary rocks of the Tetagouche Group; E, Elmtree inlier. Silurian: CG, sedimentary rocks of the Chaleur Group. Devonian: A, Antinouri Lake granite; N, Nicholas Denys granite; P, Pabineau granite. Ground magnetic traverses 1 to 5 are indicated by heavy black lines.

axis of the Tetagouche antiform falls within a broad magnetic low. The field associated with sedimentary and mafic volcanic rocks is higher and contains several strong positive anomalies:

1. An ENE trending anomaly, the Peters River High, coinciding with a sedimentary-volcanic, fold-thrust package of Ordovician rocks immediately northwest of Bathurst.
2. A N to NNE trending anomaly overlying the Nine Mile synform. It diverges northward into two branches, each trending slightly oblique to the more northeasterly trend of the synformal axis. The branch anomalies are called the Western and Eastern Nine Mile highs. They coincide with alkali and tholeiitic basalts of the Fournier and Tetagouche groups.
3. A broad N to NNE trending anomaly, also resolvable into western and eastern branches, overlying mainly alkali and tholeiitic basalts of the Tetagouche Group occurring between the Mud Lake synform and Brunswick antiform.

GROUND MAGNETIC SURVEYS

Ground magnetic traverses were completed across the Elmtree, Peters River, and the Western and Eastern Nine Mile highs (Fig. 2). Measurements were made with an EDA OMNI IV proton precession total field/gradiometer system. The instrument has a processing sensitivity of ± 0.02 nT, but rounds values to 0.1 nT. Magnetometer stations were spaced at intervals of 100 or 200 m, subject to operational and geological requirements, with the closer spacing being employed across the main part of the aeromagnetic anomaly, and wider spacing to either side. The need for a base station to correct for diurnal magnetospheric variations was eliminated by the use of a technique called TIE-POINT traversing. This involves a preliminary survey which establishes, quickly, a series of base stations at large intervals along the traverse; the first and last observations in this survey are made at the first base station. The magnetometer's software corrects for diurnal drift and reference values are computed for all base stations. The subsequent detailed surveys tie into the base stations and a correction for diurnal drift is made on the assumption that it is linear. An assessment of diurnal activity was obtained by consulting records obtained at the Ottawa Observatory, which is the closest observatory of the Canadian Magnetic Observatory Network to the research area. These showed that diurnal activity was quiet during the periods of survey; no magnetic storm activity was apparent.

IN SITU MAGNETIC SUSCEPTIBILITY MEASUREMENTS

In situ magnetic susceptibility measurements, using a Geoinstruments Model JH-8 susceptibility meter, were made on outcrops along the traverses 2 through 5. No outcrops were encountered along traverse 1. The susceptibility data are summarized in Table 1. Along traverses 2, 4 and 5, south of the Rocky Brook-Millstream fault, most measurements were made on basaltic rocks of the Tetagouche and Fournier Groups. The mean susceptibilities of these rocks for the respective traverses are 2000, 700 and 70 SI $\times$ 10^{-5}. The highest susceptibility recorded, 15 000 SI $\times$ 10^{-5}, was for a basaltic outcrop along traverse 3; this is an extreme value, all others being less than 10 000 SI $\times$ 10^{-5}. By comparison, mean susceptibilities for gabbros sampled along the traverses are significantly lower, being 500 and 40 SI $\times$ 10^{-5} for traverses 2 and 5, respectively. A rock type, tentatively identified as mudstone, along traverse 5 yielded an unexpectedly high mean susceptibility of 1500 SI $\times$ 10^{-5}. In the Elmtree inlier, along traverse 3, it is noticeable that the mean susceptibility of the Belledune tholeiite, 2400 SI $\times$ 10^{-5}, is much higher than that of the Deveraux gabbro, which is 150 SI $\times$ 10^{-5}.

Table 1. Summary of In Situ Magnetic Susceptibilities
SI Units ($\times$ 10^{-5})

	Traverse 2		Traverse 3		Traverse 4		Traverse 5	
	Mean	Max.	Mean	Max.	Mean	Max.	Mean	Max.
Shale	30	45						
Tuff	2800	5800	30	40				
Basalt	2000	9200	2400	15000	700	4000	70	100
Mudstone							1500	5500
Gabbro/ Diabase	500	1000	150	800			40	50

Values are the mean susceptibility for the particular rock type along the traverse, and the maximum susceptibility recorded

Weathering can often result in magnetic susceptibilities which are lower than those of fresh rock at depth. The field values thus should be regarded as minimum values, but nevertheless provide a valuable guide for selecting susceptibilities for modelling studies.

LABORATORY STUDIES OF ROCK MAGNETISM

Magnetic properties of rocks at two drill core sites were investigated with the objectives of (1) determining the intensity and direction of any component of natural remanent magnetization that might be present, to better constrain models derived from the ground magnetic data, and (2) using magnetic fabrics as analytical indicators of tectonic motions and paleostress. One site, A, was located along traverse 2 crossing the Peters River High on Ordovician alkali basalts of the Tetagouche Group. The second site, B, was located on Belledune tholeiite of the Elmtree inlier along traverse 3.

Natural remanent magnetization

A knowledge of natural remanent magnetization (NRM) of a rock is important in modelling studies of anomalies. In the absence of such information, an interpreter may assume that an anomaly is produced entirely by induced magnetization related to the earth's ambient magnetic field. Such an assumption can lead to incorrect modelling of the sizes and shapes of geological bodies, particularly if an NRM component is very intense and oriented obliquely to the earth's magnetic field. The NRM determined at sites A and B are moderately intense, steeply dipping and directed downwards. The NRM data for the sites are summarized as follows:

	D	I	M	α_{95}	N
Site A	334°	69°	118.0	17.0	22
Site B	247°	69°	48.6	17.2	16

D = declination, I = inclination, M = intensity of magnetization in units of 10^{-2} A/m, α_{95} = radius of the circle of confidence, N = number of samples averaged.

Anisotropy of magnetic susceptibility (AMS) demagnetization

Magnetic fabric analysis by AMS demagnetization is a recently developed technique, first described by Park et al. (1988). The method uses stepwise heating and alternating field treatments for removing fabric overprints. The total magnetic fabric, defined by the susceptibility tensor, k_{ij}, results from various forces operating on the mineral assemblages at different times. Of interest here, with respect to the AMS demagnetization technique, is the fabric imprinted by weak stresses on the magnetic domain structures of individual grains. The metastable nature of this anisotropy component has been documented by magnetic induction (Kapicka, 1981), and its response to applied stress has been observed directly (Boyd and Fuller, 1981). Systematic changes in the total magnetic fabric that occur, both under alternating field and thermal demagnetization, indicate the removal of domain anisotropy. Because this magnetic property is highly stress-sensitive, it may record tectonic events which are weaker than those that produce visible kinematic indicators, such as deformed or reoriented grains.

AMS demagnetization experiments were carried out on 40 samples from sites A and B of this study. Half of the samples from each site were step-heated to 580°C. The other half were treated in steps of alternating field up to 290 mT. All experiments were performed inside magnetically shielded spaces that reduce the effects of the earth's field to less than 2 nT. Results indicate a fundamental difference between the two localities (Fig. 3). The average axis of maximum magnetic susceptibility, k_{max}, for site A, is directed 24°E with a dip of 20° prior to treatment. During treatment, the axis rotates counterclockwise to an azimuth of 355° and attains a shallower dip of 6°. Some samples attain zero dip, and a few continue in the reversed orientation, which then plots in the southeast quadrant (samples 174 and 176). Site B has less internal consistency in AMS orientations, as indicated by higher α_{95} values. The k_{max} axes cluster in the southeast quadrant, about azimuth 130° with a 45° dip. Treatment rarely produces significant axial shifts, suggesting that here deformations were more pervasive, imposing irreversible fabrics.

The AMS results demonstrate that the magnetic fabric at site A was reoriented from northwest to northeast by a local event, which did not affect site B. The best candidate for such a tectonic disturbance is dextral displacement along the Rocky Brook-Millstream fault. Removal of the overprinted component is incomplete, because only several samples from site A attain an endpoint in the southeast quadrant, in statistical agreement with site B. The difference in dip of k_{max} between the two sites could be related to vertical motion along the fault zone. This hypothesis remains to be confirmed by more extensive sampling.

MODELLING OF PROFILES

In this section of the paper, a series of profiles derived from the ground magnetic traverses are presented and modelled. The profiles were prepared simply by plotting successive station values and thus in plan view follow the course of the road, i.e., they are not projected onto a straight line. The geology is likewise plotted. For the preliminary modelling undertaken here this does not present a problem. Note, however, that the aeromagnetic profiles which are plotted for comparison follow straight lines that approximate the actual course of the traverses. Because the lengths of ground magnetic profiles are longer than corresponding aeromagnetic profiles, the latter were enlarged to fit the length of the former. Again, for the purposes of this initial examination, it does not pose a problem.

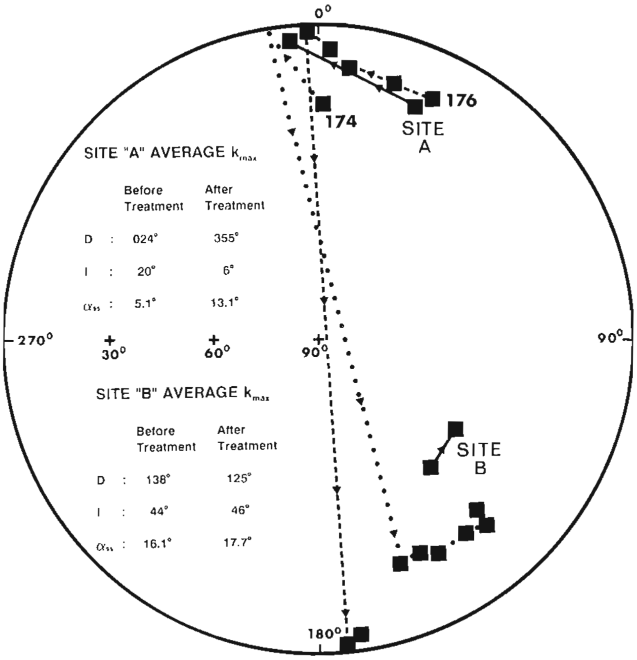

Figure 3. AMS demagnetization results for sites A and B showing shift in k_{max} with treatment. D = azimuth, I = dip, α_{95} = radius of circle of confidence.

Modelling was accomplished using a 2½-D interactive program, MAGRAV2 (Broome, 1989), installed on a desktop computer. Although this program has 2½-D capabilities, given the preliminary nature of the interpretations, an infinite strike length, i.e., a 2D geometry, was assumed for all bodies. Another assumption is that the magnetization is entirely induced, because information regarding NRM presented in the previous section was not available at the time these preliminary models were computed. Because the NRM is steeply inclined (69°), similar to the earth's field in the region (72°), the shapes of modelled bodies probably are not greatly different from those that would have been modelled had NRM been included in the computations. Many of the magnetic bodies in the models are steeply dipping and relatively narrow. The bodies have apparent maximum depths ranging generally between 1 and 2 km. The word "apparent" is used because, for narrow bodies, increasing the depth beyond a certain value, does not significantly change the magnetic expression. This is a limitation of the magnetic method. For this reason, and because the modelled bodies may extend deeper than shown, the bottoms of the bodies are shown as dashed lines. The geological legend for all profiles is shown in Figure 4.

Traverses across the Peters River aeromagnetic high

Traverses 1 and 2 cross the Peters River high, which coincides with an east-northeast trending belt of alkali basalt belonging to the Tetagouche Group. The corresponding profiles (Fig. 5, 6) exhibit a similar pattern of a relatively flat magnetic field over the northern part of the belt

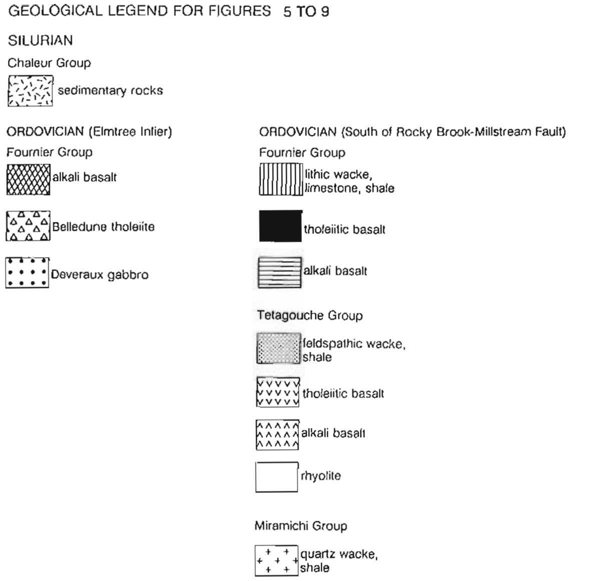

Figure 4. Geological legend for Figures 5 through 9.

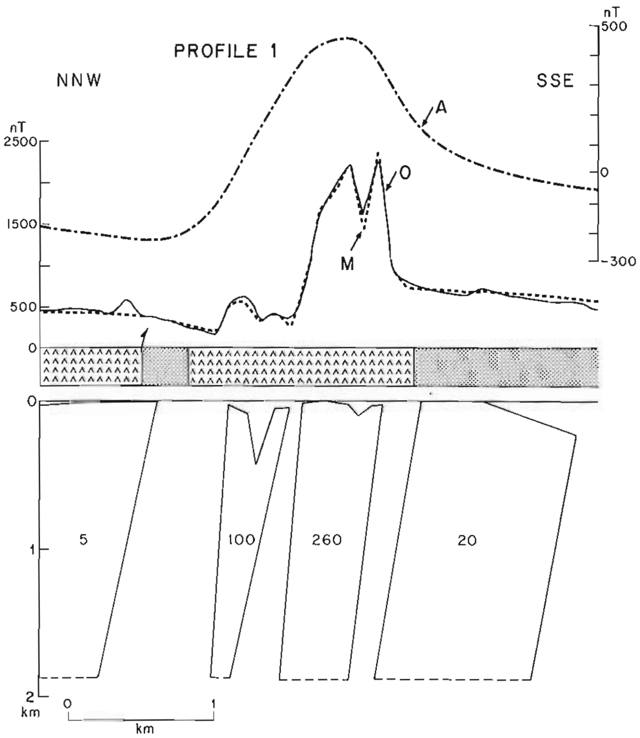

Figure 5. Magnetic profile and geology along traverse 1. Geology based on Davies (1979) and van Staal and Langton (1990). Contact with half-arrow head is a thrust fault; head is on the hanging wall. A, O and M are the aeromagnetic (scale is on right and is absolute), observed and model (scale is on left and is arbitrary) profiles, respectively. Numbers in the model bodies are magnetizations in units of 10^{-2}A/m. Geological legend given in Figure 4.

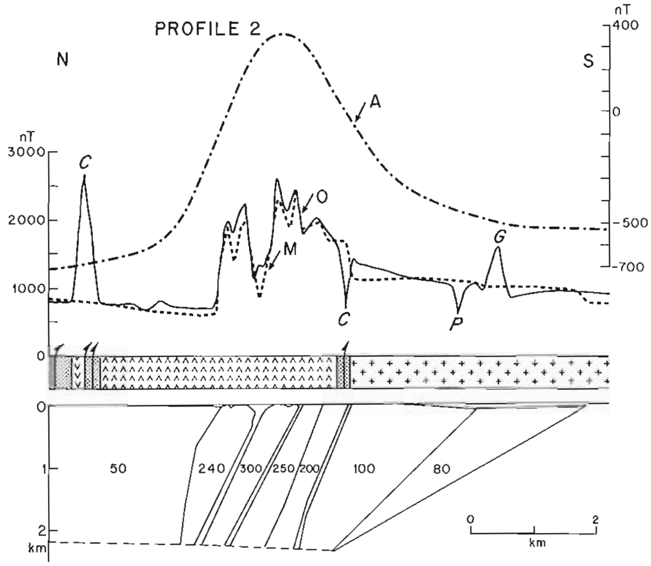

Figure 6. Magnetic profile and geology along traverse 2. Geology based on van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. A, O and M are the aeromagnetic (scale is on right and is absolute), observed and model (scale is on left and is arbitrary) profiles, respectively. Peaks on profile labelled C, G and P are cultural anomalies related to metal culverts, a metal guard rail and a power line, respectively. Numbers in the model bodies are magnetizations in units of 10^{-2}A/m. Geological legend given in Figure 4.

and strong positive anomalies with amplitudes approaching 2000 nT over the southern part. The alkali basalt is clearly separable into two gross units on the basis of its magnetic characteristics. The belt of alkali basalt forms most of the Beresford Domain, defined by van Staal et al. (1988). They found evidence for an anticlinal axis near the southern margin of the domain, north of which the structure was considered to be a north-younging homoclinal succession. The asymmetry of the magnetic field supports the homoclinal interpretation, but does not preclude the presence of an anticlinal structure within the magnetic portion of the basalt.

Magnetic models for both profiles comprise narrow magnetic units which dip very steeply towards the north. The units extend to minimum depths of roughly 2 km along both traverses.

Traverse across the Elmtree aeromagnetic high

Traverse 3 crosses the western part of the Elmtree inlier, which has an aeromagnetic signature in the form of two peaks (Fig. 7). The peaks are resolved by the ground traverses into a series of sharper peaks of shorter wavelength, ranging from about 300 to 1000 nT in amplitude. Most of the peaks over the Deveraux gabbro in the southwest are lower in amplitude than those over the Belledune tholeiite in the northwest. Modelling of profile 3 produces a picture of narrow, steeply dipping magnetic units within the Deveraux gabbro, separated by non-magnetic sections of the gabbro. Magnetic units in the Belledune tholeiite

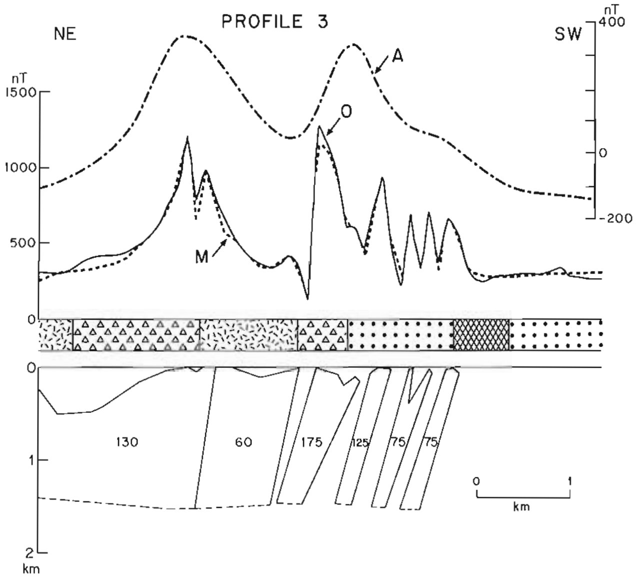

Figure 7. Magnetic profile and geology along traverse 3. Geology based on van Staal and Langton (1990). A, O and M are the aeromagnetic (scale is on the right and is absolute), observed and model (scale is on the left and is arbitrary) profiles, respectively. Numbers in the model bodies are magnetizations in units of 10^{-2}A/m. Geological legend given in Figure 4.

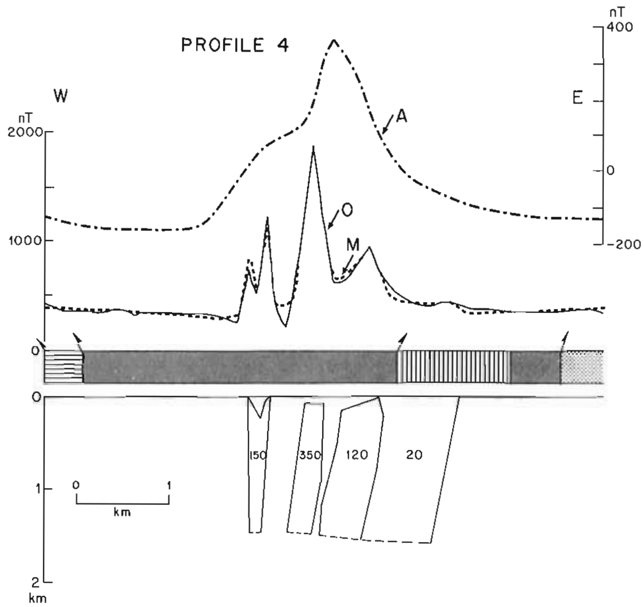

Figure 8. Magnetic profile and geology along traverse 4. Geology based on van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. A, O and M are the aeromagnetic (scale is on the right and is absolute), observed and model (scale is on the left and is arbitrary) profiles, respectively. Numbers in the model bodies are magnetizations in units of 10^{-2}A/m. Geological legend given in Figure 4.

are generally broader, but descend to around the same depth as those in the Deveraux gabbro, about 1.5 km. Analysis of gravity data along the same traverse (Thomas et al., 1991) indicated that the maximum depth of the inlier is about 400 m. Whether the 1.5 km depth indicated by the magnetic model is a minimum depth, for reasons outlined above, or a true depth may be resolved by further analysis. The aim of future research will be to produce gravity and magnetic models, which have similar depths. This will involve a more comprehensive study of rock properties data than has been possible in these preliminary reports.

Traverse across the eastern Nine Mile aeromagnetic high

Traverse 4 is located on the western limb of the Nine Mile synform, crossing mainly basalts and sedimentary rocks of the Fournier Group (Fig. 8). A single aeromagnetic peak is resolved by the ground survey into several narrower peaks having amplitudes ranging from about 400 to 1600 nT above background. These peaks occur over the eastern half of a belt of tholeiitic basalts, suggesting that a broad subdivision into two is feasible. The model indicates a series of narrow, steeply dipping magnetic units, separated by narrow, nonmagnetic sections of the basalt. The dip of the units is towards the west. The minimum depth of the units is about 1.5 km.

Traverse across the western Nine Mile aeromagnetic high

Traverse 5 also crosses the western limb of the Nine Mile synform, occupying a similar position with respect to the axis as does traverse 4 (Fig. 2). It crosses alkali and tholeii-

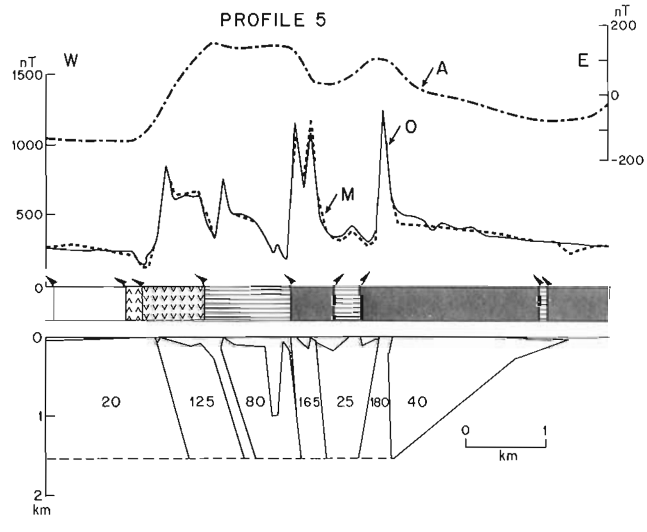

Figure 9. Magnetic profile and geology along traverse 5. Geology based on van Staal and Langton (1990). Contacts with half-arrow heads are thrust faults; heads are on the hanging walls. A, O and M are the aeromagnetic (scale is on the right and is absolute), observed and model (scale is on the left and is arbitrary) profiles, respectively. Numbers in the model bodies are magnetizations in units of 10^{-2}A/m. Geological legend given in Figure 4.

tic basalts belonging to both the Tetagouche and Fournier groups, and touches on rhyolites of the Tetagouche Group at its western end (Fig. 9). An aeromagnetic profile with two peaks is resolved into a series of short wavelength peaks with amplitudes ranging from about 500 to 900 nT by the ground surveys. These peaks fall mainly on basaltic rocks. The rhyolitic rocks produce little magnetic response. A large section of tholeiitic basalt within the Fournier Group at the eastern end of the profile also is magnetically very quiet. Magnetic units in the model are steeply inclined, dipping east, and are generally fairly narrow. Units in the central part of the model corresponding to alkali basalt of the Fournier Group may require modification, because the surface distribution of the basalt runs along the profile, rather than crossing it at a high angle, which is preferable for the modelling routine employed. The magnetic units extend to a minimum depth of about 1.5 km.

FUTURE PLANS

Magnetic rock property data will be further analyzed for input into magnetic modelling with the objective of producing refined models of the crust. Modelling will proceed in association with C. van Staal so that structural aspects of the region can be fully appreciated and integrated into the models.

ACKNOWLEDGMENTS

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The structure of the Half Mile Lake region, Bathurst Camp, New Brunswick[1]

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de Roo, J.A. and van Staal, C.R., The structure of the Half Mile Lake region, Bathurst Camp, New Brunswick; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 179-186, 1991.

Abstract

In the southwestern part of the Bathurst Camp in northern New Brunswick, complexly deformed volcanogenic sandstones and shales of the Patrick Brook Formation of the Lower/Middle Ordovician Tetagouche Group are host to at least two massive sulphide deposits in the Half Mile Lake region. The structural history of the Tetagouche Group in this region is similar to other parts of the Bathurst Camp in comprising five generations of mesoscale structures.

The regional structure is dominated by the isoclinal F_1 Half Mile Lake fold, which was refolded during D_2 into an open F_2 fold. S_2 was deformed into open to tight recumbent folds during F_3 folding. S_3 was subsequently cast into domes and basins in two open, upright folding events (F_4 and F_5). Correlation with structures at Heath Steele Mines shows that S_2, S_4 and S_5 have been rotated clockwise, suggesting large-scale open folding of all mesoscale structures during formation of the Nine Mile Synform.

Résumé

Dans la partie sud-ouest du camp de Bathurst dans le nord du Nouveau-Brunswick, des grès et des shales d'origine volcanique de la formation de Patrick Brook du groupe de Tetagouche de l'Ordovicien inférieur-moyen, déformés de façon complexe, contiennent au moins deux gisements de sulfures massifs dans la région du lac Half Mile. L'histoire structurale du groupe de Tetagouche dans cette région est semblable à celle d'autres parties du camp de Bathurst, comportant cinq générations de structures d'échelle mésoscopique.

La structure régionale est caractérisée par le pli du lac Half Mile F_1 isoclinal qui a été replissé au cours de D_2 en un pli F_2 ouvert. S_2 a été déformé en plis couchés d'ouverts à fermés au cours du plissement F_3. S_3 a par la suite été moulé en dômes et bassins au cours de deux événements de déformation en plis droits ouverts (F_4 et F_5). Une corrélation avec les structures observées dans les mines Heath Steele révèle que S_2, S_4 et S_5 ont subi une rotation horaire, indiquant un plissement ouvert à grande échelle de toutes les structures d'échelle mésoscopique au cours de la formation du synforme de Nine Mile.

[1] Contribution to the Canada-New Brunswick Mineral Development Agreement 1990-1995. Project carried by the Geological Survey of Canada.

INTRODUCTION

This paper presents preliminary results of structural and geological investigations in the Half Mile Lake region, in the southwestern part of the Bathurst Mining Camp of northern New Brunswick (Fig. 1).

The Bathurst Camp is mainly underlain by a sequence of mafic and felsic volcanic and sedimentary rocks of the Ordovician Tetagouche Group (Skinner, 1974), which can be traced throughout the Miramichi Highlands of central New Brunswick (Fyffe, 1982). The Tetagouche Group in the Bathurst Camp is host to numerous stratabound massive sulphide deposits, which are generally classified as volcanogenic (e.g. McAllister,1960; Harley,1979; and Franklin et al.,1981), although it is becoming increasingly clear that most deposits are hosted by sedimentary rather than volcanic rocks (McCutcheon, 1989; van Staal and Langton, 1990).

Despite recent U-Pb zircon dating of the volcanic rocks (e.g. Sullivan and van Staal, 1990), establishment of the chronostratigraphic relationships between the various formations in the Tetagouche Group is complicated by primary facies changes and intense deformation (van Staal, 1987; van Staal et al., 1990). Nevertheless a relatively simple tectonostratigraphy has been proposed recently (van Staal and Fyffe, in press; van Staal and Langton, 1990). Early deposition of quartz sandstones and shale of the Miramichi group (van Staal and Fyffe, in press) is inferred to have terminated in the Lower Ordovician, preceding a stage of silicic volcanism that was accompanied in the Middle Ordovician by mafic volcanism and was followed by renewed sedimentation (Helmstaedt, 1971; van Staal, 1987; van Staal and Fyffe, in press). The Tetagouche Group was intruded by granitoid plutons and gabbro, prior to the earliest recognisable regional, low-grade metamorphism and deformation

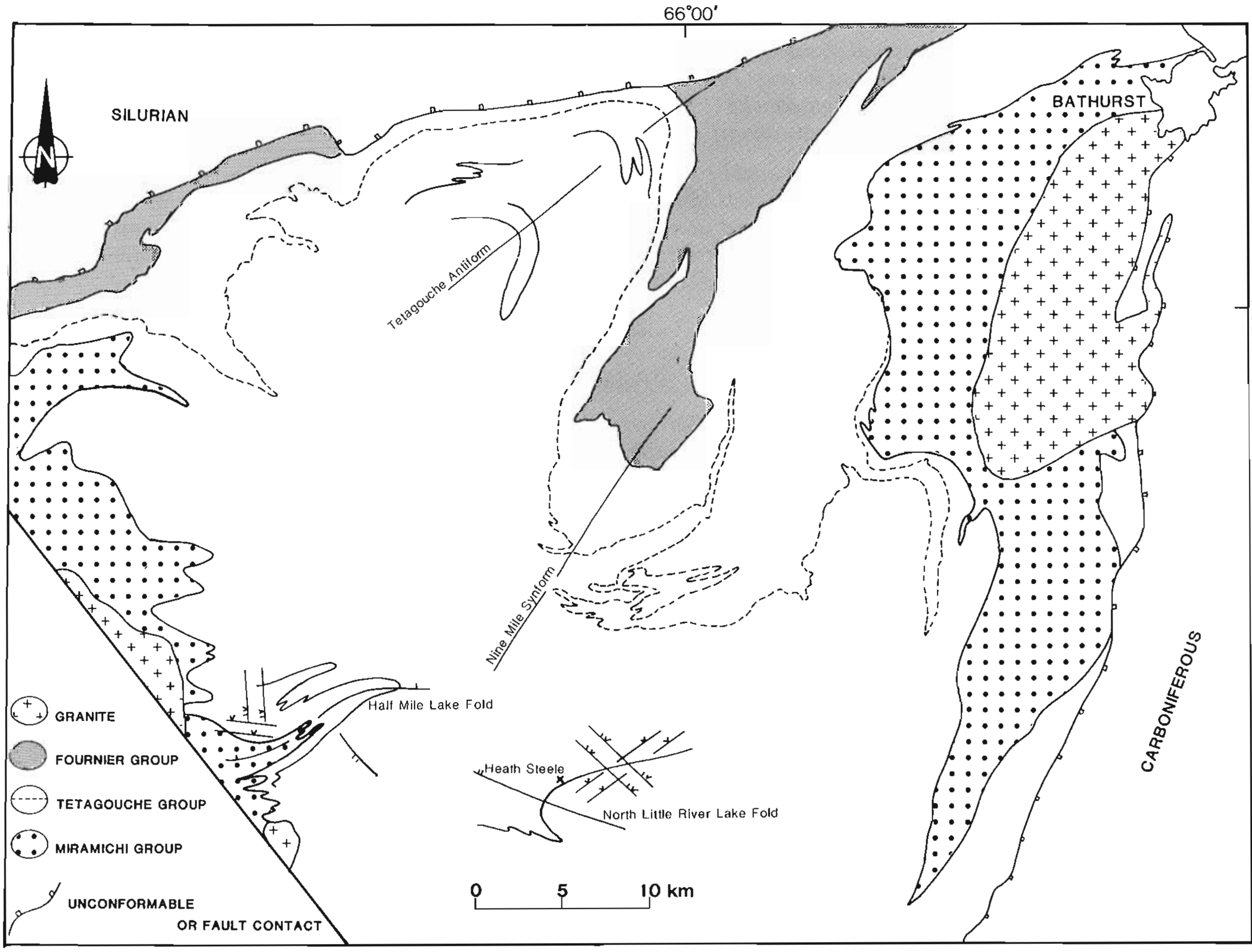

Figure 1. Simplified geological map of the Bathurst Camp in northern New Brunswick with the general distribution of the Miramichi, Tetagouche and Fournier groups (after Davies, 1979; van Staal et al., 1990) and major fold structures. The Tetagouche Group has been divided into two parts divided by the dashed line: the silicic volcanic and associated sedimentary rocks occupy the southern half and the main body of basaltic volcanic and associated sedimentary rocks the northern half. Note the rotation of the average orientations of S_2, S_4 and S_5 between the Half Mile lake and North Little River Lake folds. The relative age of the foliations and fold axial surfaces is indicated by Roman numerals.

(D_1; Helmstaedt, 1971; Skinner, 1974; Irrinki, 1986; van Staal, 1987). A younger suite of bimodal intrusions, mainly of late Silurian age (Bevier, 1988) intruded during post D_1 deformation (van Staal, 1987) and is generally referred to as the Acadian plutonic suite by Fyffe et al. (1981).

LITHOLOGIES OF THE HALF MILE LAKE REGION

Detailed mapping in the Half Mile Lake region (Fig. 1) is hindered by poor exposure and lithological changes induced by penetrative, locally intense, deformation. Lithological contacts are rarely exposed, and in many cases may be tectonic. Therefore the structural history of this area can only be understood when stratigraphic study is combined with detailed structural analysis.

The lithostratigraphy used in this paper (Fig. 2) is based on van Staal and Fyffe (in press) and van Staal & Langton (1990).

The oldest rocks exposed in the area belong to the Miramichi Group and consist of grey to black shaly and silty phyllites interbedded with light grey to greenish grey quartz arenites or wackes. The Miramichi Group rocks can be seen to be disconformably overlain by the rocks of the Tetagouche Group close to Little Falls in the Tetagouche River (van Staal and Fyffe, in press), whereas the contact between these two groups appears to be gradational in the Nepisiquit river (van Staal, unpubl. results). The nature of this contact is unknown in the Half Mile Lake area since it is not exposed.

The sedimentary rocks of the Tetagouche Group in the Half Mile Lake region consist mainly of shaly and sandy epiclastic sedimentary rocks. The sandstone beds are generally massive and tuffaceous with black quartz grains, which are interpreted as reworked phenocrysts. These sediments resemble rocks of both the Nepisiquit Falls and Patrick Brook formations defined in the northern part of the Bathurst camp. The predominance of dark grey to black shales, which are absent in the Nepisiguit Falls Formation indicates that these rocks are best included in the Patrick Brook Formation.

The sedimentary rocks assigned to the Patrick Brook Formation are host to at least two massive sulphide deposits (Half Mile Lake north and Half Mile Lake south; Davies, 1979), which are exposed in the nose area of the F_1 Half Mile Lake fold (Fig.2), near the contact with the metaporphyries. These contacts are also marked by local sulphide disseminations and iron enrichments (gossans), and the sulphide stratigraphy (i.e. a Cu-rich zone, overlain by a Pb/Zn-rich zone) and grading in the sediments suggests that the Half Mile Lake south orebody youngs towards the southeast (Harley, 1979; L. Davies, pers. comm.). The rocks of the Patrick Brook Formation are stratigraphically overlain by volcanic and minor sedimentary rocks of the Flat Landing Brook Formation. The volcanic rocks of this formation comprise mainly aphyric or feldsparphyric rhyolites with minor pyroclastic rocks, mafic volcanic rocks metamorphosed into greenschists and massive quartz and feldsparphyric metarhyolite bodies. The last are generally referred to as porphyries and are partly deformed into augen schists. The outcrop pattern of the large porphyry bodies in this area (Fig. 2) suggests that they locally crosscut the stratigraphy at a low angle and hence are thought to represent large sills. Elsewhere, the porphyries also, at least in part, represent high level intrusions (van Staal, unpubl. results). The mafic volcanic rocks of the Flat Landing brook Formation belong to the Otter Brook basalt suite (Winchester and van Staal, in review) and comprise lapilli tuffs and agglomerates or conglomerates, the fragments of which include some felsic volcanic rocks, and massive basalt flows and/or diabases. Pillow structures have not been recognised. This indicates that these rocks may have been deposited subaerially. The basalts are closely associated with greyish green chlorite-rich phyllites, which at least in part probably represent the metamorphosed equivalents of shaly sediments. The Tetagouche Group has been intruded by locally foliated gabbros that in part grade into metadiabase, and by granites. The granites (Fig.2) appear undeformed and unaltered, and are considered to form part of the Siluro-Devonian Miramichi Granite and late Silurian Mount Elizabeth granite (Whalen, 1990). The plutons have a contact aureole, in which most of the metasedimentary phyllites were transformed into hornfels with porphyroblasts of cordierite and andalusite.

DEFORMATION OF THE HALF MILE LAKE REGION

The deformational history of the Tetagouche Group in the Bathurst camp is polyphase (van Staal and Williams, 1984; van Staal; 1987; de Roo et al., 1990). In general terms, key outcrops are sought, (Fig. 3) which allow establishment of all or most generations of structures on the basis of overprinting relationships on mesoscopic scale. Sets of structures are then correlated between outcrops using a combination of criteria, including style, orientation, and/or age, relative to other structures present. This procedure resulted in the identification of five generations of structures (F_1 - F_5), which is identical to the succession of generations found in all other parts of the Bathurst Camp.

Small-scale structures

The earliest deformation (D_1) is represented by a very penetrative cleavage or schistosity (S_1; Fig. 2). In pelites, S_1 is commonly represented by a differentiated layering, and is parallel to bedding (S_0). At Heath Steele Mines (Fig. 1), adjoining Half Mile Lake to the east, it has been demonstrated that this parallelism is at least in part the result of rotation of S_0 into S_1 by F_1 transposition folding (de Roo et al. 1990). S_1 is overprinted by F_2 folds with an S_2 axial plane cleavage. S_2 is locally a differentiated crenulation cleavage. The presence of rootless F_2 folds within the layering is indication of local F_2 transposition of S_1 into S_2. The variation in the intensity of F_2 folding and S_2 cleavage development suggests a heterogeneous D_2 strain distribution, which explains why the dominant foliation in outcrops throughout the area can be S_1 or S_2. F_2 transposition also explains why

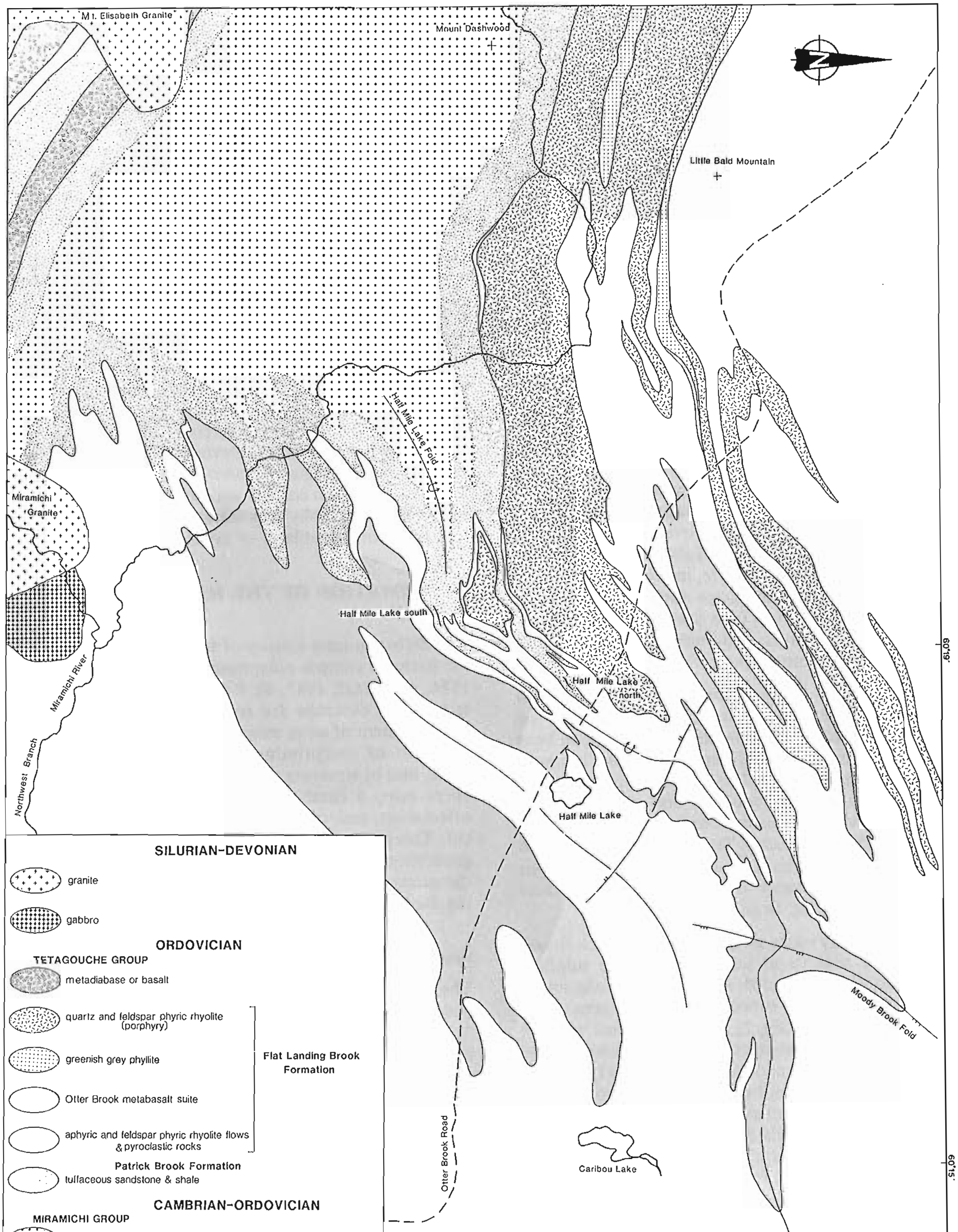

Figure 2. Geological map of the Half Mile Lake region. The geological map is in part based on a compilation of data extracted from 3 assessment reports filed at the Department of Natural Resources of New Brunswick. These files are: 1) Adair, R.N., 1987, Noranda Exploration — file 473442; 2) Goodwin, J.A., 1987, Billiton Canada — file 472953; 3) Mattagami Lake Mines — file 471609.

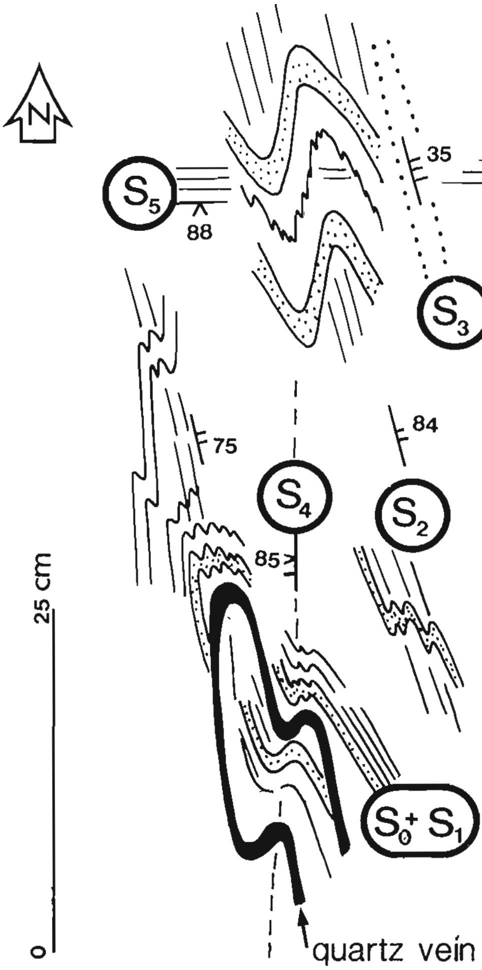

Figure 3. Sketch of a key outcrop with the presence of structures of all generations, as determined from overprinting relationships. Note that S_2 and S_3 have approximately the same strike.

the extension lineation (L1) locally lies in S_2, since elsewhere L_1 is demonstrably folded or deformed by F_2 (van Staal, 1985). This lineation is defined by a preferred orientation of minerals, microboudinaged porphyroclasts of feldspar and quartz, and by stretched lapilli and agglomerate fragments. L_1 plunges gently north in domains where later deformation was weak.

S_2 has been deformed into F_3 recumbent or reclined folds (Figs. 4a, 4b). The F_3 folds are open, locally chevron-style, structures or small crenulations. Their axial planes are generally marked by a horizontal, or gently dipping, crenulation cleavage (S_3). Locally, S_3 develops into a differentiated layering in the axial plane of very tight F_3 folds. The earlier foliations are discontinuous across such folds. Hence, in terms of style, D_3 locally mimics D_2.

Like S_2, S_3 has a variable dip. This is partly the product of upright F_4 folding. F_4 is represented in outcrop by a northerly striking crenulation cleavage (S_4) in the axial plane of open, upright, folds of S_3. S_3 and S_4 are much better developed in the Half Mile Lake area than at Heath Steele Mines, where overprinting relationships were

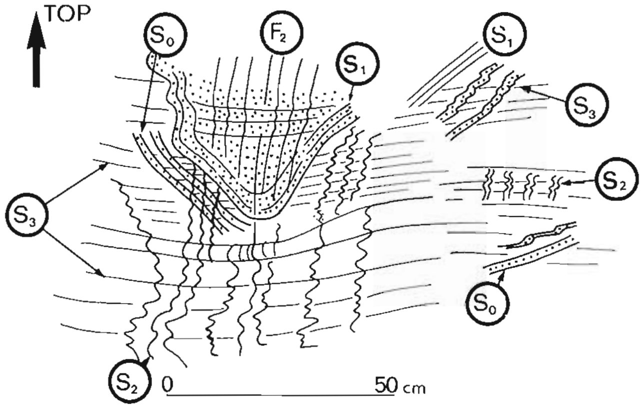

Figure 4a. Sketch showing overprinting relationships between the first three generation structures in an almost vertical exposure of tuffaceous sandstone and shale near the Half Mile Lake North deposit. S_2 is locally developed as a differentiated crenulation cleavage.

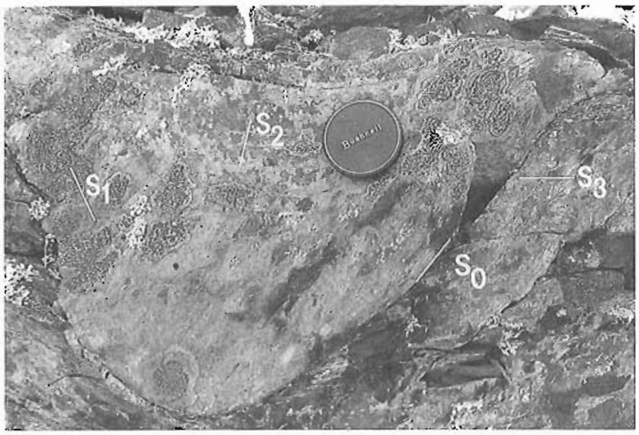

Figure 4b. Close up of F_2 fold in sandstone of fig. 4a. Note that the S_3 crenulation cleavage is weakly developed in the competent sandstone compared to the incompetent shale. In fact this structure shows a small scale example of the incipient development of 'steep and flat' belts: i.e. where S_3 is strong, S_2 is almost transposed into the shallowly dipping S_3 orientation, whereas S_2 remains steeply dipping where S_3 is weak.

inconclusive (de Roo et al., 1990; table I). On the basis of clear overprinting data presented here, the "D_h" and "D_v" structures of de Roo et al. (1990) can now be classified as D_3 and D_4 deformations, respectively.

The fifth deformation (D_5) is represented by a weak, westerly striking, cleavage (S_5) that is easily recognized as a vertical foliation of evenly spaced, dark seams. In one key outcrop S_5 overprints F_4 folds of S_3. Steeply dipping kink bands are locally present in the area. They could not be dated, but regionally they are associated with the D_5 deformation.

Large-scale structures

The almost invariable parallelism between S_1 and the trend of lithological boundaries and the symmetrical distribution of felsic volcanic rocks in the vicinity of the Half Mile Lake massive sulphide deposits, relative to the meta-

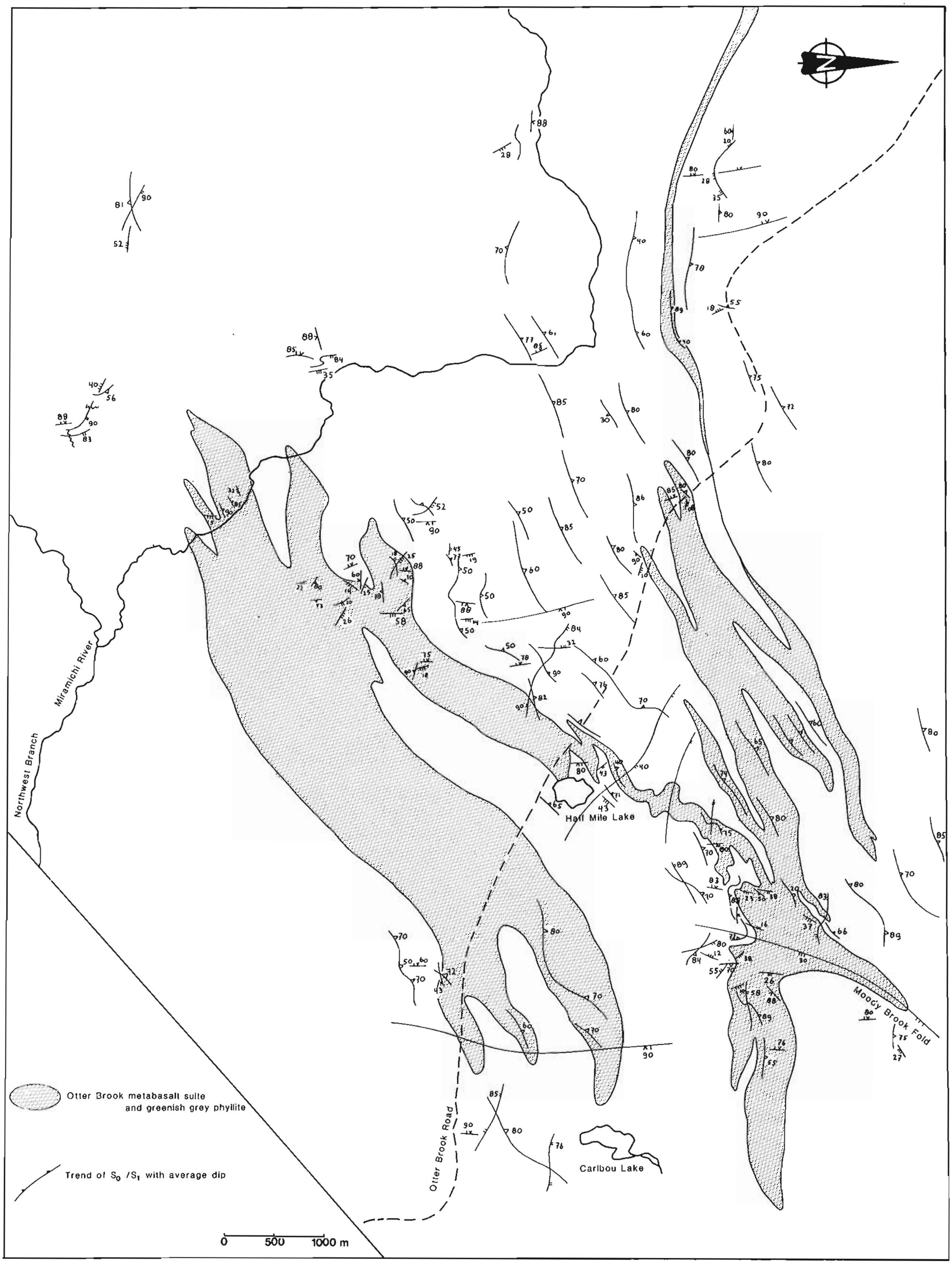

Figure 5. Structural map of the Half Mile Lake region with the trends and average dips of the foliations. The relative age of the foliation surfaces, except S_0/S_1, is marked by roman numerals. The Half Mile Lake Fold is outlined by the contact between the basalts and the rhyolites of the Flat Landing Brook Formation.

sedimentary rocks suggest that the latter mark the hinge of a macroscopic F_1 fold of S_0 (the Half Mile Lake Fold; Fig. 2). This interpretation is consistent with available younging indicators in the sedimentary rocks and massive sulphides (see above), which indicate that this fold is an anticline that is presently overturned towards the southeast. Although mesoscopically the orientation of S_1 is highly variable due to later folding, the F_1 age of the Half Mile Lake fold is supported by field relationships in an indirect way. S_5 is weakly developed, so that S_4 is mostly the youngest cleavage present in the outcrops. S_4 consistently has a northerly strike. In all outcrops where both S_2 and S_4 foliations could be demonstrated unambiguously, overprinting relationships systematically indicated that F_4 folds of S_2 have an S-asymmetry. If the Half Mile Lake Fold were an F_2 fold, S_2 should follow the trend of stratigraphic contacts, so F_4 folds of S_2 should have a Z-asymmetry, and the S_2 form surface should have a northeasterly strike (Fig. 2). Instead, the S_2 data suggest a northwesterly striking form surface. Hence, the Half Mile Lake Fold must be an F_1 structure. The overall attitude of S_2 and S_4 (Fig. 5) strongly suggest that, macroscopically, this fold has been refolded by a large NW-SE trending open fold, which is defined by the enveloping surface of smaller scale, nearly symmetrical F_2 folds (Fig. 2). This F_2 refolding generally resulted in open to tight macroscopic folds (Fig.2), even though mesoscopic transposition of S_0/S_1 into S_2 is present in outcrops. (Similar relationships have been recorded in the "B" -zone sulphide deposit at Heath Steele Mines. The "B" -zone is a slab that is oriented parallel to S_1 despite mesoscopic F_2 transposition of S_1 (de Roo et al., 1990).

Across the Half Mile Lake Fold, S_3 not only changes dip, but also strike, so that S_3 shows a dome and basin pattern (Fig. 6). Regionally, this is the result of F_4 and F_5 upright folds having axial planar cleavages at high angles to one another (S_4 and S_5). Regional-scale F_3 folds are rare in the Bathurst Camp. However, a macroscopic-scale F_3 structure, the Moody Brook fold, is outlined by the outcrop pattern of the metabasalts in the northeastern part of the Half Mile lake region (Fig. 2).

The relative orientations of S_4 and S_5, i.e. S_4 is at a high angle, clockwise from S_5, corresponds to those at Heath Steele Mines (Fig. 1), where S_4 is overprinted by a WSW- striking S_5 (McBride, 1976; de Roo et al., 1990). In the Half Mile Lake region both cleavages have rotated about 30° clockwise, because of post-S_5, open, folding on a regional scale (van Staal, unpubl. results), related to deformation that produced the giant Nine Mile Synform and Tetagouche Antiform (van Staal, 1986). The Half Mile Lake and Heath Steele Mines regions occupy opposite limbs on the Nine Mile Synform.

CONCLUSIONS

Major, relatively open F_2 folds with local, mesoscopic, transposition of S_1 have also been recognized in areas to the north and south of the Half Mile Lake area (Helmstaedt, 1971; Irrinki, 1986). To the north, F_2 folds are upright with axial planes steeply dipping to the west. Van Staal (1987) correlated this zone of upright F_2 folding with the zone of recumbent F_2 folding, and referred to them as "steep belts" and "flat belts", respectively. De Roo et al. (1990) showed that, mesoscopically, the flat belts are F_3 folded D_2 steep belts (Fig. 4b). The presence of overturned, noncylindrical F_1 folds in steep belts at Brunswick Mines (van Staal and Williams, 1984, 1986; van Staal, 1985) and at Heath Steele Mines (de Roo et al., 1990) illustrates the regional consistency of F_1/F_2 structural relationships.

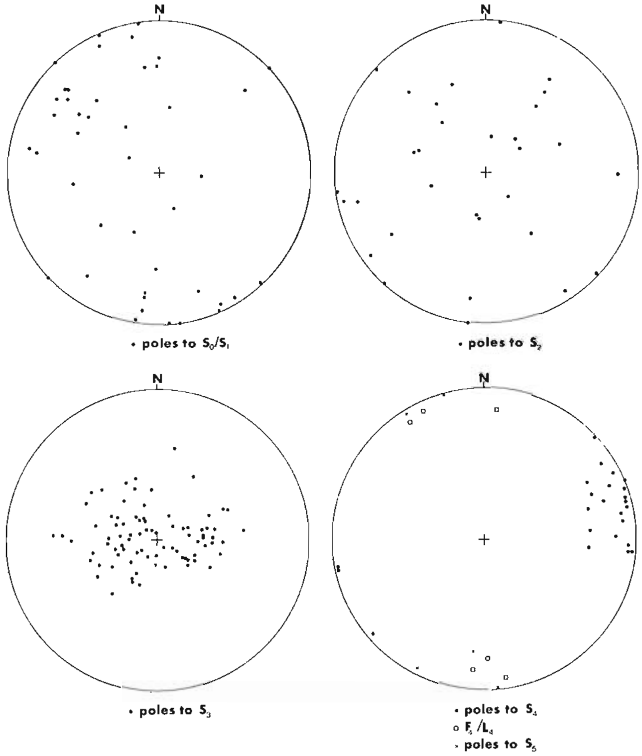

Figure 6. Lower hemisphere equal area projections of structural elements measured in the Half Mile Lake region.

Overprinting relationships in the Half Mile Lake region clarify the scheme of tectonic age relationships used so far. The "F_h" recumbent folds of de Roo et al. (1990) must be F_3 structures. Therefore the North Little River Lake Fold (Fig. 1) is transected by S_4 (de Roo et al., 1990; Fig. 10b,c) and be a major F_2 structure. This fold coincides with an F_4 synform of S_3. The clockwise rotation of S_4 and S_5 between Heath Steele Mines and Half Mile Lake suggests that the Half Mile Lake F_1 Fold and the North Little River Lake F_2 Fold have been deformed by open folding of far larger proportions, and now occupy the hinge of the giant Nine Mile Synform. The form surface of S_3 indicates the presence of several large F_4 folds in this area as well (Fig. 5).

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Revised late Precambrian stratigraphy near Saint John, New Brunswick: discussion

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Pickerill, R.K. and Fyffe, L.R., Revised late Precambrian stratigraphy near Saint John, New Brunswick; in Current Research, Part D, Geological Survey of Canada, Paper 91-1D, p. 187-189, 1991.

The recent paper by Currie (1989) on the late Precambrian and earliest Cambrian stratigraphy of the Saint John area of southern New Brunswick is a welcome contribution to the study of this geologically complex region. Nevertheless, Currie (1989) raised several controversial issues particularly with respect, but not necessarily confined to, stratigraphic nomenclature and which form the basis for this discussion.

Currie (1989) stated "Recent studies by Tanoli (1987) on the stratigraphy of the Saint John Group show that formations above the Ratcliffe Brook belong to a single environment (barrier bar complex) which did not produce the Ratcliffe Brook Formation. It would therefore appear logical to include the Ratcliffe Brook Formation and the three underlying units in a different group spanning the time period from latest Precambrian to earliest Cambrian. These rocks accumulated under subaerial conditions in an environment dominated by gradually decreasing volcanic activity." With respect to these statements we wish to draw attention to the following. First, formations above the Ratcliffe Brook Formation do not belong to a single environment of a barrier bar complex. Admittedly, the immediately overlying Glen Falls Formation represents strata interpreted by Tanoli and Pickerill (1987) as remnants of a barrier island system retreated by shoreface erosion, but all succeeding formations of the Saint John Group were deposited in shallow to deep marine shelf environments (see Tanoli and Pickerill, 1988, 1989). Second, the Ratcliffe Brook Formation was not entirely deposited under subaerial conditions. On the contrary, only the basal conglomerates, or Facies RB1 of Tanoli and Pickerill (in press), of the formation represent probable subaerial deposits, and these constitute only a small proportion of the entire thickness of the sequence. Overlying lithofacies (Facies RB2 to RB6 of Tanoli and Pickerill (in press)) were deposited in transitional to shallow marine environments, possibly in association with a delta (Hofmann and Patel, 1989; Tanoli and Pickerill, in press), as a consequence of the early Cambrian eustatic sea-level rise (Tanoli and Pickerill, in press). The terrestrial setting, first suggested by Hayes and Howell (1937) and now reiterated by Currie (1989), ignores the fact that the strata contain marine fossils (Matthew, 1889, 1899; Walcott, 1900) and a definitively marine trace fossil assemblage (Hofmann and Patel, 1989). Additionally, Facies RB2 to RB6 of Tanoli and Pickerill (in press) contain marine authigenic minerals such as chamosite and glauconies (*sensu* Odin and Letolle, 1980). Third, we wish to draw the attention of the author to Section 2.4.10 of Nystuen (1989, p. 13) and Article 15 of the North American Code of Stratigraphic Nomenclature (1983, p. 854) which both emphasize that the mode of formation has no decisive significance for the definition of most categories of geological units. Thus, despite the fact that even though units underlying the Ratcliff Brook Formation, particularly the Eocambrian Unit Es of Currie (1989), formed subaerially, any formal stratigraphic nomenclatural decision should not be dictated by depositional environment. As also noted above, much of the Ratcliffe Brook Formation is demonstrably marine in origin and, therefore, its inclusion within a different group based on such a consideration is automatically erroneous.

Currie (1989) stated that the underlying redbeds (of his Unit Es) closely resemble the Ratcliffe Brook Formation. On this observation we have no disagreement, and as noted by Tanoli et al. (1985) and Tanoli and Pickerill (1988) many previous workers in the Saint John area (e.g. Ells, 1908; Hayes and Howell, 1937; Yoon, 1970) erroneously included such Eocambrian redbeds within the Ratcliffe Brook Formation. Nevertheless, Tanoli et al. (1985) and Tanoli and Pickerill (1988) have discussed several criteria which in combination can be utilized to distinguish between the two sequences. Even Currie (1984, 1986a, 1987) apparently found little difficulty in their distinction. The fact that problems may be encountered in allocating small or isolated exposures of redbeds to either Currie's (1989) Eocambrain Unit Es or to the Ratcliffe Brook Formation, particularly where many of the criteria proposed by Tanoli et al. (1985) and Tanoli and Pickerill (1988) do not necessarily apply, is not unusual in any regional geological study. However, this should not necessitate the inclusion of the two sequences as part of a newly defined lithostratigraphic unit. The Ratcliffe Brook Formation is now firmly entrenched in the literature as representing the earliest Cambrian lithostratigraphic unit of

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the Saint John Group (Landing et al., 1989) and its potential removal from this group is not entirely logical. The descriptor Ratcliffe Brook Formation when first introduced by Hayes and Howell (1937) was simply a lithostratigraphic replacement name for the chronostratigraphic "Etcheminian Series" first proposed by Matthew (1889). It passes gradationally and conformably into the overlying Glen Falls Formation (see Tanoli and Pickerill, 1988, Fig. 5b, p. 676) whereas it disconformably overlies the Coldbrook Group and typically disconformably overlies redbeds of Unit Es of Currie (1989). As noted by Currie (1989), Unit Es also contains lithologically distinctive basalts and basaltic tuffs which are decidedly absent within the Ratcliffe Brook Formation. All the above observations suggest to us that if, in fact, a new group is proposed for the various Eocambrian units described by Currie (1989), it should not necessarily include the early Cambrian Ratcliffe Brook Formation.

Currie (1989) recognized two divisions within strata previously included within the Coldbrook Group (cf. Alcock, 1938), namely an unnamed Eocambrain bimodal volcanic sequence and an underlying subduction-related Hadrynian volcanic sequence which he referred to as the Coldbrook Group. It should be noted, however, that Article 19 of the North American Code of Stratigraphic Nomenclature (1983, p. 855) specifically recommends that when a unit is divided into two or more of the same rank as the original, the original name should not be used for any of the divisions in order to avoid the confusion engendered by such usage.

Currie (1989), contrary to the previous usage of Leavitt (1963) and Hamilton (1965), also effectively redefined the upper boundary of the Greenhead Group by excluding the Martinon Formation from it. The graded siltstones and basal limestone-cobble conglomerates of the Martinon Formation were interpreted, respectively, as distal and proximal slope facies in contrast to the carbonate-quartzite-shale shelf facies of the Ashburn Formation. An unconformity was assumed to mark the mutual boundary of these formations since strata in the Martinon Formation contained lithified clasts derived from the underlying Ashburn Formation; justification for redefining the upper boundary of the Greenhead Group rests on the significance attributed to this hiatus. The presence of basaltic sills and flows(?) within the Martinon Formation initially led Currie (1986a,b) to suggest a closer link to the Hadrynian Coldbrook Group than to the Helikian Greenhead Group, thereby implying a considerable time gap between deposition of the two formations. This view, however, is apparently no longer held by Currie (1989) who stated that the basalts resembled dykes in the Greenhead Group rather than flows in the Coldbrook Group. Given the uncertainty with regard to the nature of the unconformity, it seems premature to modify the existing nomenclature, particularly as this should, according to Article 19 of the North American Code of Stratigraphic Nomenclature (1983, p. 855), preclude retention of the well-established descriptor "Greenhead Group".

Naturally, lithostratigraphy involves many philosophical differences and these differences must all be respected. In spite of these differences, a large degree of mutual understanding and co-ordination in usage can be achieved without serious sacrifice to anyone's philosophy if a sufficiently tolerant attitude is maintained (Hedberg, 1972). As in many paleontological studies we feel that historical precedence may play an equally important role in the redefinition of stratigraphic sequences. If, however, at some future date, the Ratcliffe Brook Formation is to be included within a newly defined group and the Martinon Formation excluded from the Greenhead Group, all changes must be accommodated within the North American Code of Stratigraphic Nomenclature (1983, Article 17, p. 854), which states "Formally defined and named geologic units may be redefined, revised or abandoned, but revision and abandonment require as much justification as establishment of a new unit."

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Revised late Precambrian stratigraphy near Saint John, New Brunswick: reply

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Editorial policy of Current Research forbids introduction of new formal stratigraphic terminology. Currie (1989) therefore made no proposals for such nomenclature, but pointed out that terminology of the late Precambrian geology of the Saint John region requires revision. Fyffe and Pickerill (1991) argue that such a revision should adhere to the International Stratigraphic Code, although the names now in use were not originally defined according to the code, and strict application of the code would result in disappearance of internationally recognized stratigraphic names. Currie (1989) suggested that revision should preserve as many entrenched stratigraphic names as possible, with revised definitions to bring them into conformity with the code. In particular Fyffe and Pickerill (1991) point out that retention of the name "Coldbrook Group" for a part of the late Precambrian volcanic rocks originally grouped by Alcock (1938) into the Coldbrook Group would violate the code. This argument would carry more weight if the Coldbrook Group had a well-defined type section, lithology, base, and top as required by the code. The name appears in many and varied publications, and the general character of the rocks is internationally known. A properly defined Coldbrook Group would avoid confusion in the literature. Arguments for and against redefinition of the name Coldbrook Group apply, *mutatis mutandis*, to other Avalonian terranes in the Atlantic Provinces where similar successions have been discovered. The volcanic-dominated section in the Saint John region is bounded below by a fault and above by the base of the Saint John Group. An internal unconformity divides the volcanic rocks into a lower, generally andesitic unit, and an upper bimodal volcanic-sedimentary unit. The former corresponds exactly to the original definition of Coldbrook Group (Alcock, 1938). Possible nomenclature for these rocks can be summarized as follows: (i) restrict the name Coldbrook Group to the strata below the unconformity, and place all strata above it in the Saint John Group, defining new formations as necessary; (ii) divide the strata now placed in the Coldbrook Group into two (or more) new divisions, leaving Saint John Group intact; (iii) divide the Coldbrook Group, and redefine the base of the Saint John Group as convenient. Proposal (i) would not breach the stratigraphic code (it revises only map boundaries of established units) and preserves established nomenclature. It would also emphasize the essentially continuous stratigraphic section across the Precambrian-Cambrian boundary in the Saint John area, an important geological fact that has been obscured by present terminology. Proposal (ii), favoured by Fyfe and Pickerill (1991), would require new names for divisions of the former Coldbrook Group, even though one division would exactly match the (incomplete) definition of the original unit. They explicitly assume that the base of the Saint John Group is well defined. Proposal (iii), favoured by Currie (1989), assumes the base of the Saint John Group (base of the Ratcliffe Brook Formation) to be as poorly defined as the Coldbrook Group. Fyffe and Pickerill (1991) suggest that Ratcliffe Brook Formation (a) grades into younger units of the Saint John Group, (b) is separated from older units by unconformity or disconformity, (c) is partly or mainly marine, like the rest of the Saint John Group, and (d) lacks a basaltic volcanic component. I believe (b) to be false, and (a), (c), and (d) to be equally applicable to strata below the Ratclifffe Brook Formation as presently mapped. This factual disagreement can be resolved by further field work.

Fyffe and Pickerill (1991) also object to excision of the Martinon Formation from the Greenhead Group (or Green Head Group) on somewhat similar terminological grounds. The terminology of the Greenhead Group has as little formal foundation as that of the Coldbrook Group, and no formal case for inclusion of the Martinon Formation has ever been made. The Martinon Formation differs from the rest of the Greenhead Group in lithology, origin (turbidites versus shelf-type sediments), and deformation style. It contains clasts, some deformed, of Greenhead Group.

The proposal that we should on one side abandon long-established nomenclature, and on another refrain from correcting an obvious error in stratigraphic attribution, all to maintain a supposed "correctness" in nomenclature, seems to me an assault by terminological pedantry upon geological common sense and upon the pre-eminence of field observation in constructing both maps and stratigraphic columns. Potential loss of well-established names has motivated a number of violations of the stratigraphic code, for example division of the Gander Lake Group of Newfoundland into Gander Lake and Davidsville groups (Kennedy and McGonigal, 1972). The proposed "rectification" into Gander and Davidsville groups (Kennedy, 1975) could be paralleled in the Saint John region by renaming the residual strata Cold Brook Group, which would conform to the code rule that groups be named after topographic features.

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AUTHOR INDEX

NOTE TO CONTRIBUTORS

Submissions to the *Discussion* section of *Current Research* are welcome from both the staff of the Geological Survey of Canada and from the public. Discussions are limited to 6 double-spaced typewritten pages (about 1500 words) and are subject to review by the Chief Scientific Editor. Discussions are restricted to the scientific content of Geological Survey reports. General discussions concerning sector or government policy will not be accepted. All manuscripts must be computer word-processed on an IBM compatible system and must be submitted with a diskette using Word Perfect 5.0 or 5.1. Illustrations will be accepted only if, in the opinion of the editor, they are considered essential. In any case no redrafting will be undertaken and reproducible copy must accompany the original submissions. Discussion is limited to recent reports (not more than 2 years old) and may be in either English or French. Every effort is made to include both *Discussion* and *Reply* in the same issue. *Current Research* is published in January and July. Submissions should be sent to the Chief Scientific Editor, Geological Survey of Canada, 601 Booth Street, Ottawa, Canada, K1A 0E8.

AVIS AUX AUTEURS D'ARTICLES

Nous encourageons tant le personnel de la Commission géologique que le grand public à nous faire parvenir des articles destinés à la section *discussion* de la publication *Recherches en cours.* Le texte doit comprendre au plus six pages dactylographiées à double interligne (environ 1500 mots), texte qui peut faire l'objet d'un réexamen par le rédacteur en chef scientifique. Les discussions doivent se limiter au contenu scientifique des rapports de la Commission géologique. Les discussions générales sur le Secteur ou les politiques gouvernementales ne seront pas acceptées. Le texte doit-être soumis à un traitement de texte informatisé par un système IBM compatible et enrégistré sur disquette Word Perfect 5.0 ou 5.1. Les illustrations ne seront acceptées que dans la mesure où, selon l'opinion du rédacteur, elles seront considérées comme essentielles. Aucune retouche ne sera faite au texte et dans tous les cas, une copie qui puisse être reproduite doit accompagner le texte original. Les discussions en français ou en anglais doivent se limiter aux rapports récents (au plus de 2 ans). On s'efforcera de faire coïncider les articles destinés aux rubriques *discussions* et *réponses* dans le même numéro. La publication *Recherches en cours* paraît en janvier et en juillet. Les articles doivent être envoyés au rédacteur en chef scientifique: Commission géologique du Canada, 601 rue Booth, Ottawa, Canada, K1A 0E8.

Geological Survey of Canada Current Research, is now released twice a year, in January and in July. The four parts published in January 1991 (Paper 91-1, parts A to D) are listed below and can be purchased separately.

Recherches en cours, une publication de la Commission géologique du Canada, est publiée maintenant deux fois par année, en janvier et en juillet. Les quatre parties publiées en janvier 1991 (Étude 91-1, parties A à D) sont énumérées ci-dessous et vendues séparément.

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Partie D, Est du Canada et programmes nationaux et généraux