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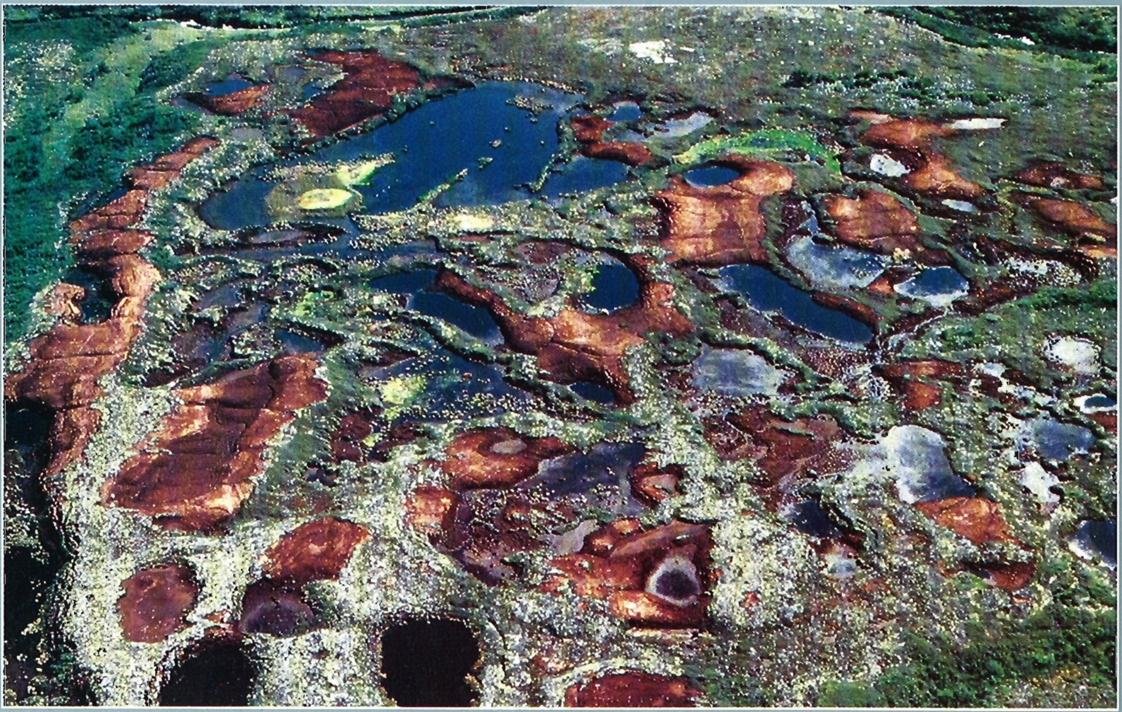

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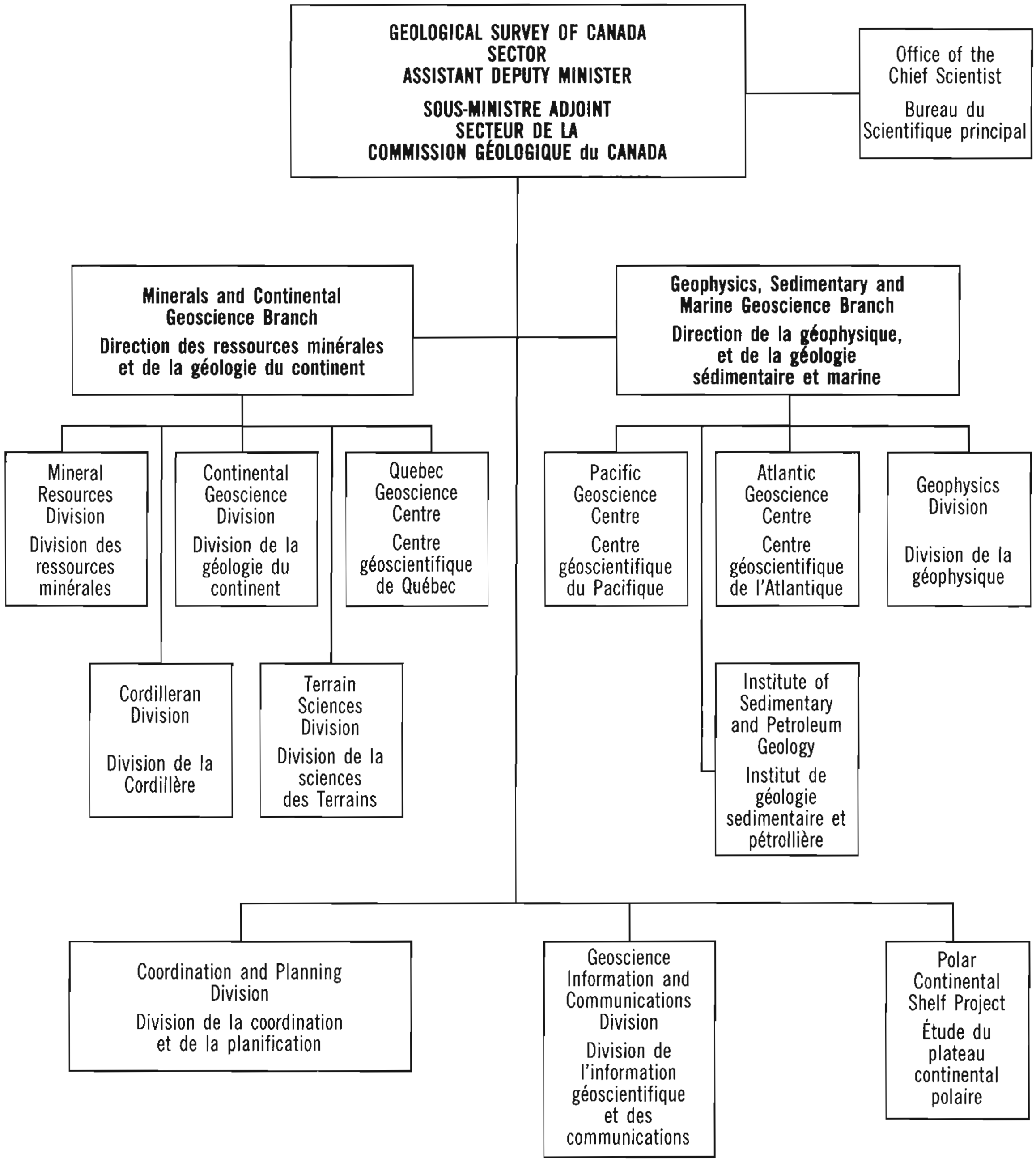
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CONTENTS

Preliminary report on the structural control of the Rendell-Jackman gold deposit, Springdale Peninsula, Newfoundland[1,2]

B. Dubé, K. Lauzière, and D. Gaboury[3]
Quebec Geoscience Centre, Quebec

Dubé, B., Lauzière, K., and Gaboury, D., 1992: Preliminary report on the structural control of the Rendell-Jackman gold deposit, Springdale Peninsula, Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 1-10.

Abstract

The Rendell-Jackman gold deposit is one of the most significant mesothermal vein-type gold deposits in the Canadian Appalachians. It is located within a deformed Ordovician metavolcanic dominated sequence containing felsic dykes.

The first episode of deformation is evidenced in the metavolcanics by an S_1 planar fabric containing a down dip stretching lineation. The second episode (D_2) produced steeply plunging E-W folds, an S_2 foliation and discrete brittle-ductile D_2 high strain zones, affecting the metavolcanics and the felsic dykes. These brittle-ductile D_2 high strain zones are particularly well developed in the hinges of F_2 folds.

They are also strongly controlled by the layer anisotropy induced by the felsic dykes which are competent compared to the incompetent metavolcanics. The gold bearing quartz veins are of shear vein type and hosted within and sub-parallel to the D_2 high strain zones. The last deformation episode produced F_3 sub-horizontal folds and late NE and ENE brittle faults.

Résumé

Le gisement aurifère de Rendell-Jackman est un des plus importants dépots mésothermaux de type filonien dans les Appalaches canadiennes. Le gisement est situé dans une séquence métavolcanique d'âge ordovicien, déformée et injectée de dykes felsiques.

La première phase de déformation se caractérise par le développement dans la séquence de roches métavolcaniques, d'une foliation S_1 contenant une linéation d'étirement à fort plongement. La deuxième phase (D_2) a produit des plis d'orientation est-ouest à fort plongement, une foliation S_2 et des zones discrètes D_2 fragiles et ductiles très déformées, présentes dans les métavolcaniques et les dykes felsiques. Ces dernières sont particulièrement bien développées dans les charnières des plis P_2.

Ces zones sont également fortement contrôlées par l'anisotropie de couches induite par les dykes felsiques compétents par rapport aux roches métavolcaniques incompétentes. Les veines de quartz aurifères sont de type «cisaillement» et sont contenues et orientées parallèlement aux zones D_2 très déformées. La dernière phase de déformation a produit des plis P_3 sub-horizontaux et des failles cassantes tardives, d'orientation nord-est et est-nord-est.

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[1] Published with the authorization of Noranda Exploration Co. Ltd

[2] Contribution to Canada-Newfoundland Corporation Agreement on Mineral Development 1990-1994. Project C1.124, carried out by the Quebec Geoscience Centre of the Geological Survey of Canada.

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INTRODUCTION

The Rendell-Jackman gold deposit in northwestern Newfoundland is one of the most significant mesothermal vein-type gold deposit in the Canadian Appalachians (Dubé, 1990). It is located near King's Point on the Springdale Peninsula (Fig.1). It was discovered by Noranda Exploration in 1987 and recent calculations estimate geological reserves (undiluted) in the Hammer Down deposit at 429,593 tonnes grading 11.6 g/t Au (cut to 34.3 g/t) (Andrews, 1990). The mineralization, located in sulphide-rich quartz veins, is within discrete high strain zones spatially associated with the Green Bay fault, a poorly exposed northeasterly striking fault (Fig.1). The area has been previously studied on a regional scale by, among others, Neale et al., (1960), Marten (1971), Kean (1988) and Szybinski and Jenner (1989) and on a local scale by Noranda Exploration geologists (Andrews, 1990; Huard, 1990).

Detailed structural mapping was undertaken in order to unravel the structural history and determine the controls on gold mineralization. It is part of a long-term project on structure and gold deposits in the Canadian Appalachians (Dubé, 1990). This paper summarizes the results of fieldwork done on the stripped outcrops of this area during the summers of 1990 and 1991.

REGIONAL GEOLOGICAL SETTING

The Springdale Peninsula is located in the northwestern part of the Dunnage Zone of Williams (1979). It is made up of Cambro-Ordovician ophiolitic rocks of the Lush's Bight Group, overlain by a thick sequence of Lower to Middle Ordovician volcanic, volcaniclastic and epiclastic rocks, which include the Catcher's Pond Group, and Silurian subaerial volcanics and fluviatile redbeds of the Springdale Group (Kean, 1988). Carboniferous sedimentary rocks are unconformably deposited over these rocks. According to Kean (1988), the pre-Middle Ordovician rocks contain an L-S foliation which trends northeast or southeast and contains a steeply plunging stretching lineation. This foliation is axial planar to regionally developed early folds. Mineralized and barren chloritic shear zones are also widely developed in the Lush's Bight Group. Marten (1971) recognized a series of sinistral chlorite schist faults sub-parallel to a northeasterly trending cleavage and interpreted them as related to folding. Later structural events include crenulations, north-northeast trending kink bands, open folds and faults (Kean, 1988). The poorly exposed north-east trending Green Bay fault has both a postulated dextral (Marten, 1971) and sinistral (Jenner and Szybinski, 1987) movement.

GEOLOGY OF THE RENDELL-JACKMAN GOLD DEPOSITS

Three gold zones outcrop on the Rendell Jackman property and are known as the Hammer Down deposit and the Muddy Shag and Rumbullion showings (Fig.1). Excellent exposure of these three mineralized zones displays the relatively complex geology.

Gold mineralization on the Rendell-Jackman property occurs within the Lower to Middle Ordovician Catcher's Pond Group (Andrews, 1990). The deposit is located within a metavolcanic sequence which consists of pillowed, brecciated and massive basaltic flows, volcaniclastics, tuffs, gabbroic dykes and sills. The common occurrence of small

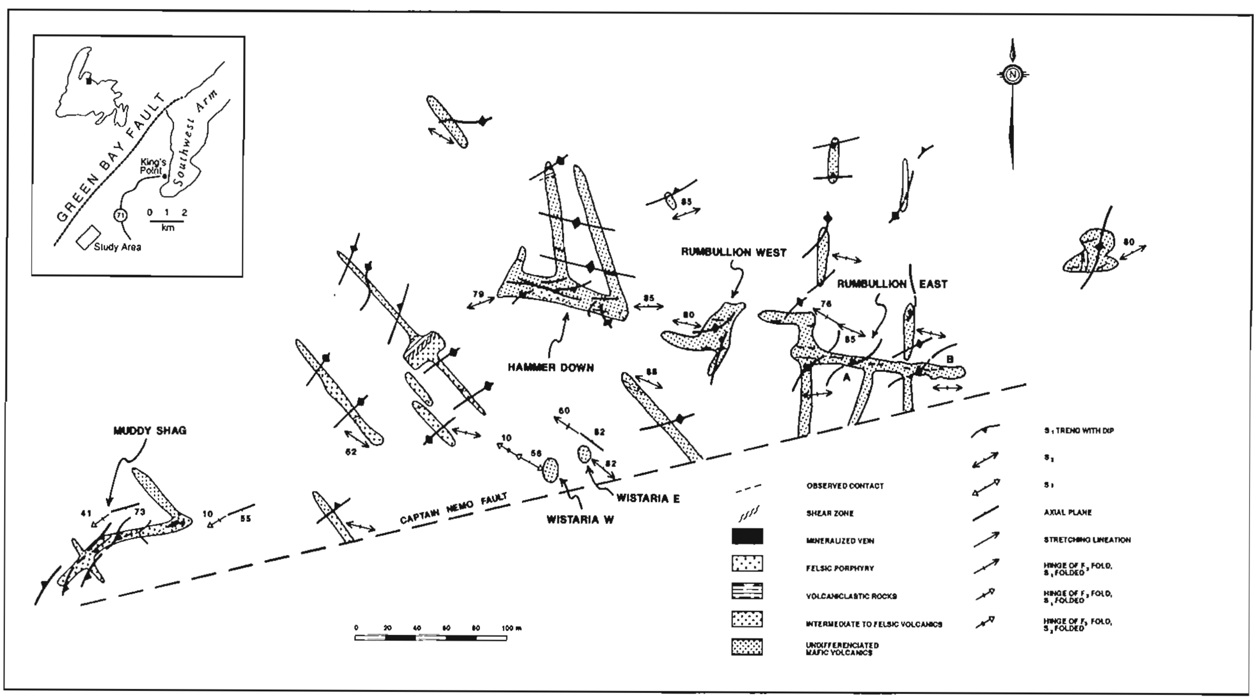

Figure 1. Simplified and preliminary geological map of the Rendell-Jackman gold deposit, showing the location of the stripped outcrops as well as the main structural elements.

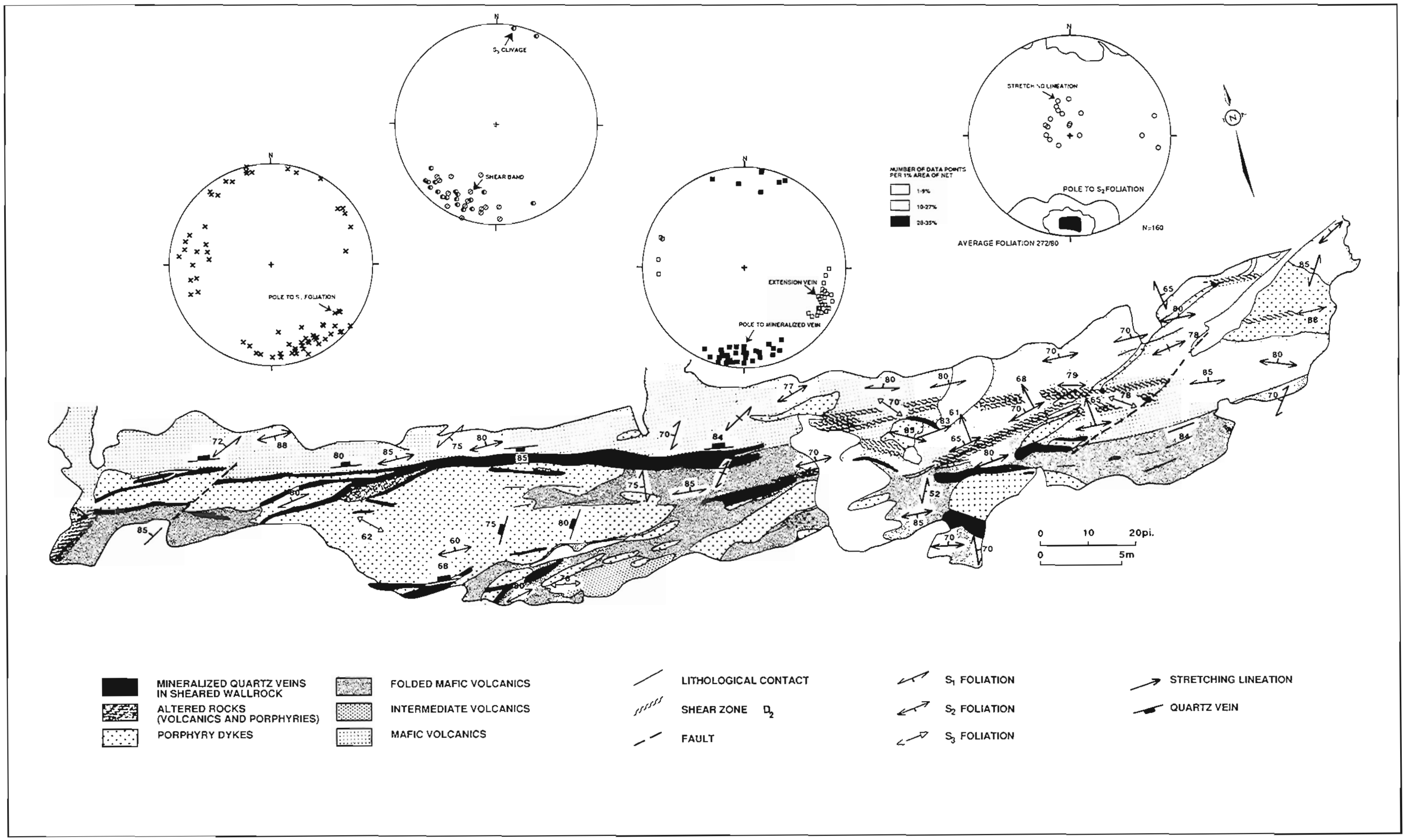

Figure 2. Detailed geological map of Hammer Down showing the structural elements, the spatial relationship between the felsic porphyry dykes and the mineralized veins and equal area projections of structural data (lower hemisphere).

leucoxene grains suggests that these mafic rocks are of tholeiitic affinity. Felsic to intermediate rocks, including lapilli and ash flow tuffs, rhyolites and dacites, as well as sedimentary rocks such as argillites and polymictic conglomerates are also present. These rocks have been intruded by pinkish porphyritic felsic dykes which contain up to 20% feldspar phenocrysts (1-3mm), 5 -15% amphiboles and 5% quartz eyes. Similar quartz-porphyry dykes are common within the Catcher's Pond Group and are Devonian in age (Dean, 1977). All but the felsic porphyry dykes have undergone at least three episodes of deformation. The felsic porphyry dykes were introduced after the first episode of deformation. All units have been subject to greenschist facies metamorphism.

Host rocks

The mineralized veins are located at the boundary between the felsic porphyry dykes and mafic volcanics-volcaniclastics at Hammer Down (Fig. 2) or are hosted solely by mafic volcanics, as on Rumbullion (Figs. 3-4-5). At Rumbullion, the mafic volcanics are strongly banded to massive with small leucoxene grains. Most of the banded variety are deformed pillows and pillow breccias (Fig. 6A) but a few mafic dykes are also present. Numerous pyritic veinlets with orientations varying from sub-parallel to S_1 to sub-parallel to S_2 impart a rusty colour to the outcrop surface. At Muddy Shag, the highly banded host rocks comprise intermediate to felsic units alternating with mafic units, banded mafic volcanics and schistose to massive porphyry dykes.

Structure of the deposits

The distribution of lithologic units is complex, but critical observations on the structural control of the gold mineralization are reported here.

The first deformational episode in the metavolcanic sequence is represented by a steeply dipping bedding-parallel S_1 planar fabric defined by flattening of elements such as pillows, pillow breccias, fragments and minerals, mostly chlorite and leucoxene. The S_1 fabric is the main foliation and is particularly well developed at Rumbullion (Figs. 3-4-5). This north-east trending foliation contains a well developed down dip stretching lineation and probably corresponds to the regional L-S foliation of Kean (1988). This foliation is possibly related to thrusting or reverse high-angle faulting as suggested by the down dip stretching lineation. Many millimetre-scale calcite-rich veinlets are sub-parallel to S_1 foliation. The felsic porphyry dykes cut the S_1 foliation.

The second deformation episode (D_2) is responsible for the steeply plunging, upright, open to tight "E-W" asymmetrical parasitic folds that affect the S_0-S_1 fabric and produce an axial planar S_2 cleavage (Figs. 3-4). The hinges of these F_2 folds are coaxial with the stretching lineation present on the S_1 foliation as well as with the intersection lineation between the S_1 and S_2 foliations. The S_2 cleavage is only locally well developed and its intensity ranges from a spaced fracture cleavage to a strong schistosity transposing the S_1 foliation (Figs. 6B-C). These zones of strongly developed schistosity correspond to discrete centimetre- to metre-scale brittle-ductile D_2 high strain zones, affecting both the metavolcanic sequence and the porphyry dykes. These high strain zones are injected by quartz veins and pyrite disseminations and/or veinlets. The brittle-ductile D_2 high strain zones are related to the development of the F_2 parasitic folds and are mostly present in the hinge of F_2 fold as it is observed at Rumbullion (Figs. 3-4). These high strain zones are inferred to have developed as accommodation structures during growth and tightening of folds. At the SE part of Hammer Down, an almost perfectly perpendicular relationship between the S_2 cleavage and the S_1 penetrative fabric is observed (Fig. 2), clearly indicating the location of a F_2 fold hinge (Figs. 6B-C). However, the overall trends of lithological contacts at Hammer Down are not as suggestive of a fold closure; folds are probably cut and contacts displaced by D_2 high strain zones located in the middle part of the outcrop.

The S_2 cleavage is developed within the felsic porphyry dykes. At the eastern part of Hammer Down, discrete brittle-ductile D_2 deformation zones cut a felsic porphyry dyke and show a transposition and apparent dextral offset (Fig. 6D).

The development of the D_2 high strain zone is also strongly controlled by the layer anisotropy induced by the presence of the competent felsic porphyry dykes with the foliated and incompetent mafic volcaniclastics. This relationship is particularly obvious on Hammer Down were almost all the contacts between the felsic porphyry dykes and the mafic metavolcanics are characterized by an S_2 foliation or D_2 high strain zones (Fig. 2).

No well developed non-coaxial fabric is associated with either the S_1 or the D_2 fabrics. But, local millimetre to centimetre scale "E-W" non-coaxial fabrics which resemble asymmetrical extensional shear bands are present within the D_2 high strain zone on the Hammer Down showing. The "shear bands" oriented on average 300/74° clearly drag an E-W trending foliation. These "shear bands" are mainly observed in vertical section view but are also locally present in plan view. The intersection between the "shear bands" and the E-W foliation is at a relatively high angle (60°) to poorly developed northwesterly trending and moderately plunging stretching lineations locally developed on the S_2 foliation (Fig. 2). This relationship suggests that they may be indicative of a reverse-oblique movement. However, a locally developed S_3 fracture cleavage sub-parallel to the shear bands could produce a geometrically similar structure by crenulation of the S_2 foliation (Fig. 2). If this is the case, the asymmetrical fabric is kinematically insignificant and no sense of movement can be deduced. Ongoing petrographical studies of oriented thin sections will help to determine the veracity of these "shear bands" and their inferred sense of movement.

Local sub-horizontal striations plunging either east or west are developed on the S_2 foliation and are probably related to a late brittle event.

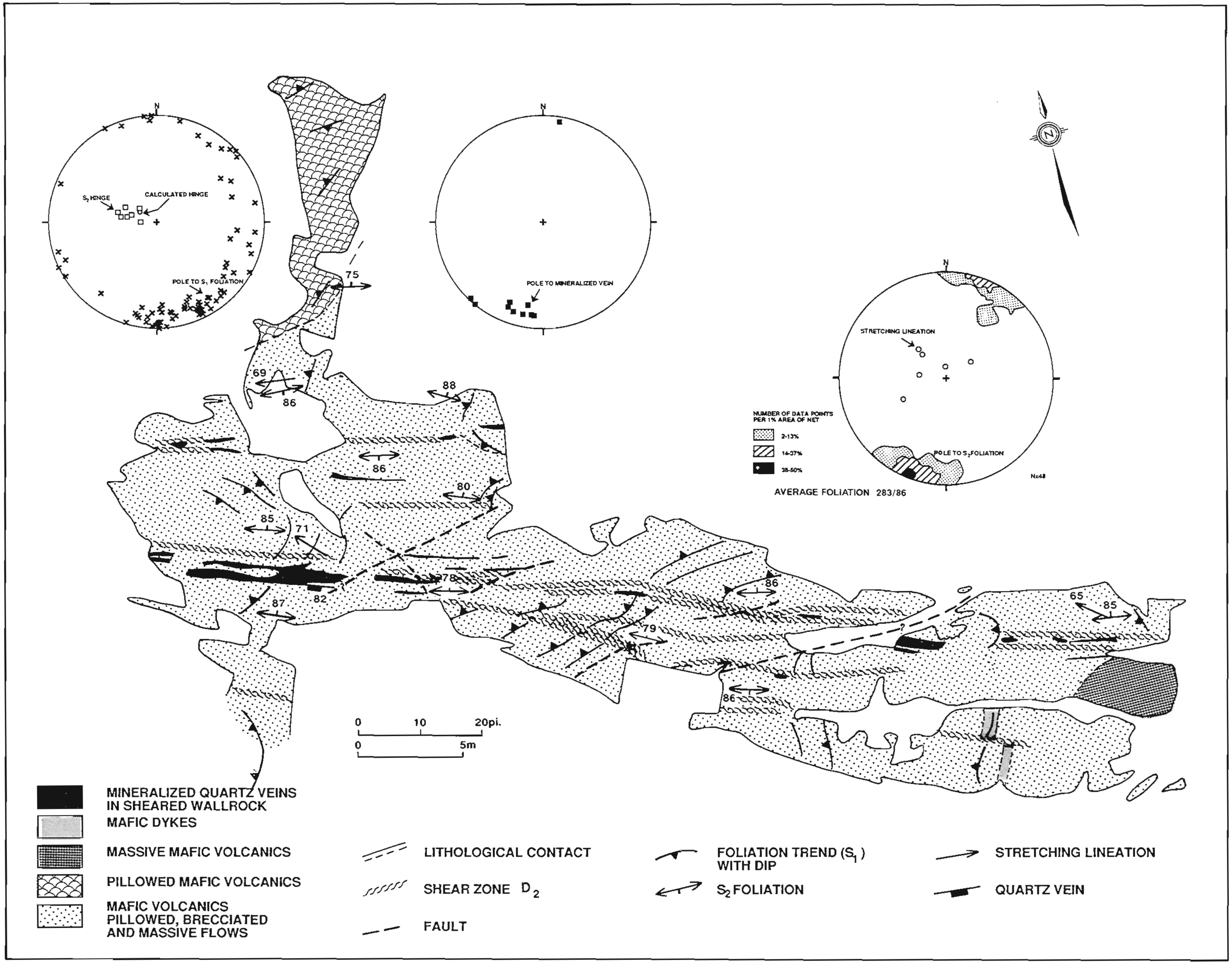

Figure 3. Detailed geological map of the Rumbullion East (A) stripped outcrop showing the structural elements, the location of the mineralized veins in the hinges of the F2 folds and equal area projections of structural data (lower hemisphere).

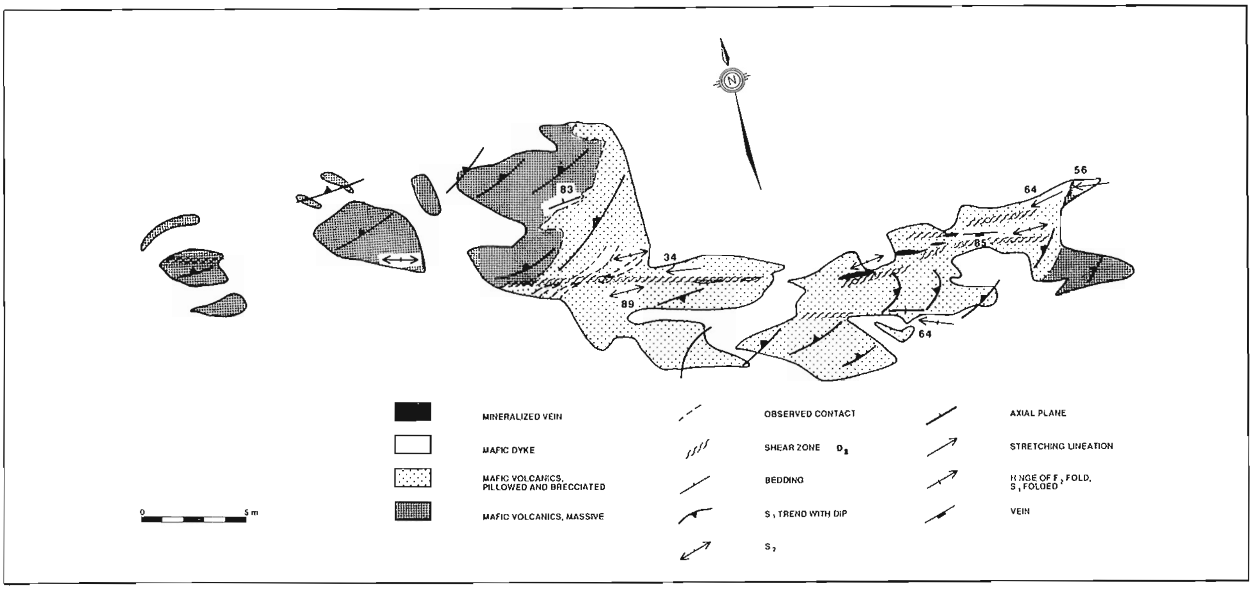

Figure 4. Detailed geological map of Rumbullion East (B) stripped outcrops showing the relationship between the S_0-S_1 fabric, the F_2 folds and the mineralized veins within the D_2 high strain zones.

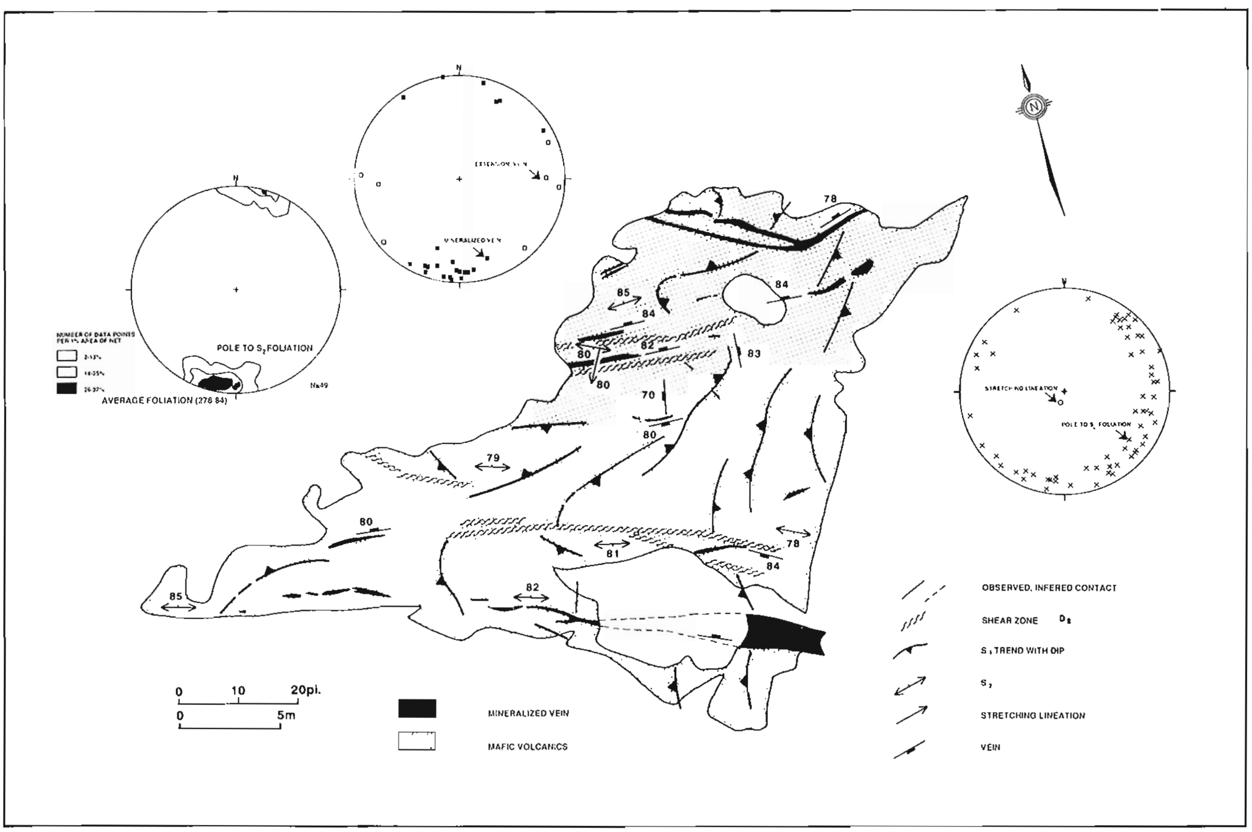

Figure 5. Detailed geological map of Rumbullion West stripped outcrops showing the relationship between the S_0-S_1 fabric and the mineralized veins and equal area projections of structural data (lower hemisphere).

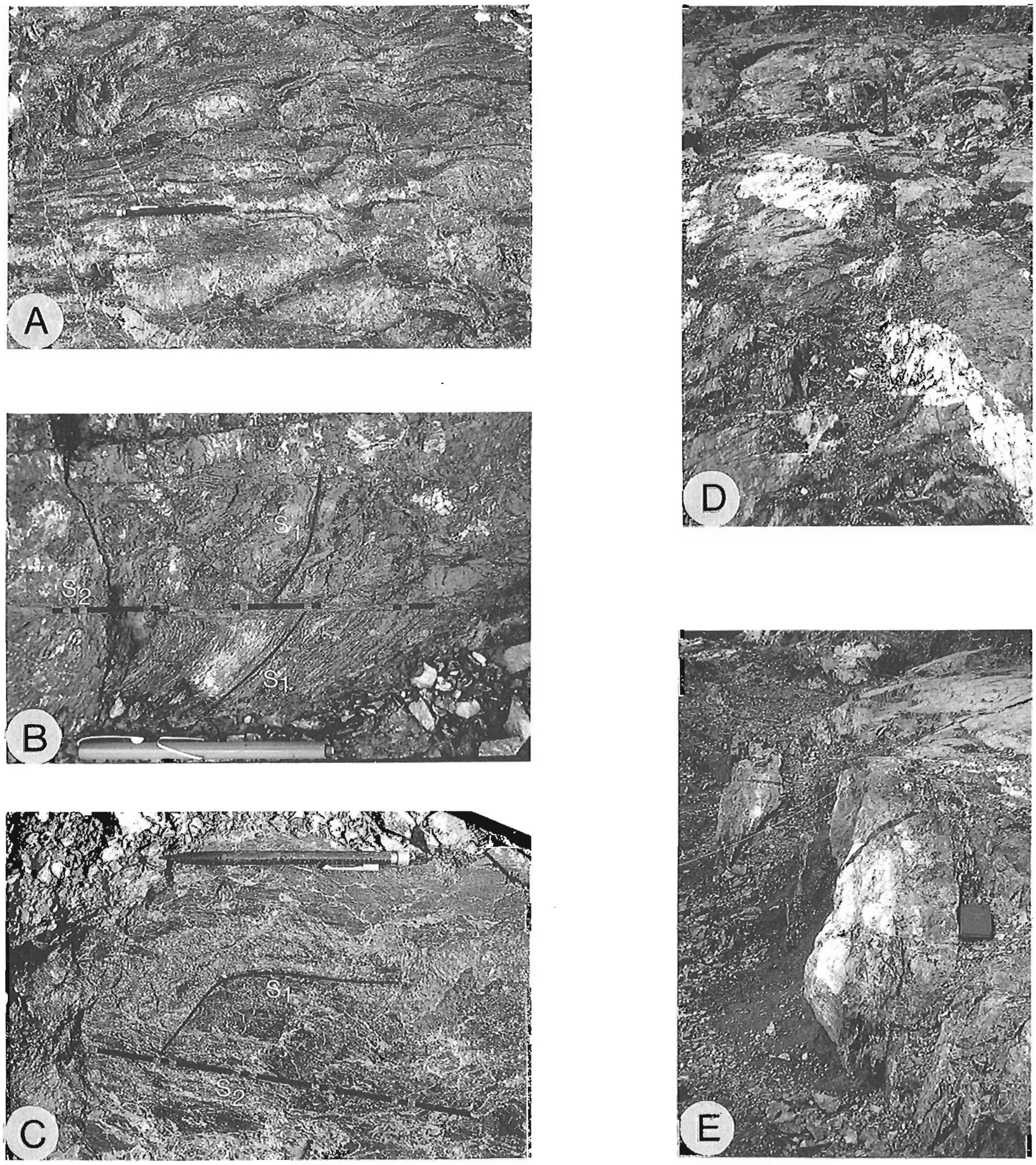

Figure 6. A) Banded aspect of the mafic volcanics produced by flattening of pillows; **B)** Spaced fracture cleavage at a high angle to the penetrative S_1 foliation, Movement occurred along some cleavage planes; **C)** Penetrative S_2 foliation transposing the S_1 foliation in the hinge of a F_2 fold; **D)** Brittle-ductile D_2 deformation zones cutting and displacing a felsic porphyry dyke with an apparent dextral offset; **E)** Example of a laminated gold-bearing quartz vein. All photographs are in plan view.

The last episode of ductile deformation produced open to tight F_3 sub-horizontal to shallow plunging folds. These F_3 folds are commonly chevron folds and are particularly well exposed on the Wistaria outcrop which represents a key area to study this deformational event (Fig. 1). There, a penetrative S1 fabric within mafic volcaniclastic rocks is folded, producing F_2 folds with an associated S_2 cleavage which, in turn, are clearly folded by F_3 folds (Fig.7). The S_2 cleavage and F_3 folds also affect the porphyry dykes. The axial plane of these F_3 folds is oriented at 298/56° and the fold axes plunge 301/10°. Similar chronological relationships have been observed at Muddy Shag (Fig.8), where felsic porphyry dykes containing strongly boudinaged sulphide-rich high grade mineralized quartz veins are both folded by sub-horizontal chevron-type F_3 folds. The shallow plunge of these folds explains the surface geometry of the mineralized veins: long limbs are straighter and thinner than short limbs and fold noses. Plunges of these F_3 folds vary because of the different pre-fold attitudes of the folded surfaces (commonly S_1 but locally S_2). The axial plane of the F_3 folds at Muddy

Shag (N072/75°) is sub-parallel to the S_2 fabric, whereas at Wistaria S_3 is different from S_2. The reason for this difference in angular relationship between the two sets of F_3 folds is unknown but they could represent conjugate D_3 fold sets.

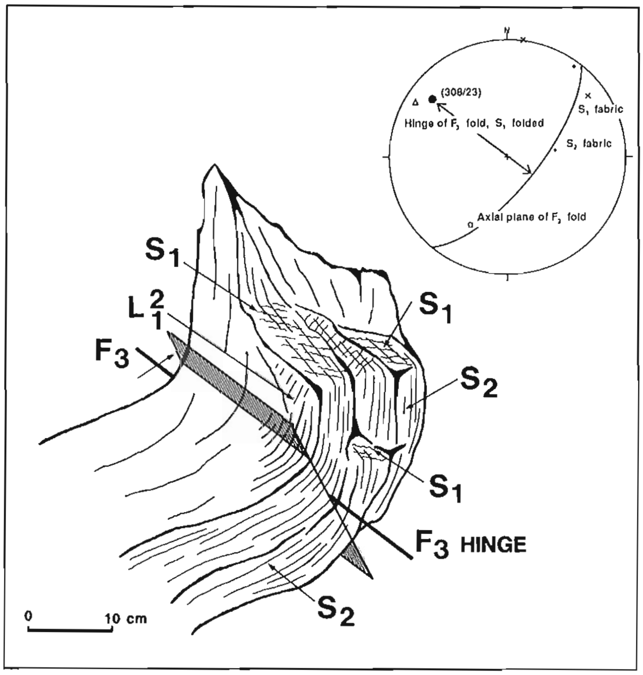

Figure 7. Sketch from a photograph showing a F3 fold and the chronological relationship between the different structural elements; Wistaria west outcrop.

Evidence of late brittle deformation is present at Hammer Down where a millimetre-scale 235/85° striking fault, observed at the western end, cuts across the mineralized veins, producing an apparent dextral transcurrent offset of 1 m (Fig.2). The fault is filled with quartz and contains traces of pyrite. Other evidence of SW-NE apparent dextral strike-slip faulting is present at the eastern end of this same outcrop where a fault clearly cuts a S_2 high strain zone. The extension of this fault to the southwest is unknown. Locally cataclasites within already strongly foliated rocks are also developed. A major ENE striking fault known as the "Captain Nemo fault" or the "Lower brittle fault" was produced during a late brittle faulting event (Andrews, 1990; Huard, 1990) (Fig.1). Drill core contains quartz vein fragments that have been incorporated within this cataclasite strongly suggesting that the actual geometry of the ore zone could also be partly controlled by movement along this fault. No significant mineralization is observed on the south side of the Captain Nemo fault (Huard, 1990). The F_3 folds are spatially associated with this fault but their timing relationship remains to be determined and ongoing structural work will

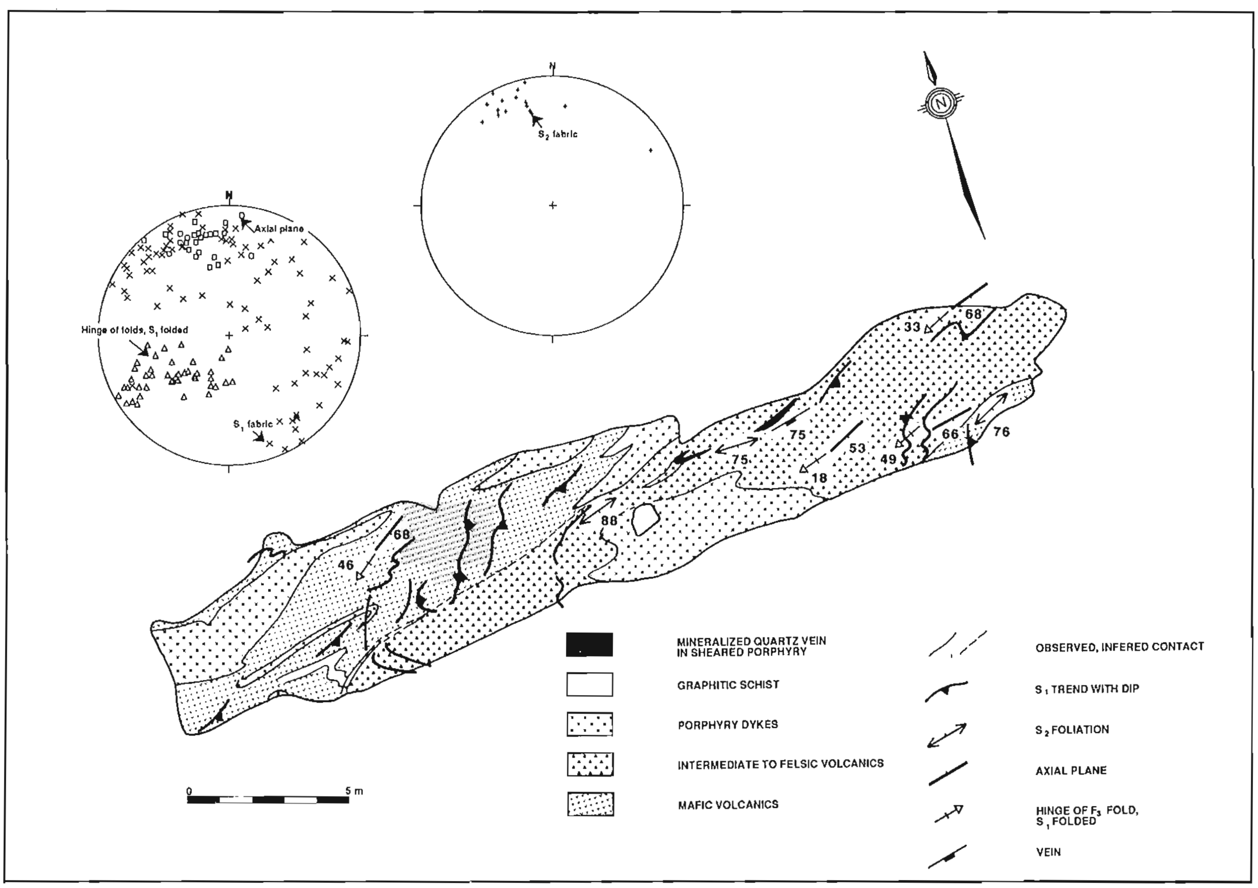

Figure 8. Detailed geological map of Muddy Shag showing the structural elements and equal area projections of structural data (lower hemisphere).

address this uncertainty. Other ENE faults were observed on the Rumbullion outcrop and show apparent sinistral offset of the vein.

Mineralized veins

Both gold bearing quartz veins and late extensional barren quartz veins have been recognized, the former being volumetrically and economically the most significant.

The gold-bearing quartz veins are laminated shear veins and are hosted within the brittle-ductile D_2 high strain zones (Fig. 6E). As mentioned above, the formation of these D_2 high strain zones is controlled by both the development of the F_2 parasitic folds (Fig. 3) and by the strong layer anisotropy induced by the behaviour of the competent felsic dykes compared to the foliated and anisotropic mafic volcanics (Fig. 2). For example, in the western part of the Hammer Down outcrop most of the felsic dyke contacts with the surrounding rocks are sites of mineralized quartz veins and their orientation is clearly spatially controlled by these contacts (Fig. 2). The mineralized veins are, as a general rule, concordant with the S_2 fabric, but some splays are oblique. Their widths vary from a few centimetres to 1 metre. The veins formed in the hinge of the F_2 folds are generally not as well developed as those adjacent to the felsic porphyry dykes. They are boudinaged and are not of great surface lateral extent. Therefore from an economic point of view, the layer anisotropy induced by the dykes was a lot more efficient in the development of the secondary permeability allowing vein formation than the F_2 parasitic folds.

The veins are very rich in gold with assays commonly greater than 170 g/t Au (Huard, 1990). The mineralogy of the mineralized veins is dominated by white to grey highly strained quartz with traces of carbonate. Sulphide percentages in the quartz veins vary from less than 5% to near massive proportions with up to 75% sulphides, composed mainly of pyrite with minor amounts of sphalerite, galena and chalcopyrite (Andrews, 1990). The sulphides are present as disseminated millimetre-size euhedral crystals or as centimetre scale aggregates. Commonly, they form millimetre to centimetre wide veinlets sub-parallel to the vein and commonly located close to the edges. Wall rock fragments incorporated within the veins are strongly elongate sub-parallel to the trend of the veins and are altered. They have been chloritized or sericitized depending on the primary composition of the protolith. Adjacent wall rocks show, generally over less than a metre, a weak hydrothermal alteration and a well developed S_2 foliation sub-parallel to the vein direction indicating that they are of shear vein-type.

At the western limit of the Hammer Down deposit, a zone of oblique mineralized veins is developed to the south of the main mineralized quartz vein in a "horse tail" type of structure. According to Andrews (personal communication) and Huard (1990), the mineralized zone terminates farther to the southwest (before reappearing at the Muddy Shag showing) whereas the deformation zone goes on (Fig. 2). This deformation zone corresponds to an area of penetrative S_2 foliation which is approximately in the same orientation as the oblique mineralized veins at the southwestern end of Hammer Down outcrop (Fig. 2). This oblique vein pattern is analogous to "termination splay faults" produced where individual faults branch into a number of diverging termination splays at their ends (Ramsay and Huber, 1987).

Veins of the second type are late extensional sub-vertical and trend north-south (Fig. 2). These veins are up to 1-2 cm wide and they commonly crosscut the mineralized shear veins. They contain euhedral millimetre- to centimetre-scale quartz crystals oriented perpendicular to the vein walls indicating open space filling. Traces of pyrite are also present. Locally, these veins have been offset along the S_2 fabric in a dextral or sinistral sense over a few centimetres suggesting late reactivation of the S_2 fabric.

Hydrothermal alteration

As mentioned by Andrews (1990) and Huard (1990), megascopically, there is no well developed hydrothermal alteration associated with the gold mineralization. Only local centimetre-scale sericite alteration in the felsic porphyry dykes and a chloritization within the mafic metavolcanics has been observed adjacent to the mineralized veins. The strong carbonatization shown by abundant calcite veinlets present in the metavolcanics adjacent to the mineralized veins at Hammer Down is probably not related to the mineralization as the veinlets are folded and oriented sub-parallel to S_1 foliation. Fluorite near and rarely at the margins of the mineralized veins is reported by Huard (1990) and Ritcey (personal communication), but no well developed alteration pattern is present. The alteration is being studied by D. Ritcey of Memorial University and will not be dealt with further here.

DISCUSSION AND CONCLUSION

This study indicates that the gold bearing quartz veins are laminated shear vein-type and are hosted within and sub-parallel to the brittle-ductile D_2 high strain zones. The formation of these D_2 high strain zones is controlled by both the development of the F_2 parasitic folds and by the strong layer anisotropy induced by the competent felsic dykes compared to incompetent and foliated mafic volcanics. This anisotropy has played a key role in the development of the secondary permeability to which the mineralized veins are related as is indicated by the common occurrence of wider and more continuous veins on the Hammer Down showing. As an empirical rule, the veins are better developed in the vicinity of the felsic dykes whereas they are smaller and much more discontinuouss where related to the F_2 folds within the metavolcanics. In the latter, these brittle-ductile D_2 high strain zones are mostly present in the hinge of the folds and are inferred to have developed as accommodation structures during fold growth and tightening.

From a genetic point of view, the presence of fluorite near, and rarely on the margins of, the mineralized veins led Huard (1990) to propose a magmatic source linking the felsic dykes to the gold-bearing hydrothermal fluids. The structural control of the dykes on the mineralization is obvious and the timing relationship between the gold hosting structures and the mineralization indicates that the dykes are older than the mineralization as they are cut by the D_2 high strain zone. But how much older are they? Neither can a magmatic source for the fluid be ruled out nor can a metamorphic fluid system. Ongoing U/Pb geochronological dating of the felsic dykes and of appropriate hydrothermal minerals by D. Ritcey at Memorial University may give a definite answer on the timing relationship between the dykes and the mineralization and will greatly help to evaluate the potential for a felsic magmatic source to the mineralizing fluids.

This paper has presented the different structural elements and their relationship with the mineralization. But, work has still to be done on the overall distribution of the units and the geometry of the veins. The reason for the two different orientations of the axial surface of the F_3 folds is unknown. It is proposed that they represent conjugate sets of late F_3 folds but this hypothesis has to be tested. From an exploration point of view, the recognition of the F_3 folds is critical in the definition of the geometry of the ore zone and localization of ore shoots because they are folded. This could possibly explain the apparent lack of down-dip extension of the mineralization with depth reported by Huard (1990). It is not clear what the distribution of the units is at the property scale. It is also critical to understand why the veins are rotated from 060° on Muddy Shag to "E-W" on Hammer Down and Rumbullion and back again at the N060° trend in a showing located further east. Is this variation in trend produced by a flexure in the shear direction creating dilation and allowing vein formation on Hammer Down as proposed by Huard (1990) or is it related to the F_3 folds, or again are we looking at two differently oriented shears? The relationship with the Green Bay fault is also unknown mainly because of the lack of exposure of this fault, but it is geometrically possible that the F_2 folds, and the related mineralization, are controlled by strike slip movement along the Green Bay fault. It also possible, as proposed by Andrews (1990) and by Huard (1990), that the NE trending "shear zones" of the Muddy Shag showing are in splays of the Green Bay Fault.

Ongoing work leading to two M.Sc. theses will lead to better understanding of the distribution of the lithologies and veins and of the structural setting at the property and regional scales (D. Gaboury, Laval University) as well as the alteration and stable and radiogenic isotopic characteristics of the deposit (D. Ritcey, Memorial University).

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Quaternary geological studies, Buchans area of central Newfoundland[1]

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Abstract

Quaternary geological studies within the Buchans (12A/15) and Badger (12A/16) map areas are designed to establish a geological framework for mineral exploration through integration of mapping, stratigraphy, and compositional analyses of glacial sediments. In this region drift prospecting has been confounded by thick glacial deposits and complex ice flow history. In the western part of the Badger area, the dominant northeastward ice flow was preceded by southward flow. Within the Red Indian Lake basin, near Buchans, regional ice flow trending northeast–southwest was superseded by flows toward the south and the north and northwest. P-forms indicate subglacial meltwater flowing both northeast and southwest. The predominant surficial deposit is till of varying thickness and morphology. It is commonly sandy, contains evidence of internal sorting, and is overlain in places by thin outwash deposits which suggests deposition from an ablating ice sheet. Near Buchans, extensive, subtill glaciofluvial and glaciolacustrine deposits occur to 320 m elevation a.s.l. They overlie mineralized bedrock, and their incorporation within overlying till complicates the compositional relationship between bedrock and surficial sediments.

Résumé

Les études géologiques du Quaternaire, dans les régions cartographiques de Buchans (12A/15) et de Badger (12A/16), ont été conçues de façon à établir un cadre géologique pour la prospection minérale, par intégration de la cartographie, de la stratigraphie, et des analyses de la composition des sédiments glaciaires. Dans cette région, la prospection glacio-sédimentaire (drift) a été rendue très difficile par la présence d'épais sédiments glaciaires et par l'évolution complexe de l'écoulement des glaces. Dans la partie ouest de la région de Badger, l'écoulement dominant des glaces vers le nord-est a été précédé d'un écoulement vers le sud. Dans le bassin du lac Red Indian, près de Buchans, l'écoulement régional des glaces dans une direction générale nord-est-sud-ouest a fait place à des écoulements vers le sud et vers le nord et le nord-ouest. Des formes P indiquent l'écoulement d'eaux de fonte sous-glaciaires à la fois vers le nord-est et vers le sud-ouest. Les dépôts de surface prédominants sont des tills d'épaisseur et de morphologie variables. Ils sont généralement sableux, contiennent les indices d'un triage interne, et sont par endroits recouverts de minces épandages fluvio-glaciaires qui pourraient indiquer la mise en place de matériaux par suite de l'ablation d'un inlandsis. Près de Buchans, de vastes dépôts fluvioglaciaires et glaciolacustres sous-jacents au till apparaissent jusqu'à 320 m d'altitude au-dessus du niveau de la mer. Ils recouvrent un substratum minéralisé, et leur incorporation au till sus-jacent complique la relation caractérisant les compositions du substratum et les sédiments superficiels.

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[1] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-1994

INTRODUCTION

As a contribution to the Canada-Newfoundland Cooperation Agreement 1990-94, studies of Quaternary geology and glacial history in central Newfoundland have been designed to provide a geological basis for drift prospecting. In that region mineral exploration has been confounded by thick glacial deposits and multiple ice flow directions. The work reported here integrates Quaternary mapping with stratigraphic studies and compositional analyses of glacial sediments to define pathways of glacial transport and to establish relationships between surficial sediments and bedrock. The area of study occupies part of the Central Volcanic Belt of Newfoundland, including much of the volcanic and sedimentary terrane of the Buchans Group which has been a focus for mineral exploration (Fig 1). The town of Buchans has long been a major producer of base metal sulphide ore in Newfoundland. Mining operations, however, ceased in 1979 when known ore reserves were exhausted.

During the summer of 1991, mapping of surficial deposits at a scale of 1:50 000 within NTS areas 12 A/15 (Buchans) and 12 A/16 (Badger) was initiated. Throughout both areas samples of glacial sediments, primarily till, were collected for lithological and geochemical analyses. Indicators of ice flow trend and their relative ages were mapped to establish a history of ice flow and to interpret likely directions of glacial transport. This report presents an account of initial field observations and outlines implications for drift prospecting in the area.

PREVIOUS WORK

In central Newfoundland, varied directions and ages of ice flow have been defined based on glacial striations and streamlined landforms, and on crosscutting relationships among them. The geological record, however, has proven to be complex, and conclusions regarding glacial history have been inconsistent and contradictory. On the Glacial Map of Canada (Prest et al., 1969) landforms define prominent ice flow towards the northeast and the southwest across central Newfoundland, and some few striations are shown that indicate other directions. Similar trends are evident on radar images, along with prominant northwest-southeast trends across the Buchans area (Graham and Grant, 1991). Among earlier workers, the most complete record of glacial striations is that of Murray (1955) who described ice flow south of Red Indian Lake based on his own observations and those of exploration geologists. Based on the distribution of granitic erratics, he defined the principal direction of glacial transport toward the northeast.

For the Buchans map area (12 A/15), (Sparkes, 1987) proposed: 1) early radiating ice flow toward the south (140°–160°) from a centre located north of Red Indian Lake; 2) northeastward ice flow (020°–060°), possibly resulting from change in the location of an ice divide and thinning of the ice sheet; 3) formation of a glacial lake within Red Indian Lake basin, likely representing either interstadial or interglacial conditions; and 4) southwestward ice flow into Red Indian Lake from a centre located either north or northeast of Buchans townsite.

Within the Badger map area (12 A/16) most striations indicate ice flow toward the northeast (050°-075°), which is consistent with the observations of Murray (1955), and more northerly in the eastern part of the area (Sparkes and Vanderveer, 1980; St. Croix and Taylor, 1990). In its western part, near Exploits River, and across map areas to the south (12 A/10, 12 A/9) and west (12 A/15), there is widespread evidence for older ice flow toward the south (175°-195°) (Vanderveer and Sparkes, 1979; Sparkes and Vanderveer, 1980), which is likely related to the early southward flow described by Sparkes (1987) in the Buchans area.

Based on ice marginal landforms and glaciofluvial deposits, a deglacial history of central Newfoundland has been proposed that indicates general recession toward an ice divide arcing across the central part of the island, and its later devolution into separate, shrinking ice caps, one of which was last located within Red Indian Lake basin (Grant, 1974). The landforms define northeastward retreat of an ice lobe within the lake basin, and southward retreat of the ice margin across Hinds Lake and eastward toward the Topsails Plateau (Grant, 1975), in apparent contradiction to the final conclusion of Sparkes (1987). According to Grant, final disintegration of the ice was within Red Indian Lake basin. That general model is supported by the conclusions of (St. Croix and Taylor, 1991) derived from striation mapping east of Badger.

Stratigraphically and compositionally distinct tills have been reported in the areas of Buchans and the southwestern part of Red Indian Lake basin, along with intertill waterlain sediments of possible interglacial or interstadial significance (Grant and Tucker, 1976; Sparkes and Vanderveer, 1980; Vanderveer and Sparkes, 1982; Sparkes 1984, 1985; Mihychuk 1985).

Drift prospecting activities near Buchans have been summarized by James and Perkins (1981) who identified three ice flow directions including "... a dominant ice movement from the northeast, a prominent movement from the northwest, and an obscure movement from the west–southwest" (p. 281). Ice flow toward the southwest is shown by a glacial dispersal train defined by geochemical analyses that extends more than 8 km from the principal mine sites. Mineralized erratics of Buchans–type ore also occur east and northeast of Buchans and, although their origins have not been established, they are possibly related to glacial transport in the prominent northeast flow direction described by James and Perkins.

GEOLOGY

The area is underlain by sedimentary, volcanic, and intrusive rocks of Paleozoic age comprising part of the Newfoundland Central Volcanic Belt (Kean et al., 1981; Colman–Sadd et al., 1990). North of Red Indian Lake, volcanic and sedimentary rocks of the Buchans and Roberts Arm Groups underlie the central part of the Buchans map area and extend to the northeast across the Badger map area (e.g., Kirkham, 1987) (Fig. 1). To the north and northeast is the Topsails igneous terrane, which includes the peralkaline and granitic rock of the Topsails Intrusive Suite, volcanic and sedimentary rocks of the Springdale

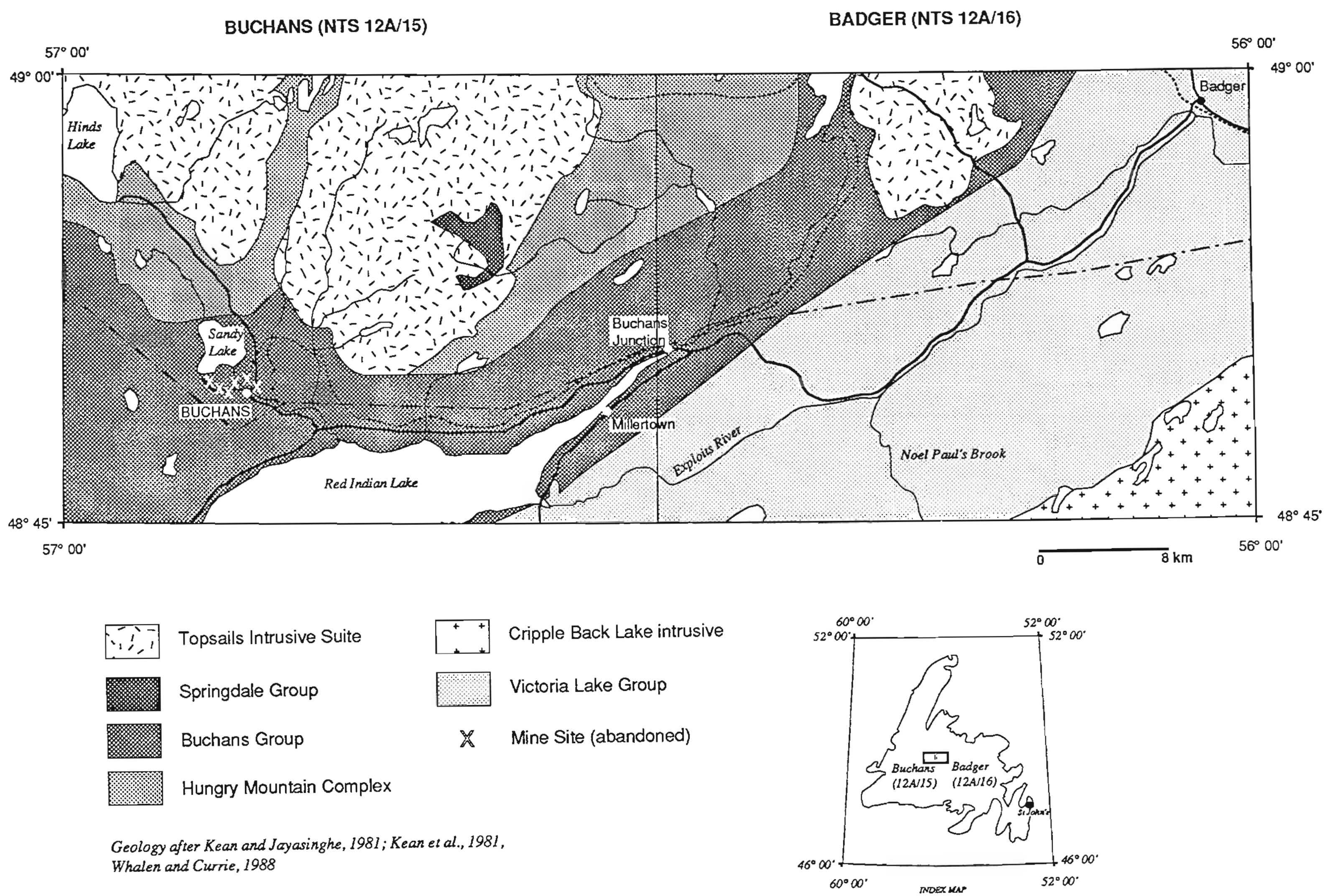

Figure 1. Bedrock geology of the study area.

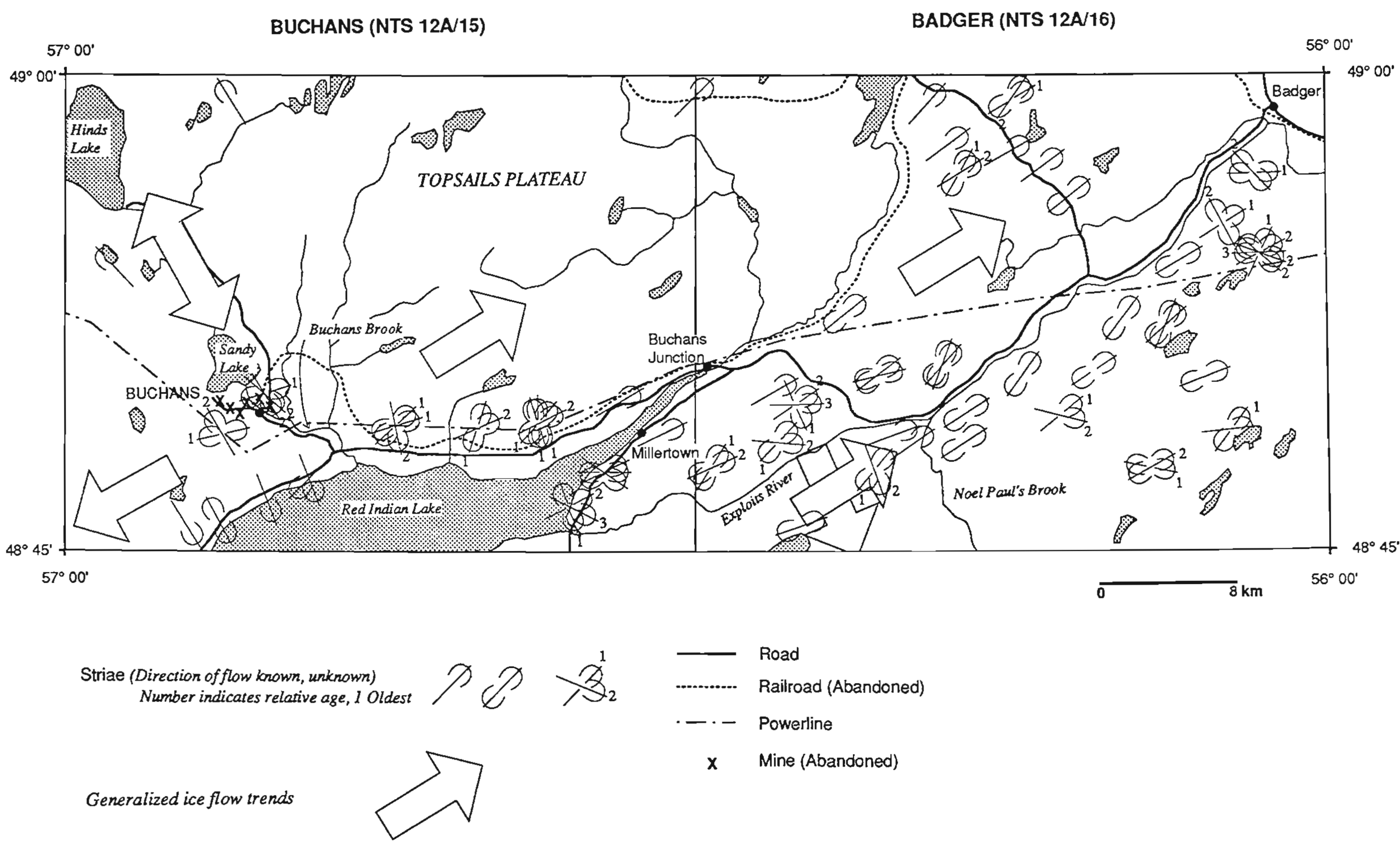

Figure 2. Map of the study area showing the trends of striations and their relative ages.

Group, and intrusive and metamorphic rock of gabbroic to granitic composition of the Hungry Mountain Complex (Kean and Mercer, 1977; Whalen and Currie, 1987, 1988).

South of Red Indian Lake and across most of the Badger map area, volcanic and sedimentary bedrock is part of the Victoria Lake Group. The southeastern corner of the Badger map area is underlain by the Cripple Back Lake quartz monzonite (Kean and Jayasinghe, 1980).

The basin occupied by Red Indian Lake is part of a major northeast trending fault zone extending across much of central Newfoundland (Kirkham, 1987). Within the basin, southwest of the map areas, inliers of Carboniferous conglomerate, sandstone, shale, and siltstone are present.

PHYSIOGRAPHY

Central Newfoundland is characterized by broad hills having relief of tens to more than a hundred metres and they are generally elongate in a northeast–southwest trend. Elevations vary between 150 and 500 m a.s.l., but are greatest over Topsails Plateau, which is part of the Long Range Mountains and is underlain by bedrock of the Topsails igneous terrane (Fig. 2). Red Indian Lake occupies a structural lineament crossing the central part of the Buchans map area, and is drained by Exploits River which flows northeast across the Badger map area. The townsite of Buchans lies within a northwest trending saddle occupied by Hinds and Sandy lakes.

FIELD METHODS

Access to the field area was by truck along regional highways, and by all–terrain vehicle along logging roads, abandoned rail lines, and powerlines. Logging roads permitted access to most of the Badger map area, although in its western part they are poorly maintained and, as a result, difficult to travel. In contrast, there has been little logging within the Buchans area. Roads there are associated primarily with mining operations and are largely restricted to a narrow corridor along Red Indian Lake and the valley north of Buchans; elsewhere access was by helicopter. Canoe traverse along the shores of Red Indian and Hinds lakes permitted examination of shoreline exposures.

Surficial sediments were mapped through examination of natural and man-made exposures along roads and streams, hand-dug pits and, near Buchans, backhoe pits. Striated surfaces, where exposed, were examined to determine trend, sense, and relative age of ice flow. More than four hundred samples were collected for geochemical, lithological, and mineralogical analyses.

STRIATIONS AND OTHER ICE FLOW FEATURES

Throughout both map areas striations and streamlined landforms are widespread. Most trend 020° to 060°, indicating regional ice flow generally toward the northeast (Fig. 2). Within Red Indian Lake basin, including the southwestern part of the Badger map area, an older phase of ice flow toward the south and southeast is defined by striations on outcrop surfaces 'sheltered' from erosion by later, northeastward flowing ice. Within the valley occupied by Sandy and Hinds lakes, north of Buchans, the last record of ice flow appears contradictory, indicating movements both toward the southeast and the northwest.

Along the shore of Red Indian Lake, and at a few sites within the Badger map area, P-forms occur, including furrows, musselbruch, and sichelwannen (e.g., Kor et al.,1991). They indicate flow toward the northeast, defining what may be the result of subglacial meltwater flow within the area during late glacial time. Southwest of the Buchans map area, P-forms indicate flow toward the southwest.

GLACIAL LANDFORMS

Glacially streamlined landforms, including flutes and crag-and-tail hills, are few and throughout much of the map area are aligned with the general northeast trend of bedrock structure (Grant, 1975; Sparkes, 1985; Graham and Grant, 1991) making them difficult to distinguish on aerial photgraphs. As previously noted by Grant (1975), north and east of Buchans, ice marginal meltwater channels outline the southward retreat of the ice margin across Hinds Lake toward Red Indian Lake, as well as eastward up valleys toward highland areas of the Topsails Plateau. From their form, late glacial ice flow was strongly controlled by topography.

SURFICIAL GEOLOGY

Till is the predominant surficial sediment. Throughout much of the area it is commonly a grey to brown sandy diamicton characterized by internal sorting in the form of thin, discontinuous lenses and pockets of sorted sediment; silt caps on clasts; and sandy linings to clast moulds (Fig. 3). The evidence is consistent with deposition from a disintegrating ice sheet. Within the basin occupied by Red Indian Lake, till is a red to brown, muddy sand diamicton having little or no internal stratification and a compact nature, and it is interpreted as a lodgment till. Its color lies chiefly within the finer fraction, and reflects bedrock provenance. The red is derived from glacial erosion of red sedimentary rocks of several formations, including red sandstone of the Springdale Group, red siltstone and argillite of the Buchans Group, and red conglomerate of carboniferous strata.

Till is generally thickest near Red Indian Lake and overlying Buchans Group terrane north of the Lake. It is thin (<2 m) to discontinuous over topographic highs such as the Topsails Plateau, and along parts of the northern margin of Exploits River valley. Elsewhere, it is typically more than 2 m thick, and it forms a broad plain with minor hummocky areas localized in depressions. Along the sides of most valleys till is eroded and dissected by numerous, closely spaced gullies or channels trending into the valley, and is locally overlain by thin outwash deposits. Areas of ribbed moraine are

Figure 3. Till section showing its stony, poorly stratified character. (GSC 1991 572A)

present in valleys or broad lows within upland areas, and they are best developed in the southwest quadrant of the Badger map area.

Ice contact glaciofluvial deposits are well developed along the southern margin of the Badger map area, east of Noel Paul's Brook. There, sinuous, sharp-crested eskers up to 20 m high with segments more than 2 km long occur in association with glaciofluvial outwash sediments. They consist of well sorted, well stratified sands and gravels. The eskers trend approximately east northeast-west southwest, and structures within them indicate deposition by water flowing eastward. Shorter, more poorly developed esker segments occur elsewhere, primarily in the northern part of the Badger map area. Within Buchans Brook and its tributaries, moderately to well sorted deposits of sand and sandy gravel form a series of ice contact deposits, including kames, kame terraces, and deltas graded to elevations between 310 and 270 m. The deposits have not been glacially overidden and are thereby considered younger than extensive subtill waterlain deposits seen in stratigraphic sections exposed in gravel pits southeast of Sandy Lake.

Glaciofluvial outwash sediments are characterized generally by angularity of clasts, poor sorting, and by marked facies changes over short distance. The evidence indicates relatively little transport and modification by meltwater, and deposition could in part be subglacial. The most extensive and thickest deposits occur within the valley occupied by Sandy and Hinds lakes, near the shore of Red Indian Lake, and along Exploits River valley where they occur at elevations 10-20 m above the modern river level. Shoreline exposures along Hinds Lake reveal deposits of cobble to boulder gravel having moderately to well rounded clasts, and that degree of rounding is notable within the Buchans area. The deposits are most likely associated with northward flow of meltwater indicated by the slope of meltwater channels across hill sides overlooking the lake.

STRATIGRAPHY

As noted by Sparkes (1985), stratigraphic sections near Buchans include subtill glaciofluvial and glaciolacustrine sediments, the thickest accumulations of which are exposed in gravel pits south and east of Sandy Lake. There, the sediments are over 30 m thick and coarsen upward from fine sand and silt, to sand, sand and gravel, and cobble gravel. They are capped by a sandy till containing large (1-2 m) striated erratics derived from the Topsails igneous terrane to the northeast as well as rounded clasts from the underlying glaciofluvial gravels. Despite having been overidden by glacier ice and capped by till derived from the outwash, it is evident on aerial photographs that the waterlain sediments occur as a deltaic landform graded to about 260 m elevation.

At several locations near Buchans, and up to 10 km southwest, stratified sediments occur beneath till, and can be incorporated within till as deformed inclusions. The sediments are finely laminated to thin bedded, comprising well sorted, fine to medium grained sand and silt. Beds are graded, fining upwards, and at some sites can include minor, discontinuous beds of pebbles or coarse sand (Fig. 4). The deposits occur to elevations of about 320 m a.s.l., and the thickest known accumulations occur along the south shore of Sandy Lake. At least part of the subcropping mineralization at Buchans is directly overlain by sand and silt. The sediments are interpreted as glaciolacustrine, and are inferred to result from deposition within a lake dammed by glacier ice within Red Indian Lake basin. The location of the presumed ice dam and extent of the glacial lake have not been defined.

On the north shore of Red Indian Lake, till contains deformed, angular inclusions of finely laminated, grey-brown sediments with sedimentary structures intact. Near the Exploits dam similar waterlain sediments occur along shear planes separating large blocks of carboniferous conglomerate that have been thrust stacked by glacier ice. Stratification within the sediments is preserved, although defining chevron folds. The waterlain sediments are several centimetres to 40 cm thick, and the thrust blocks are several metres thick. The thrust planes generally strike southeast and dip about 30° southwest, and the blocks are thus presumed to have been emplaced by ice flowing generally toward the northeast.

DISCUSSION, PRELIMINARY CONCLUSIONS, AND IMPLICATIONS FOR MINERAL EXPLORATION

Striations

As reported by previous workers, crosscutting striations on bedrock surfaces define several directions of ice flow having different relative ages. Perhaps the most complex record lies within Red Indian Lake basin, particularly near Buchans where flows to the southwest, south, northeast, and northwest are recorded. Striations can be difficult to interpret fully in terms of glacial history and ice sheet reconstruction. The older record is commonly incomplete due to glacial erosion during suceeding events. Although the youngest flow trends are predominant and well defined, they are not necessarily the same age throughout a map area, and cannot be used collectively to define the shape of the ice sheet at any one time. Reconstruction of glacial history requires compositional evidence that can be used to illustrate the relations among striations, glacial dispersal trends, and distances of glacial transport from specific bedrock sources.

Figure 4. Finely laminated to thin bedded silt and sand form subtill glaciolacustrine deposits that are widespread in the Buchans area. (GSC 1991 572F)

The following are impressions derived from striations measured during the present study:

1) Contradictory estimates of relative age among striations having similar trends indicate that ice likely moved in the same direction on more than one occasion.
2) Based on the occurrence of P-forms, late glacial movement within the basin of Red Indian Lake may have been associated with the flow of subglacial meltwater along the axis of the lake, outwards to the northeast and southwest.
3) Near Buchans, there is widespread evidence for last flow toward the south and southeast, although at two sites striations also indicate ice flow toward the northwest. The northwest ice flow directions are consistent with the slope of the ice surface defined by ice marginal landforms along hillsides adjacent to Hinds Lake and are thus presumed late glacial.
4) Although striations within the Badger map area are generally northeast, they are more complex within Exploits River valley, particularly in the eastern part of the map area. On one outcrop near Badger the multiple directions of flow include trends that are not recognized elsewhere in the area. They may represent local reorientation of ice flow towards the valley during thinning and disintegration of the ice sheet.

Sediments

Subtill rhythmites interpreted as glaciolacustrine have been reported in the Tulks and Lloyds river valleys, southwest of Red Indian Lake (Vanderveer and Sparkes,1982). Near Victoria River, outside the study area south of Buchans, a glacial lake shoreline occurs at an elevation of about 211 m, which is 60 m above the surface of Red Indian Lake and is similar to the elevation of glaciolacustrine deposits to the southwest described by Vanderveer and Sparkes (Mihychuk 1985). The glacial lake has been informally named 'Glacial Lake Shanadithit' by Mihychuk. Subtill waterlain sediments near Buchans occur between 260 and 320 m a.s.l. and their greater elevation indicates that they could represent either an earlier stage in the evolution of that glacial lake or a separate event. Their location indicates that glacial lake drainage could have been northward out of Hinds Lake and that substantial amounts of ice remained within Red Indian Lake blocking flow either to the southwest or to the southeast via Exploits River.

The recognition of subtill waterlain sediments has some clear implications for drift prospecting, especially in the vicinity of Buchans. Although a great deal of exploration has occurred there, geochemical analyses results have not been interpreted in the context of a detailed stratigraphic framework. From work reported here, the surface till is partly derived from the waterlain sediments by glacial erosion, and thus the compositional relationship between surficial deposits and bedrock cannot be explained by a simple model of glacial erosion and transport in one ice flow direction.

Based on geochemical analyses of surficial sediments, a prominent glacial dispersal train characterized by 'anomalous' zinc concentrations extends more than 8 km southwest of the mine site in the direction of principal ice flow (James and Perkins, 1981). From the work of this study, it is conceivable that the anomalous results could also reflect transport of fine grained ore constituents within a glacial lake system. James and Perkins did note a broad accordance between the dispersal pattern and low topography which is consistent with the expected distribution of glacial lake sediments. Testing of that hypothesis requires geochemical analyses of the glacial lake sediments and thorough examination of glacial stratigraphy within the area of the dispersal train. The preservation of glaciolacustrine sediment overlying subcropping mineralization at Lucky Strike Mine indicates that bedrock there was not subject to glacial erosion during the last ice flow event, and possibly earlier ones as well.

East of Buchans the surface till incorporates abundant red granitic debris from bedrock to the northeast, as well as rounded clasts from the underlying deposits of glaciofluvial sands and gravels overidden by the ice. The glaciofluvial sediments appear to have masked underlying bedrock, limiting its glacial erosion. Within one of the open pits (Oriental) east of Buchans, an older till in contact with bedrock is reported to contain mineralized debris in contrast with the surface till directly overlying it that is effectively barren of such debris (Sparkes, 1987).

Interpretation of aerial photographs and examination of till exposures indicate that surficial sediments have been subjected to reworking or remobilization by meltwater flowing into the valleys, leading to varying degrees of sorting and sediment transport. The effects of such remobilization on drift composition and glacial dispersal patterns is unknown. The implication is that the sedimentology of surficial deposits can be important to drift prospecting, and especially so at the detailed scales of investigation that commonly characterize mineral exploration.

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Revision of upper Precambrian-Cambrian stratigraphy, southeastern Cape Breton Island, Nova Scotia[1]

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Abstract

Recent remapping of uppermost Precambrian and overlying Cambrian rocks in southeastern Cape Breton Island has resulted in preliminary revision of stratigraphic units and their distribution. East and north of the Mira River, the ca. 560 Ma Main-à-Dieu sequence is overlain by uppermost Precambrian to lowermost Cambrian redbeds, lower Cambrian quartz arenite and shale, and various Middle and Upper Cambrian units. West of the Mira River, a different upper Precambrian to Lower Cambrian unit (Kelvin Glen Group) is recognized. Stratigraphic sequences in the Boisdale Peninsula begin with Middle Cambrian units. Cambrian volcanic rocks are present only in the central Boisdale Peninsula.

Résumé

Après avoir récemment recartographié le Précambrien sommital et les roches cambriennes sus-jacentes dans le sud-est de l'île du Cap-Breton, on a procédé à une révision préliminaire des unités stratigraphiques et de leur répartition. À l'est et au nord de la rivière Mira, la séquence de Main-à-Dieu, datée d'environ 560 Ma, est recouverte par des couches rouges qui s'échelonnent de la partie supérieure du Précambrien à la partie inférieure du Cambrien, un quartzite sédimentaire et un shale du Cambrien inférieur, et diverses unités du Cambrien moyen et supérieur. On a identifié, à l'ouest de la rivière Mira, une unité différente s'échelonnant du Précambrien supérieur au Cambrien inférieur (groupe de Kelvin Glen). Les séquences stratigraphiques de la péninsule de Boisdale commencent par des unités du Cambrien moyen. Il n'existe de roches volcaniques d'âge cambrien que dans la région centrale de la péninsule de Boisdale.

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INTRODUCTION

Uppermost Precambrian and overlying Cambrian rocks in southeastern Cape Breton Island have been remapped as part of a regional project involving mapping, petrological studies, and geochronology (Barr et al., 1988, 1989, 1990a, b, 1991; Barr and White, 1989). This report summarizes preliminary results from this project for the Mira River area (Fig. 1), and indicates how these results differ from earlier interpretations (Fig. 2). In addition, some preliminary results are described for two areas underlain by possibly correlative rock units in the Boisdale Peninsula, located northwest of the Mira River area (Fig. 1, inset map).

BACKGROUND

Hutchinson (1952) and Weeks (1954) subdivided the pre-Carboniferous sedimentary and volcanic rocks in the Mira River area of southeastern Cape Breton Island into the Upper Precambrian Fourchu Group and various, mainly Cambrian, overlying groups and formations (Fig. 2). Smith (1978) modified the units of these earlier workers for the area southwest of the Mira River. He introduced new names for Cambrian units and suggested that a large part of the area is underlain by Upper Precambrian rocks, probably equivalent to the Fourchu Group, that he termed the Giant Lake Complex (Fig. 2). This interpretation was subsequently followed by Keppie (1979).

Recently, Landing (1991) re-examined Upper Precambrian-Lower Cambrian stratigraphy in southeastern Cape Breton Island, and extended the use of unit names of eastern Newfoundland to their inferred correlative units in southeastern Cape Breton Island (Fig. 2). However, Landing (1991) did not attempt to modify the pre-existing geological maps of Hutchinson (1952), Weeks (1954), and Smith (1978).

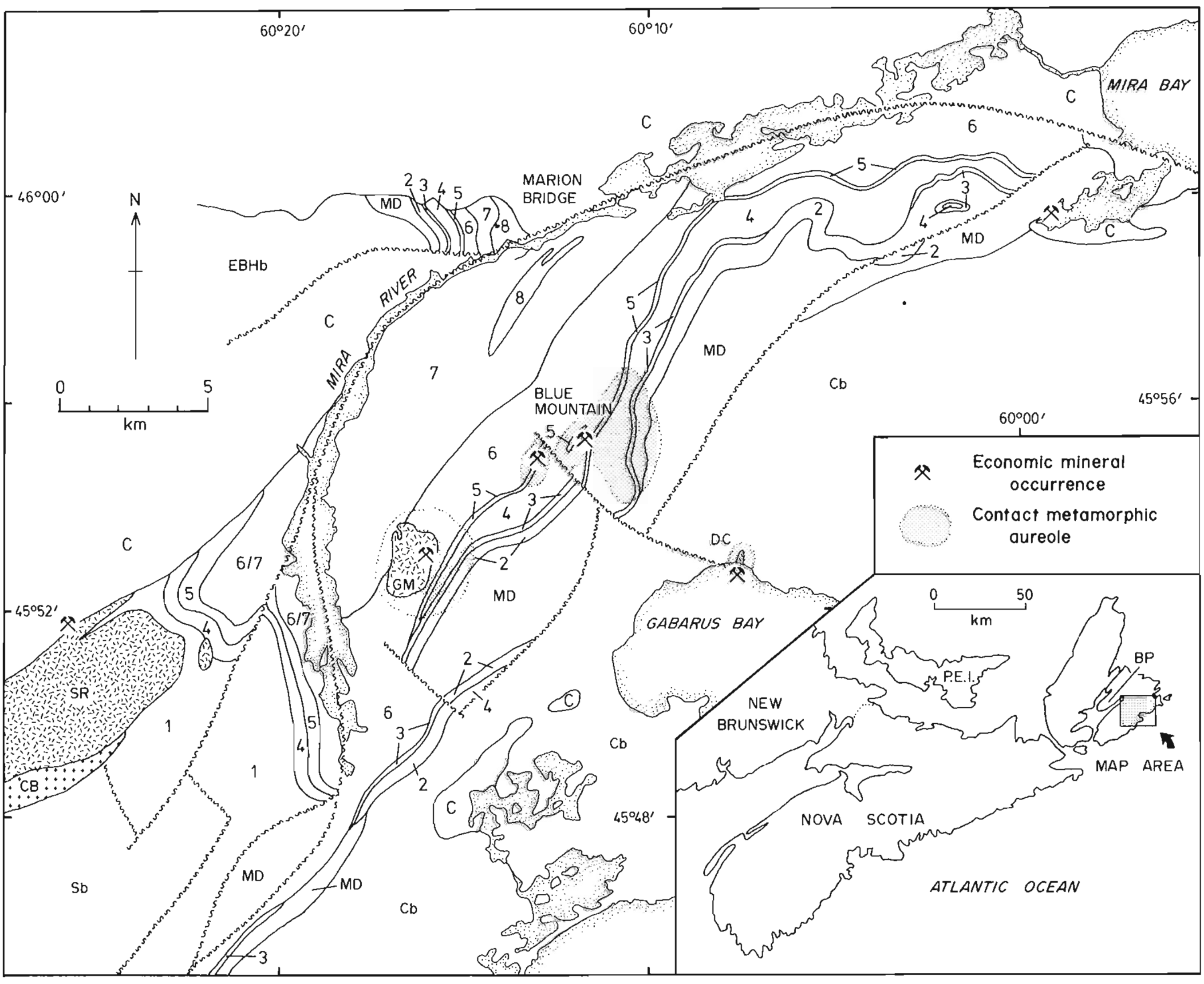

Figure 1. Simplified geological map of the Mira River area showing the distribution of major map units. Inset map shows the location of the Mira River area and the Boisdale Peninsula (BP) in southeastern Cape Breton Island.

Cambrian rocks of the Boisdale Peninsula were first mapped by Bell and Goranson (1938) but were not subdivided until the work of Hutchinson (1952). Middle Cambrian volcanic and sedimentary rocks of the Bourinot Group and overlying Upper Cambrian to Lower Ordovician units were interpreted to unconformably overlie Proterozoic rocks and to have been intruded by younger granitic intrusions.

MAP UNITS EAST OF MIRA RIVER

Our mapping east of the Mira River (Fig. 1) has resulted in modification of the areal distribution of units from that shown by Hutchinson (1952), Weeks (1954), and Landing (1991), in part because additional outcrops are now available in the area. An important difference from the earlier maps is the recognition of a sequence of volcanic and sedimentary rocks (Main-à-Dieu sequence) that is distinct, at least in rock types and deformational history, from the Coastal belt of the Fourchu Group (Barr and White, 1989; Barr et al. 1990a, 1991). The upper part of the Main-a-Dieu sequence is characterized by laminated siliceous siltstones that vary from red or maroon to grey or grey-green, and by maroon and green-grey volcanogenic conglomerate and sandstone, locally interlayered with amygdaloidal basalt, rhyolite, and felsic to intermediate lithic tuff (Barr and White, 1989).

A dominantly red clastic sedimentary unit (unit 2, Fig. 1) overlies the Main-à-Dieu sequence. Contacts are mainly faulted or unexposed, but a disconformable relation seems likely. The redbed unit has a distinctive quartzite/ quartz-pebble conglomerate at or near its base, and is interlayered with, and overlain by, red sandstone and siltstone, commonly with well developed crossbedding and graded bedding. Landing (1991) correlated this redbed unit with the Rencontre Formation of Newfoundland, and suggested that the name should also be used in Cape Breton Island. These clastic sedimentary rocks differ from those of the underlying Main-à-Dieu sequence in that they contain abundant detrital muscovite, and lack associated volcanic and pyroclastic rocks.

The redbed unit is overlain by a distinctive white-weathered, rarely maroon, crossbedded quartz arenite with local conglomeratic (quartz pebble) lenses (unit 3, Fig. 1). This unit has been correlated with the Random Formation of Newfoundland (Landing, 1991). Landing (1991) recognized a separate unit of shales between the redbed and quartz arenite units that he correlated with the Chapel Island Formation of Newfoundland. We observed minor green-grey shales and siltstones within the redbed unit, but these rocks to not appear to form a mappable unit separate from the redbed unit, and hence we question the presence of the Chapel Island Formation as a major unit in Cape Breton Island.

The quartz arenite is overlain by grey and green siltstone and shale that we assign to the Lower Cambrian MacCodrum Formation (Fig. 1, 2), following Hutchinson (1952). On the basis of comparison with Newfoundland, Landing (1991) assigned these rocks to the Bonavista Group. He suggested that an unconformity is present between the shales and the underlying quartz arenite because the contact between the two units is sharp in most places. However, we observed interlayered quartz arenite and shale at two localities and hence, like Hutchinson (1952), consider the relationship between these units to be conformable.

The Canoe Brook Formation overlies the MacCodrum Formation. It consists of red-brown, carbonate-rich mudstone and siltstone, maroon siltstone containing grey-green reduction spots, and minor pink to red limestone. The limestone may be locally stromatolitic (E. Landing, pers. comm., 1991), although Landing (1991) assigned the stromatolitic limestone to the underlying MacCodrum/Bonavista Formation. Our definition of the Canoe Brook Formation does not differ significantly from the original description by Hutchinson (1952), other than the recognition of the presence of carbonate material. Our mapping shows that the mineralized skarns in the Blue Mountain area (Macdonald, 1989; Macdonald and Barr, in press) are developed in these carbonate-rich rocks of the Canoe Brook Formation. According to Landing (1991), the

LEGEND

CARBONIFEROUS	
C	Horton and Windsor groups: red conglomerate, sandstone, siltstone, and limestone
DEVONIAN	
	Plutonic rocks: Gillis Mountain (GM) granite, monzodiorite; Deep Cove (DC) granite; Salmon River (SR) rhyolite porphyry
CAMBRIAN	
Late Cambrian	
8	MacNeil Formation: dark grey shale, siltstone, and black limestone
Middle Cambrian	
7	MacLean Brook Formation: grey quartz sandstone, siltstone, and shale; minor light grey quartz sandstone
6	Trout Brook Formation: dark grey to rust brown cleaved shale and siltstone
Early Cambrian	
5	Canoe Brook Formation: red-brown carbonate-rich mudstone and siltstone ± grey reduction spots; rare pink limestone east of Mira River
4	MacCodrum Formation: light grey to green, laminated to thin-bedded micaceous siltstone and sandstone
3	White to maroon quartz arenite; locally black due to contact metamorphism
LATE PRECAMBRIAN TO EARLY CAMBRIAN	
2	Maroon to red quartzite/quartz-pebble conglomerate; maroon to red medium- to coarse-grained micaceous sandstone and minor red siltstone
1	Kelvin Glen Group: pale orange arkosic sandstone, pebble to cobble conglomerate, purple to grey wacke ± micaceous red sandstone
MD	Main-a-Dieu sequence: volcanic and sedimentary rocks, including basalt, rhyolite, laminated siltstone and chert, red sandstone, siltstone, conglomerate, varied tuffaceous rocks
LATE PRECAMBRIAN	
CB	Chisholm Brook granodiorite
Cb	Coastal belt volcanic and volcaniclastic rocks
EBHb	East Bay Hills belt volcanic and volcaniclastic rocks
Sb	Stirling belt volcanic, volcaniclastic, and sedimentary rocks

Figure 1 Legend

Canoe Brook Formation, as well as the upper part of the underlying MacCodrum Formation, are equivalent to the Brigus Formation in Newfoundland (Fig. 2).

The lower Middle Cambrian Trout Brook Formation (Hutchinson, 1952) is a sequence of dominantly dark grey to rust-brown, well cleaved shale and siltstone that overlies the Canoe Brook Formation. Toward the stratigraphic top of the Trout Brook Formation, the shales become locally maroon, with thin, graded beds of fine grained sandstone.

The fossiliferous Middle Cambrian MacLean Brook Formation (Hutchinson, 1952) overlies the Trout Brook Formation, and consists of interbedded grey quartz sandstone, siltstone and shale with minor light grey quartz sandstone and maroon shale. The MacLean Brook Formation appears to be conformable with the underlying shales of the Trout Brook Formation. Cleavage is poorly developed and sedimentary structures are well preserved. The sandstone is typically crosslaminated whereas the siltstone and shale exhibit parallel to wavy laminations, with local ripple marks. Flute and load marks as well as trace fossils are abundant.

Dark grey shale, siltstone, and limestone of the Upper Cambrian MacNeil Formation (Hutchinson, 1952) overlie the MacLean Brook Formation. The dark grey to black limestone is the most characteristic rock type in the MacNeil Formation. It occurs in layers and concretions, and typically displays cone-in-cone structure.

MAP UNITS NORTH OF THE MIRA RIVER

A small area of Upper Precambrian to Cambrian rocks is preserved at the eastern tip of the East Bay Hills near Marion Bridge (Fig. 1). Stratigraphy in this area is very similar to that east of the Mira River. Flow-banded rhyolite, rhyolitic tuff, and interlayered laminated red-maroon siltstone, probably correlative with the Main-à-Dieu sequence, are overlain by the redbed unit, with the distinctive quartzite/quartz-pebble conglomerate at the base. The redbed unit is overlain by the white-weathered quartz arenite unit, in turn overlain by the *Psammichnites*-bearing MacCodrum Formation. The latter unit is well exposed in MacCodrum Brook. Overlying Cambrian formations are not well exposed, except the MacLean Brook and MacNeil formations near Marion Bridge. The MacNeil Formation in this area contains characteristic black limestone beds.

MAP UNITS WEST OF THE MIRA RIVER

Differences between rock units west and east of the Mira River were recognized by Weeks (1954), who assigned much of the area to the Middle Cambrian "Bourinot Group" (Fig. 2). The "Bourinot Group" in this area is now known to be Precambrian in age (ca. 677 Ma, and intruded by ca. 620 Ma granodiorite and diorite; Barr et al. 1990b, 1991; Keppie et al., 1990), as suggested by Smith (1978) who termed it the "Giant Lake Complex" (Fig. 2). It is here referred to informally as the Stirling belt, to distinguish it from younger (ca. 574 Ma) volcanic units of the Coastal belt (Fig. 1, 2).

The Precambrian volcanic and sedimentary rocks of the Stirling belt are overlain by arkosic sandstone, conglomerate, and wacke, here assigned to the Kelvin Glen Group (Fig. 2). Many of the contacts appear to be faulted. The Kelvin Glen Group, as here defined, corresponds more or less to the Kelvin Lake Formation within the Kelvin Glen Group of Smith (1978), although unlike Smith (1978), we consider that the age may extend into the Late Precambrian (Fig. 2). Based on lithological similarities, we recognize units that appear to be equivalent to the Early to Middle Cambrian MacCodrum, Canoe Brook, Trout Brook, and MacLean Brook formations within the area assigned by Smith (1978) to the Kelvin Glen Group. Hence our definition of the Kelvin Glen Group is more restricted than that of Smith (1978).

		Hutchinson (1952) and Weeks (1954)		Smith (1978)		This study		Landing (1991)
		WEST of MIRA	EAST of MIRA	WEST of MIRA		WEST of MIRA	EAST and NORTH of MIRA	
ORDOVICIAN TO DEVONIAN		Middle River Gp.						
CAMBRIAN	Late		MacNeil Fm.	MacNeil Fm.			MacNeil Fm.	
	Middle	Kelvin Glen Gp.	MacLean Brook Fm.	Kelvin Glen Group	Victoria Brook Fm.	MacLean Brook and Trout Brook Fm. equivalent	MacLean Brook Fm.	
		Bourinot Gp.	Trout Brook Fm.		Gillis Brook Fm.		Trout Brook Fm.	
	Early	Canoe Brook Fm.	Canoe Brook Fm.		Kelvin Lake Formation	Canoe Brook Fm. equivalent	Canoe Brook Fm.	Brigus Fm.
		MacCodrum Fm.	MacCodrum Fm.			MacCodrum Fm. equivalent	MacCodrum Fm.	Bonavista Gp.
		Morrison River Fm.	Morrison River Fm.			Kelvin Glen Group	Quartz arenite	Random Fm.
							Redbeds	Chapel Isl. Fm
								Rencontre Fm.
PRECAMBRIAN		Fourchu Gp.	Fourchu Gp.	Giant Lake Complex		Stirling belt ca. 677 Ma	Main-a-Dieu seq.	Fourchu Gp.
							Coastal belt ca. 574 Ma	

Figure 2. Comparative stratigraphic interpretations in the Mira River area. Sources are cited in the text.

Furthermore, we interpret the rhyolites assigned by Smith (1978) to the "Victoria Brook Formation" to be sills of rhyolite porphyry. The sills are similar in composition to rhyolite porphyry of the Devonian Salmon River intrusion (McMullin, 1984) to the southwest and are probably also of Devonian age. Mafic rocks assigned by Smith (1978) to the basaltic "Gillis Brook Formation" (Fig. 2), although locally amygdaloidal, are more likely to be basaltic sills and dykes, and therefore are of uncertain age.

Hence, contrary to previous studies, our mapping indicates that no volcanic rocks are present in the uppermost Precambrian and Cambrian units west of the Mira River. Direct correlatives of the Main-à-Dieu sequence, the redbed unit, and the quartz arenite unit east of the Mira River do not appear to be present west of the Mira River (Fig. 1, 2). Units in these two areas correlate only after the mid-Early Cambrian, suggesting that the two areas were separate sedimentary basins prior to that time.

BOISDALE PENINSULA

Cambrian rocks occur in the southern and central parts of the Boisdale Peninsula. Mapping in the southern part showed that Cambrian rocks are more extensive than previously recognized, and form a continuous belt instead of occurring in two separate areas as interpreted by previous workers (Bell and Goranson, 1938; Hutchinson, 1952; Thicke, 1987). The oldest unit consists of thinly to thickly bedded, grey to grey-green siltstone with minor light grey silty sandstone and dark grey to rusty orange shale. Grey to brown-grey limestone concretions and beds are locally present. Locally at the base of the unit, a pebble- to boulder-conglomerate, containing clasts of granite and metavolcanic rocks, lies unconformably on the Precambrian Spruce Brook Pluton, although most contacts with volcanic and plutonic rocks appear to be faulted. No evidence for intrusive contacts, as described by Hutchinson (1952), were found during the present study. Hutchinson (1952) correlated these rocks with Middle Cambrian units elsewhere in the Boisdale Peninsula, but lithologically they most resemble the MacLean Brook Formation of the Mira River area. These rocks are overlain by black (locally rust-brown) shale and limestone (MacNeil Formation), in turn overlain by relatively unfossiliferous, quartz-rich siltstone and sandstone that may be Ordovician in age.

Cambrian rocks in the central part of the Boisdale Peninsula occur in a narrow, fault-bounded, northeast-trending belt. Detailed mapping in this belt showed that structurally complex relationships exist, both within the lowermost unit (the Middle Cambrian Bourinot Group of Hutchinson, 1952) and between the Bourinot Group and adjacent Middle Cambrian through Lower Ordovician units. A continuous stratigraphic succession does not appear to be present, contrary to previous interpretations (e.g. Hutchinson, 1952), because contacts between some units are faulted. U-Pb dating of zircon from rhyolite in the Bourinot Group is now in progress, and may resolve relations between the volcanic units and adjacent fossiliferous sedimentary units. The Middle Cambrian volcanic and sedimentary rocks exposed in the central Boisdale Peninsula are lithologically distinct from rocks of similar age elsewhere in southeastern Cape Breton Island, and direct lithological correlation throughout the area is not apparent until the Late Cambrian (MacNeil Formation).

DISCUSSION AND CONCLUSIONS

Uppermost Precambrian to Cambrian rocks in southeastern Cape Breton Island appear to have formed in several separate sedimentary basins. Stratigraphic correlations are limited until the Middle or Late Cambrian. Evidence for Cambrian volcanism is lacking in the Mira River area, and appears to be present only in the central belt of the Boisdale Peninsula. The Precambrian-Cambrian boundary is likely to occur within the Kelvin Glen Group west of the Mira River and within the redbed unit (Rencontre Formation) east and north of the Mira River (Fig. 2). Because of the relatively old age of the underlying rocks of the Stirling belt, as well as the widespread presence of faulted contacts, the lower age limit of the Kelvin Glen Group is not well constrained. The redbed unit, in contrast, overlies the Coastal belt that has a U-Pb (zircon) age of ca. 574 Ma, and the Main-à-Dieu sequence, rhyolite from the lower part of which has yielded a U-Pb (zircon) age of ca. 560 Ma (Barr et al., 1991). The absolute age of the Precambrian-Cambrian boundary must be considerably younger than these dates. It may be better constrained by U-Pb dating, currently in progress, of zircon from felsic volcanic units underlying the redbed unit both east and north of the Mira River.

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A new look at Gander-Dunnage relations in Carmanville map area Newfoundland

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Abstract

The structural top of the Gander Zone (Cuff Pond pelite), lithologically and chemically resembles the base of the Dunnage Zone (Davidsville Group), but is more complexly deformed. Plutonic-metamorphic belts intersecting on the Gander River Ultramafic Belt suggest that thermal activity accompanied and locally outlasted deformation related to ductile shearing separating the Gander and Dunnage zones. Within the Dunnage Zone, the Hamilton Sound sequence can be matched unit by unit with the Exploits Group to the west. Early thrusts probably separate the Hamilton Sound sequence from the Davidsville Group. Disposition of sedimentary sequences and plutonic belts suggest presence of two or more major southerly-directed thrust sheets within the Dunnage Zone, with thermal activity concentrated along frontal and lateral ramps.

Résumé

Le sommet structural de la zone de Gander (pélite de Cuff Pond) ressemble lithologiquement et chimiquement à la base de la zone de Dunnage (groupe de Davidsville), mais est déformée de façon plus complexe. La présence de zones métamorphiques et plutoniques se recoupant dans la zone ultramafique de Gander River porte à croire qu'une activité thermique a accompagné la déformation liée au cisaillement ductile qui sépare les zones de Gander et de Dunnage, et qu'elle s'est poursuivie, par endroits, après cette déformation. À l'intérieur de la zone de Dunnage, la séquence de Hamilton Sound se laisse corréler unité par unité avec le groupe d'Exploits à l'ouest. Les failles de chevauchement plus anciennes séparent probablement la séquence de Hamilton Sound du groupe de Davidsville. La disposition des séquences sédimentaires et celle des zones plutoniques semblent indiquer la présence d'au moins deux grandes nappes de charriage de direction sud dans la zone de Dunnage, l'activité thermique étant concentrée sur le bord de rampes frontales et latérales.

INTRODUCTION

Dunnage and Gander zones of northeastern Newfoundland meet along the Gander River Ultramafic Belt (GRUB). The zonal scheme assumes that such a boundary represents a major suture or transcurrent fault. Kennedy and McGonigal (1972) determined that the rocks below the GRUB exhibited an additional period of deformation compared to those above, and assumed that the more deformed Gander Group was significantly older than Davidsville Group, and separated from it by a period of orogeny. Currie et al. (1980) showed that the GRUB was not continuous, and concluded that lithologies above and below the GRUB line were transitional to one another where the GRUB was absent. They suggested that the structural relations emphasized by Kennedy and McGonigal were an effect of emplacement of the GRUB, and did not represent a significant time interval. Both points of view have been expressed in recent work. Williams et al. (1991) emphasized the presence of mylonite along the Gander-Davidsville boundary, and suggested a significant time break between Gander and Davidsville deposition. van Staal and Williams (1991), based mainly on New Brunswick examples, but applied to the Indian Bay Big Pond Formation of the Gander Group, considered Davidsville equivalent to have conformably overlain Gander Lake equivalent in its original configuration and therefore denied a significant time gap. Recent LITHOPROBE seismic reflection work across the GRUB, together with earlier geophysical work (Miller, 1988) suggest that the Gander-Dunnage boundary does not form a seismically or gravitationally detectable feature extending to depth. The Gander-Davidsville-GRUB problem must therefore be attacked mainly by surface geological methods.

In any discussion of Gander-Davidsville-GRUB relations, the Carmanville map-area (2E/8) is of critical importance because (a) outcrop is more abundant than farther south, (b) the GRUB is discontinuous, permitting observation of relations where the GRUB is present and where it is absent, and (c) metamorphic isograds and associated granitic plutons overprint the Gander-Davidsville boundary in this region, giving other parameters to characterise the contact. For these reasons a multi-disciplinary group including H. Williams and D.H.Johnston, Memorial University of Newfoundland, and M.A.J.Piasecki, University of Keele has undertaken a re-examination of the Carmanville map-area.

DESCRIPTION OF UNITS

The geology of the Carmanville sheet is shown in Figure 1. In general terms the rocks can be divided into five "packages", which from south to north are (1) Gander Group, (2) GRUB, (3) Davidsville Group, (4) Hamilton Sound sequence, and (5) granitoid plutons. The first four packages are interpreted to be fault-bounded.

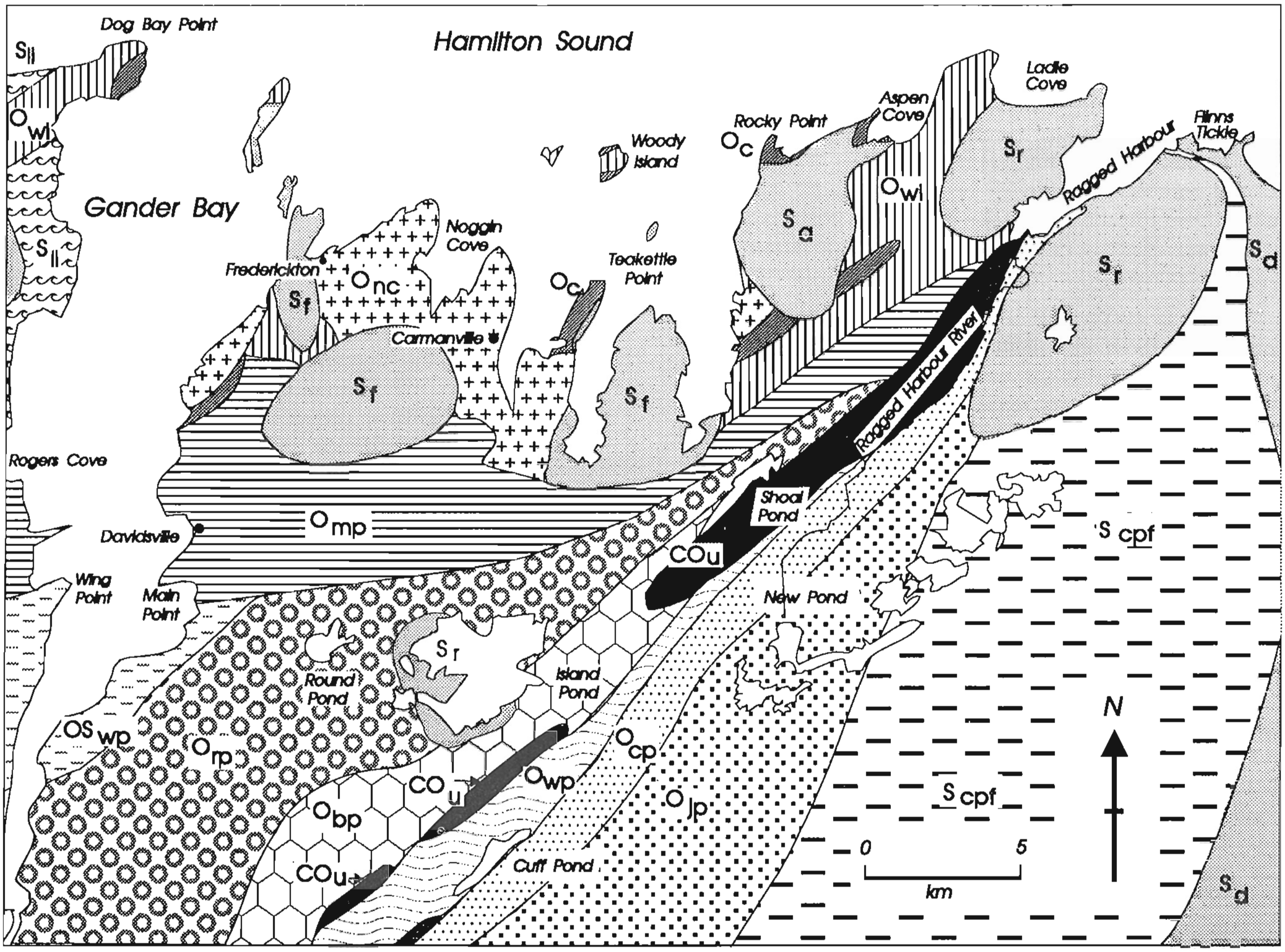
The age of the Gander River Ultramafic Belt (GRUB, unit CO_u) has not been directly determined by isotopic dating, but general considerations suggest that it is of early Ordovician or Cambrian age. Ultramafic rocks form several slabs near Cuff Pond and Shoal Pond consisting mainly of talc-chlorite schist after pyroxenite in which lenses with large porphyroblasts are separated by anastamosing narrow zones of phyllonite. Lenses up to 100 m long consist of massive black serpentinite probably derived from dunite or peridotite. Mafic volcanic rocks, locally pillowed, and associated gabbroic sills outcrop around the edges of the ultramafic rcoks. The volcanic rocks are strongly deformed except in local low-strain augen. Trondhjemite occurs in isolated outcrops which are pervasively shattered and brecciated. The GRUB is bounded to the southeast by a major mylonite zone, and internally dissected by numerous smaller shear zones.

The Gander Group (units O_{jp} and O_{cp}) consists of two units according to O'Neill and Blackwood (1989), namely the psammitic Jonathon's Pond Formation, and varied lithologies of the Indian Bay Big Pond Formation, which are known to be of late Arenig age. In Carmanville map-area the Jonathon's Pond Formation (unit O_{jp}) is easily recognized, but a pelitic unit (Cuff Pond pelite (unit O_{cp})) does not correspond to any formation proposed by O'Neill and Blackwood (1989). The Jonathon's Pond Formation consists of pink-weathering green to grey green feldspathic siltstone, with some beds of coarser, more quartzitic sandstone and pebble conglomerate, and numerous cm scale beds of pelite or semi-pelite. Amphibolitic layers or boudins occur locally, particularly along the chutes of the Ragged Harbour River. The formation everywhere exhibits cm scale layering, and closely spaced, well-developed pressure solution cleavage. Pervasive metamorphism and cleavage make if difficult to determine reliable facing directions.

The Cuff Pond pelite, which forms the northwestern fringe of the Gander Group, consists of homogeneous black pelite with widely-spaced, millimetre scale beds of grey-green psammite, and local composite psammite intervals up to 1 m thick. The pelite south of Shoal Pond contains many lits and boudins of amphibolite and quartz-feldspar porphyry, possibly metamorphosed volcanics. The Cuff Pond pelite cannot be lithologically distinguished from part of the Davidsville Group, but Cuff Pond pelite exhibits ubiquitous folding of cleavage whereas the Davidsville Group exhibits only folds of bedding.

Near the Ragged Harbour pluton (unit S_r) the Cuff Pond pelite is strongly metamorphosed and complexly deformed. Andalusite and sillimanite-rich segregations, migmatite and agmatite with rotated blocks of amphibolite and pelite occur in a complex network of granitic sheets and dykes. This complex, termed the Flinns Tickle complex (unit S_{CPf}), forms a septum 200 to 400 m wide separating the Ragged Harbour and Deadman's Bay plutons which broadens southward into a large area in the south-eastern corner of the map sheet.

O'Neill and Blackwood (1989) divided the Davidsville Group into three formations. In the Carmanville map-area the group can also be divided into three units, but with the exception of the basal Weir's Pond Formation (unit O_{wp}), these units do not agree with those defined by O'Neill and Blackwood (1989). The Weir's Pond Formation contains a significant proportion of red shale with a local basal limestone member which contains a Llanvirn-Llandeilo conodont assemblage. The Weir's Pond Formation is



LEGEND

S_d DEADMANS BAY PLUTON: quartz-plagioclase-biotite matrix with 5-60% of K-spar masses 3 to 10 cm across; border phases may be red and foliated

S_r RAGGED HARBOUR PLUTON: Foliated white biotite granite, massive to foliated biotite-muscovite granite, massive biotite-muscovite-garnet granite, muscovite-garnet pegmatite and associated aplite

S_a ASPEN COVE PLUTON: White, layered biotite granite; foliated granite sheets; marginal biotite-muscovite granite with minor garnet

S_f ROCKY BAY and FREDERICTON PLUTON: quartz-plagioclase-biotite tonalite with large poikilitic biotite; foliated biotite granodiorite to granite; fine-grained to porphyritic granitoid dykes

HAMILTON SOUND SEQUENCE

S_{II} INDIAN ISLANDS GROUP; Laminated grey quartz-rich sandstone with coralline limestone lenses

OS_{wp} WING POINT greywacke; turbiditic black and white greywacke with minor black siltstone and shale; matrix-supported conglomerate lenses including drop-stones.

O_{MP} MAIN POINT shale; black graphitic and pyritic shale, black chert, rare black sandstone up to 30 cm thick. (a) grey sandstone and conglomerate.

O_{NC} NOGGIN COVE FORMATION; pillow lavas,tuff agglomerate, debris flows, resedimented conglomerate; minor black shale and siltstone

O_{WI} WOODY ISLAND siltstone; grey and black siltstones and shale laminated on a cm scale, blue-black-weathering brown Mn-rich siltstone bands, commonly deformed (a) coticule bearing metamorphic rocks

O_C CARMANVILLE MELANGE (equivalent to O_{WI}, O_{NC} and O_{MP} in part); homogeneous black pyritic shale matrix, abundant sedimentary olistoliths and rare volcanic and ultramafic olistoliths; grey pebbly mudstone; carbonate-hosted breccia and minor limestone beds

DAVIDSVILLE GROUP (units O_{rp}, O_{bp}, O_{wp})

O_{rp} ROUND POND SILTSTONE; rhythmically bedded grey to black siltstone to shale sequences; minor granule and shale-flake conglomerate beds

O_{bp} BARRY'S PONDS conglomerate; green sandstone and conglomerate with clasts from the GRUB; fines southwest to siltstone and green-black shale

O_{wp} WEIR'S POND FORMATION; red shale and siltstone; basal dolostone and calcareous shale.

GANDER RIVER ULTRAMAFIC BELT

CO_u deformed mafic volcanic flows and tuff fine-grained gabbro; talc-chlorite schist and phyllonite, serpentinite; red to white, fine-grained, porphyritic

GANDER GROUP (units O_{JP}, O_{cp})

S_{CPF} FLINN'S TICKLE COMPLEX; pelitic migmatite with aluminosilicate nodules, agmatite with amphibolite blocks, all developed from O_{CP}. Massive to deformed granite sheets and veins related to S_r

O_{cp} CUFF POND pelite; strongly cleaved black pelite, with sparse grey psammite beds to 30 cm thick; porphyry and amphibolite lits and boudins

O_{JP} JONATHON'S POND FORMATION; fine-grained greenish feldspathic siltstone with rare pelitic beds; amphibolite boudins Strong pressure solution cleavage

Figure 1. Geological sketch of the Carmanville map-area (2E/8).

conformably overlain by Barry's Pond conglomerate (unit O_{bp}) which contains cobbles and pebbles of the GRUB in a green sandy to silty matrix. This unit fines to the southwest where it digitates with and passes gradationally upward into the Round Pond siltstone (unit O_{rp}). This unit comprises a thick succession of finely laminated grey to black fine siltstone and shale. Typically beds are 5 to 10 mm thick, grading from fine siltstone at the base to discontinuous shaly tops. Seams of shale flake and granule conglomerate occur locally at bed bases.

The sequence Gander Group-GRUB-Davidsville Group occurs throughout east-central Newfoundland, capped by shallow-water to terrestrial red beds and acid volcanics of the Botwood Group. The sequence contains no mid-Ordovician volcanic rocks. This sequence is incompatible with that found along Hamilton Sound in the northern part of the Carmanville map-area. This Hamilton Sound sequence contains abundant volcanics within the sedimentary sequence, spectacular melange, and an entirely marine and sedimentary Silurian section (Indian Islands Group).

The Hamilton Sound sequence below the Indian Islands Group consists of the Woody Island siltstone, the Noggin Cove Formation (volcanics), the Main Point Shale, and the Wing Point greywacke. The Woody Island siltstone (unit O_{wi}) exhibits alternating shale and siltstone beds from 1 to 5 cm in width. The shales are dark grey to greenish and locally silicified, but weather white. The most typical feature of the Woody Island siltstone is manganese-rich beds 1 to 2 cm thick, which are brownish to blue-black weathering, tough, very fine siltstone at low grade, and pink fine-grained rounded garnet and quartz (coticules) at higher grade. Manganese-rich beds are typically intricately boudinaged and folded, even where the rest of the beds are little deformed. Coticules occur in the Flinns Tickle complex southeast of New Pond but ubiquitous, abundant occurrence (up to 10 per cent by volume) typifies the Woody Island siltstone.

The Noggin Cove Formation (Williams et al., 1991, unit O_{nc}) is interbedded with the Woody Island siltstone at Beaver Cove along the east shore of Gander Bay. The Noggin Cove Formation consists of basaltic pillow lavas, tuffs, tuff breccia, and a variety of resedimented volcanic rocks, including spectacular debris flows (Pickerill et al., 1981). Thin interbeds of black siltstone and shale similar to units above and below also occur. The Noggin Cove Formation includes a wider variety of lithologies than the volcanic rocks of the GRUB, and is considerably less deformed.

The Noggin Cove Formation is faulted against a varied sequence of sedimentary rocks here termed the Main Point shale. The most characteristic lithology of this sequence is black, graphitic and pyritic shale which contains Caradocian graptolites (Williams, 1964), but a variety of other lithologies occurs with it, notably black pyritic chert as nodules and beds up to 5 cm thick, and Mn-rich siltstone similar to the Woody Island beds. Massive black sandstone beds up to 1 m thick occur locally, as well as pale grey sandstone, and conglomerate with sandstone clasts. The lower part of the section contains at least one layer of mafic lava >5 m thick, and gabbro sills occur higher in the section.

The Carmanville melange (Pajari et al., 1979, unit O_c) forms a non-stratigraphic unit which sampled the Woody Island siltstone, the Noggin Cove Formation and the Main Point shale, and is in part equivalent to them. In addition to unstratified fragmental rocks, the melange includes bed-like lenses of carbonate-hosted breccia and moderately disrupted limestone beds on Rocky Point. Features of the melange have been described by Pajari et al. (1979). The unit varies from homogeneous black, rusty shale with rare, but large (>5 m) blocks to grey mudstone with 20-30 per cent of small, angular to rounded, matrix-supported, shale and siltstone fragments. Many of the fragments are locally derived, so that there is a crude zonation to the fragment load with volcanic fragments near the Noggin Cove Formation, and ultramafic fragments near the GRUB. However provenance of many clasts (gabbro on Dog Bay Point, deformed and metamorphosed volcanic clasts occur on Teakettle Point, Carmanville Point and near Frederickton) is unknown. Relations between melange and more coherent stratigraphic section are spectacularly exposed on Woody Island, where siltstone beds are progressively boudinaged and incorporated into melange. Locally melange is intruded in the form of dykes along cleavage. In such dykes the clasts are invariably more flattened and deformed than in the body of the melange.

The Wing Point greywacke consists of turbiditic dark grey greywacke with well developed Bouma cycles grading from sandstone at the base to chloritic shale at the top. The unit contains lenses of matrix supported conglomerate in which some clasts appear to be drop-stones (Pickerill et al., 1979). Relations of the Wing Point greywacke to other units are not well established due to lack of outcrop, but it is interpreted to overlie the Main Point Shale. Because tops can readily be determined in this unit, it can be demonstrated to contain downward-facing folds, indicating early recumbent folding (Karlstrom et al., 1982).

The Indian Islands Group outcrops along the west shore of Gander Bay from Rogers Cove to Dog Bay point, but is not found east of Gander Bay. The rocks consist of thick-bedded grey to buff, quartz-rich sandstone with millet-seed quartz grains indicating a mature, well sorted sediment. Limestone lenses up to a metre long are common and locally contain a coral fauna (*Favosites* cf. *favosus*, Williams, 1964). Near Dog Bay point the Indian Islands Group is in contact with melange and Woody Island siltstone. On the west side of the point this contact is clearly faulted, but on the east side of the point relations are unclear.

Five plutons occur within the Carmanville map-area, namely the Rocky Bay, Frederickton, Aspen Cove, Ragged Harbour, Island Pond plutons. In addition, small parts of the Deadman's Bay and Tim's Harbour plutons outcrop within the area. With the exception of the Deadman's Bay pluton, granitoid rocks petrographically and chemically form a continuum. It is convenient to distinguish three major sub-types, namely (i) Rocky Bay-Frederickton type, biotite tonalite with large round, poikilitic biotite and no K-spar (unit S_r) and maginal massive to foliated biotite granodiorite (unit S_a), (ii) Ragged Harbour type, foliated biotite granite, biotite-muscovite granite, and muscovite pegmatite, with distinctive quartz-rich muscovite-garnet pegmatite and aplite which is found as lits and veins in all the metamorphic rocks (unit S_r), The Island Pond pluton is a pink massive leucogranite within the Davidsville Group, and is chemically related to the Ragged Harbour pluton. (iii) Deadman's Bay pluton, coarse-grained biotite granodiorite to tonalite matrix with varied proportions of masses of K-feldspar (megacrysts) ranging in habit from subhedral to globular, and in size from 2 to 15 cm (unit S_d). In general the proportion of Rocky Bay-Frederickton type decreases eastward, while the proportion of Ragged Harbour type increases up to the margin of the Deadman's Bay pluton.

The youngest rocks on the Carmanville map-sheet form a swarm of biotite-pyroxene lamprophyre dykes along the west shore of Gander Bay. These dykes commonly trend northeast, and rarely exceed 1 metre in width. Multiple intrusions with internal chill zones occur. These dykes are believed to form part of the Jurassic lamprophyric province around Bay of Exploits described by Strong and Harris (1974).

GEOCHEMISTRY

The results of 90 chemical analyses of sedimentary and plutonic rocks are summarised in Table 1. More detailed discussion of these analyses will be presented elsewhere. Psammites of the Jonathons Pond Formation (column 1) clearly contain debris from mafic to intermediate rocks (Fe and Mg contents, but the aluminum and alkali content suggest a continental source. The overlying Cuff Pond pelite (column 2) is broadly similar in chemistry with higher alumina and alkali contents expected from clays. Rocks of the Flinns Tickle complex (column 3) are essentially identical to the Cuff Pond pelite, although the higher K/Na ratio may indicate some degree of metasomatism. The composition of Cuff Pond pelites is remarkably similar to that of pelites from the Davidsville Group (column 4) which contain debris from both GRUB and continental sources (Pajari et al., 1979). There is a marked contrast between these siliceous, soda-dominant pelites and the potassic, Mn-rich pelites of the Main Point shale (column 4). The high Mn content of the latter presumably indicates a significant component of exhalative volcanic material.

The geochemistry of the granitic rocks is illustrated in columns 6-8. With the exception of K and Rb, which petrographic evidence suggest to be metasomatically enriched in the Deadman's Bay pluton, there is a general progression in values from the Frederickton and Rocky Bay plutons through Deadman's Bay to Ragged Harbour. The generally siliceous, highly evolved character of the all the granitoid rocks, the lack of mafic members, and the local concentration of B and Be (petrographically evidenced by presence of quartz-tourmaline symplectites and beryl pegmatites) is characteristic of Gander zone granites (compare Williams et al., 1988). These features are shared by plutons emplaced into the Gander Group, the Davidsville Group (Island Pond Pluton) and the Hamilton Sound sequence, and imply derivation of plutons in all three terranes from similar sources which contained significant amounts of continentally derived sedimentary material.

Table 1. Selected geochemical data for the Carmanville area

	1	2	3	4	5	6	7	8
SiO2	68.14	64.68	64.60	66.04	58.37	64.73	69.80	73.54
TiO2	0.80	0.60	0.72	0.34	0.94	0.49	0.57	0.11
Al2O3	14.04	17.18	16.58	15.98	17.35	16.31	14.42	14.54
Fe2O3	1.38	0.82	2.06	0.46	1.83	0.65	0.68	0.32
FeO	4.72	3.95	3.32	3.32	6.60	3.53	2.32	0.60
MnO	0.12	0.13	0.15	0.11	1.12	0.11	0.07	0.05
MgO	2.41	1.71	1.48	1.77	3.13	2.19	1.08	0.31
CaO	1.23	0.99	0.64	1.58	1.49	4.08	1.29	0.64
Na2O	2.60	3.98	2.76	4.40	1.58	3.73	3.15	3.93
K2O	2.21	3.01	4.32	3.54	3.66	1.60	4.77	4.40
P2O5	0.14	0.19	0.22	0.13	0.20	0.14	0.15	0.13
H2O	2.34	2.08	2.16	1.91	3.04	1.84	1.32	0.81
CO2	0.06	0.21	0.24	0.62	0.63	0.23	0.13	0.00
S	0.04	0.11	0.06	0.02	0.06	0.07	0.05	0.02
Total	100.23	99.64	99.31	100.22	100.00	99.70	99.80	99.40
Trace elements (in ppm)								
Rb	91	210	208	129	162	69	235	269
Zn	64	76	94	30	102	73	80	24
B	48	42	26	34	104	nd	nd	51
Ba	345	421	576	241	462	297	654	229
Co	10	9	7	8	26	3	3	nd
Cr	91	53	31	203	129	29	7	nd
Cu	15	35	16	14	74	21	7	nd
Ni	39	37	18	27	98	21	11	nd
Sr	139	130	133	114	114	271	206	71
V	95	131	71	57	116	88	48	11
Be	nd	2.1	3.2	nd	nd	nd	nd	2.6
Yb	2.7	4.1	5.5	2.5	5.6	4.9	4.3	2.1
Zr	179	102	102	152	156	74	148	59

nd - not detected

column 1; Jonathon's Pond Formation, average of 6 analyses
column 2; Cuff Pond pelite, average of 5 analyses
column 3; Flinn's Tickle complex, average of 6 analyses
column 4; Round Pond pelite, average of 7 analyses
column 5; Main Point pelite, average of 7 analyses
column 6; Frederickton and Rocky Bay plutons, average of 6 analyses
column 7; Deadman's Bay Pluton, average of 4 analyses
column 8; Ragged Harbour pluton, average of 10 analyses

Analyses by Rapid Methods Group, Geological Survey of Canada

STRUCTURE

Rocks of the Carmanville area exhibit a dominant northeast-trending cleavage which commonly dips steeply northwest east of Gander Bay, and southeast west of the Bay. The cleavage is axial planar to upright, tight to isoclinal large folds with wave lengths of hundreds of meters to kilometers. Plunges of these folds vary from 0 to 90°, and in some cases they resemble sheath folds. In shale-dominated successions the dominant cleavage virtually obliterates older structure, including sedimentary structures necessary to establish facing directions. However within the Wing Point greywacke facing directions can be readily determined, and folds not associated with the dominant cleavage can locally be shown to be downward-facing. Karlstron et al. (1982) pointed out that this configuration implies early recumbent folding. Piasecki (personal communication, 1991) has determined from microscopic observations that the large folds in the Hamilton Sound succession represent D3 deformation. According to Karlstrom et al. (1982), downward-facing folds affect the Indian Islands Group, and are therefore of Silurian or younger age. The age of other deformation must be younger.

Evidence for recumbent folding has not been found in the Davidsville Group. This could be due to destruction of evidence by the dominant cleavage, but the map pattern does not suggest any repetition of stratigraphy, and facing directions, where discernable are consistently westward. The Davidsville Group therefore appears to have undergone a different style of deformation from the Hamilton Sound succession involving only a single major period of folding.

The Gander Group has long been known to exhibit a more complex structural history than the Davidsville Group, as witnessed by common intrafolial folds of early cleavage (Kennedy and McGonigal, 1972). The GRUB and a narrow fringe of Cuff Pond pelite exhibit yet another style of deformation dominated by transcurrent faulting. Movement included an early sinistral phase, followed by a dextral phase and a latest sinistral phase (M.A.J. Piasecki, personal communication, 1991). This latest phase was approximately coeval with emplacement of the Ragged Harbour pluton, some phases of which are mylonitized, while others show layering or foliation only. Since this pluton gave K-Ar ages as old as 410 Ma (Currie and Pajari, 1981), folding, faulting and plutonism in Carmanville map-area appears to be confined to a short period of Silurian time spanned by deposition of the (folded) Indian Islands Group and emplacement of the plutons.

METAMORPHISM

High-grade metamorphic rocks outcrop in two belts, each about 5 to 10 km wide. A northern belt extends across the map-sheet in an east-west direction from the Gander Bay pluton to Ragged Harbour. Currie and Pajari (1981) showed that in the eastern part of this belt metamorphic isograds parallell the western (White Point) of the Ragged Harbour pluton, and that index minerals appeared in the sequence garnet, andalusite "spots", staurolite, coarse pink andalusite, cordierite, sillimanite. They estimated peak temperatures and pressures at about 680° and 4 kilobars, and showed that extensive partial melting had occurred. Several lines of evidence, including numerous quartz-aluminosilicate segregations, suggested high water fugacity during melting.

Similar grades of metamorphism occur as aureoles around plutons west of the Ragged Harbour pluton. These aureoles are a few tens or hundreds of meters wide around the major plutons, but along the east shore of Gander Bay and on two islands cordierite-andalusite rocks occur over a considerable area, suggesting that a more regional development may be present. Textural relations among metamorphic minerals indicate that peak metamorphism accompanied and outlasted D2 deformation of the Hamilton Sound sequence. Early poikiloblastic andalusite overgrows the dominant fabric, but is itself folded and broken in stretching lineations. Quartz-andalusite segregations commonly cut cleavage, but are locally folded. Further south in the Davidsville Group, the Island Pond pluton exhibits a narrow (<50 m) andalusite-cordierite aureole which overprints the main foliation. The Davidsville Group lies outside the belt of metamorphic rocks, and is little affected except for a small area north of Shoal Pond.

An eastern belt of metamorphism extends south from Ragged Harbour to Ocean Pond on the Weir's Pond map-sheet (2E/1, see O'Neill and Knight, 1988). Peak grade and mineralogy closely resemble those in the northern belt, but migmatization and partial melting are more extensive. Isograds trend approximately north-south, but are truncated by the Ragged Harbour and Deadman's Bay plutons. There is a striking similarity between the two metamorphic belts in texture, mineralogy and relation of metamorphism to structure. Even such details as rosettes of large pink andalusite crystals (splendid andalusite of Currie and Pajari, 1981) are common to both belts.

Table 2. Comparative stratigraphy of the Bay of Exploits area and the Hamilton Sound sequence

δ		Bay of Exploits	Hamilton Sound
Llandovery		Indian Islands Group	Indian Islands Group*
Ashgill		Sansom greywacke	Wing Point greywacke
Caradoc		Lawrence Harbour shale	Main Point shale*
Llanvirn-Llandeilo	Exploits Group	Lawrence Head volcanics	Noggin Cove Formation
		New Bay Formation	Woody Island siltstone
		Saunders Cove Formation	
		Tea Arm volcanics	

* Age fixed by fossils in the Hamilton Sound sequence. Ages of other units of the Hamilton Sound sequence are uncertain, but interpreted to be as shown.

DISCUSSION

Previous mapping in the Carmanville region grouped all sedimentary rocks between the GRUB and the Indian Islands Group (Williams, 1964; Kennedy and McGonigal, 1972; Currie et al., 1980). In this scheme the type section of the Davidsville Group at Davidsville appeared anomalous relative to sections further south since it contained volcanic rocks and melange. Currie et al. (1980) attempted to explain this anomaly by assuming that the volcanics were allochthonous and that melange occurred throughout the Davidsville Group. Both these postulates appear unfounded. The Noggin Cove Formation appears to be interbedded with both the Woody Island siltstone and the Main Point shale, and true melange with large olistoliths occurs only in the Hamilton Sound sequence. When the melange and volcanic-bearing sequence is considered separately, it can be matched unit by unit with the Exploits Group to the west (Table 2). Further, the trend of units within this Hamilton Sound sequence is approximately east-west, whereas the Davidsville Group to the south trends northeast for tens of kilometres.

These relations suggest that Carmanville map-area contains parts of three sequences rather than the two previously recognized. Original depositional relations, if any, of these three sequences remain uncertain, but geochemical evidence suggests that all three may be related. The Cuff Pond pelite resembles Davidsville shales, but contains Mn-rich horizons (coticules), an unusual lithology typical of the Hamilton Sound sequence. It is plausible that all three sequences formed parts of a once continuous tract. Ultramafic material in all three sequences presumably was derived from GRUB equivalents, but possibly from regions far removed from present outcrop of the GRUB, since all of the Dunnage zone is assumed to be floored by similar material (Williams, 1978). The mechanism and timing of juxtaposition remain unclear. The Gander and Davidsville Groups are separated by a major fault with a large mylonite zone of probable Silurian age. The southern edge of the Hamilton Sound succession is locally strongly deformed, but a fault, if present, must be strongly deformed with a sinuous surface trace. As pointed out by Karlstrom et al. (1982), this is a necessary condition for a pre-Silurian fault in this region.

Previous failure to separate the Davidsville Group and the Hamilton Sound succession resulted in large part from the abundance of black shale in both sedimentary successions. All these shales were correlated in the past, but it is now clear that several units are involved. Shales of the Woody Island siltstone commonly are greenish-black, and associated with Mn-rich beds. Shales of the Main Point shale are pyritic and graphitic, weather rusty, and associated with chert. The Round Pond shales weather to pale grey or silvery shades, and occur in rhythmic alternation with siltstone.

Within the Carmanville map-area plutonism and metamorphism occur in two linear belts which lie close to, but not on, boundaries of diverse sedimentary successions. Currie and Pajari (1981) suggested that metamorphism and plutonism resulted from self-heating due to tectonic thickening of the sedimentary pile. The present observations can be explained by this model if the thickening is due to thrusting and the sites of the metamorphic belts represent ramps where the thickness of section is essentially doubled. The east-west belt is assumed to represent a frontal ramp, whereas the north-south boundary could represent a lateral ramp. Zen (1988) calculated that after initial thickening, 20 Ma would commonly be sufficient to develop high temperatures, and that as obsrved in this region high temperatures would outlast deformation.

If the Hamilton Sound succession is correlative to the Exploits Group, then this succession forms an allochthonous, southerly transported sheet, and the Exploits Sub-zone of the Dunnage zone consists of two or more sheets with distinct stratigraphies. The northern sheet is characterised by mid-Ordovician volcanics and a shallow to deep marine Silurian succession (Indian Islands Group), while the southern sheet lacks mid-Ordovician volcanics and contains a shallow marine to terrestrial Silurian succession containing acid volcanics (Botwood Group). On this reasoning, correlation of deep marine strata of Exploits Bay with the Botwood Group (Boyce et al., 1991) is inappropriate. They should be correlated with the Indian Islands Group. Such a model is compatible with the tectonic model of Piasecki and Currie (1988) according to which the Dunnage zone consists of a stack of southerly transported thrust sheets.

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Geological Survey of Canada Project 730044

The "Lorneville Beds": a latest Precambrian sequence near Saint John, New Brunswick

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Abstract

The "Lorneville beds", an informally defined unit, underlie the Tommotian Ratcliffe Brook Formation of the Saint John Group conformably or slightly disconformably and overlie Precambrian volcanic rocks (Coldbrook Group) disconformably or unconformably. The "Lorneville beds" comprise an upper red massive conglomerate member up to 55 m thick (L1), a volcanogenic member up to 300 m thick (L2), and a member up to 200 m thick composed of thinly layered red siltstone or pink sandstone with interbedded red and green siltstone (L3). Tholeiitic basalts of the volcanogenic interval are chemically distinct from older and younger units. Complete sections of "Lorneville beds" are rare due to extensive faulting. The "Lorneville beds" outcrop over an area at least 25 by 60 km, and probably correlate with other units extending farther. They form part of a volcano-sedimentary sequence emplaced during subsidence spanning the Cambrian-Precambrian boundary.

Résumé

Les «couches de Lorneville», unité définie de façon informelle, sont sous-jacentes, en concordance ou en légère disconformité, à la formation tommotienne de Ratcliffe Brook du groupe de Saint John, et recouvrent des roches volcaniques précambriennes (groupe de Coldbrook) en disconformité ou en discordance. Les «couches de Lorneville» englobent un membre conglomératique supérieur, massif et rouge, atteignant 55 m d'épaisseur (L1), un membre volcanogénique atteignant 300 m d'épaisseur (L2) et un membre atteignant 200 m d'épaisseur et composé d'un siltstone rouge ou d'un grès rose finement stratifiés, avec interstrates de siltstone rouge et vert (L3). Les basaltes tholéiitiques de l'intervalle volcanogénique sont chimiquement distincts des unités plus anciennes et des unités plus récentes. Les profils stratigraphiques complets des «couches de Lorneville» sont rares en raison du grand nombre de failles. Les «couches de Lorneville» affleurent sur une superficie d'au moins 25 km sur 60 km, et se laissent probablement corréler avec d'autres unités se prolongeant plus au loin. Elles constituent une partie d'une séquence volcano-sédimentaire mise en place durant l'épisode de subsidence qui a englobé la limite entre le Cambrien et le Précambrien.

INTRODUCTION

The stratigraphy of the Saint John area of southern New Brunswick has been studied for more than 150 years, but due to the structural complexity of the region, recurrence of similar lithologies at different stratigraphic levels, and relative paucity of fossils, several fundamental stratigraphic relationships remain uncertain and controversial. Tanoli and Pickerill (1988) recently revised the formal stratigraphy of the Cambrian and lower Ordovician Saint John Group, and in another publication considered specifically the lowest formation, the Tommotian Ratcliffe Brook Formation. In both these contributions they note the presence of red beds, commonly interlayered with volcanics, lying below the Ratcliffe Brook Formation. These strata were informally termed "Eocambrian" by Currie (1984). A well exposed and complete section of these rocks occurs along the shore of Lorneville Harbour (Fig. 1) and was named informally "Lorneville beds" by Rast et al. (1978), who believed them to be of Carboniferous age. I here draw attention to correlatives of the "Lorneville beds" and information which fixes their age as latest Precambrian.

DEFINITION AND LITHOLOGY

The "Lorneville beds" are here defined to be those strata which can be observed, or reasonably inferred, immediately to underlie the lowest units of the Saint John Group, either the Ratcliffe Brook Formation, or the overlying Glen Falls Formation where the Ratcliffe Brook Formation is absent. Figure 1 shows the distribution of "Lorneville beds" and related rocks in the Saint John region. The beds are well exposed along the coast of the Bay of Fundy from Musquash Harbour to the Saint John Dry Dock and sections can also be examined in the Mystery Lake area of East Saint John, in Ratcliffe and Hanford Brooks northeast of Saint John and on the shores of the Long Reach north of Saint John. Fault-bounded slivers of similar rocks occur within several fault zones, and strata probably correlative to the "Lorneville beds" occur up to 50 km west of Saint John (Currie, 1987). A minimum age for the "Lorneville beds" is given by Tommotian fossils in the overlying Ratcliffe Brook Formation (Tanoli and Pickerill, 1990). An absolute age for part of the group may be given by a Pb-U zircon age of 555 ± 8 Ma on the Musquash pluton (Currie and Hunt, 1991) which grades into volcanic rocks correlative to the "Lorneville beds".

An idealised section of the "Lorneville beds" (Fig. 2) consists of three members, an upper red conglomerate (L1), a volcanic interval (L2) and a lower siltstone and sandstone interval (L3). The "Lorneville beds" along Lorneville Harbour include mafic volcanic rocks including flows, tuff and peperite (member L2) exposed on both limbs of an anticline, underlain by thinly bedded red and green siltstone and sandstone, locally calcareous (member L3). Neither the top nor the bottom of the section is exposed. On the north side the section is truncated by purple quartz-cobble conglomerate of the Carboniferous Balls Lake Formation, whereas to the

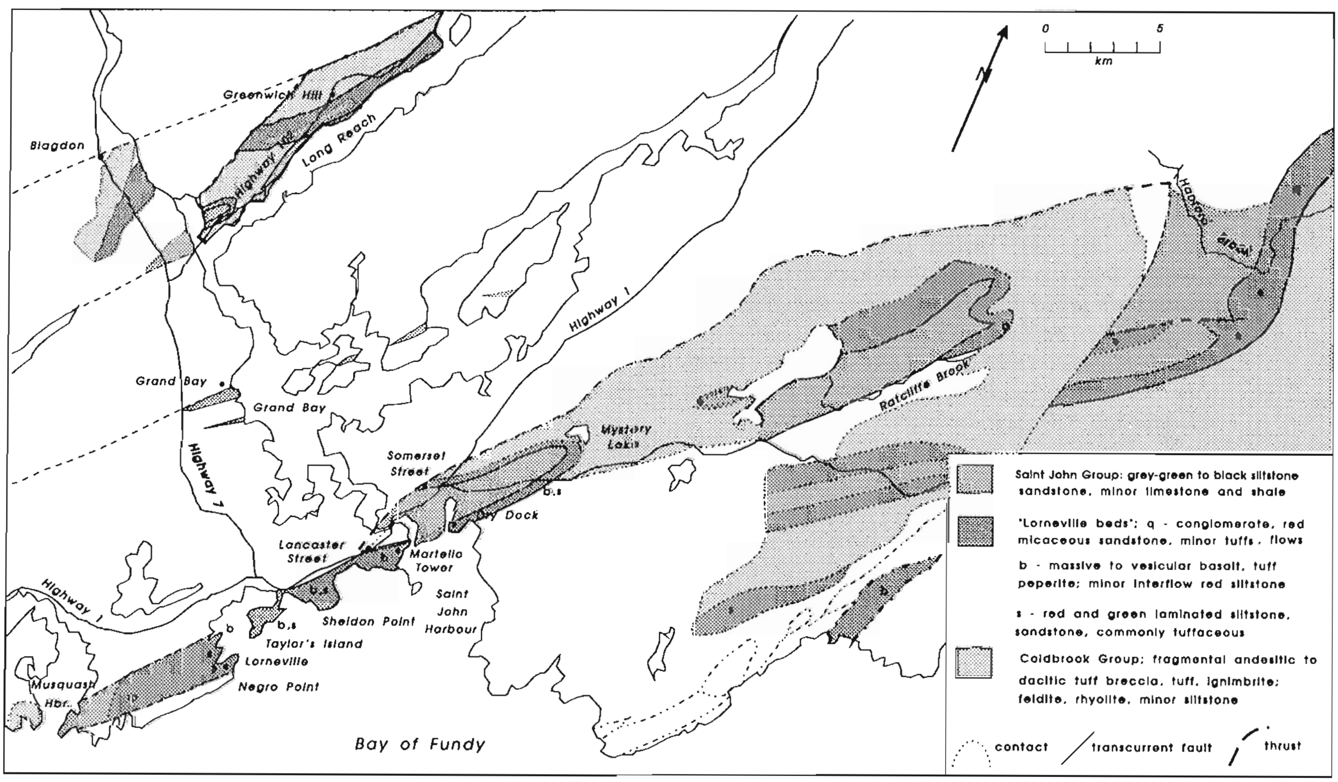

Figure 1. Distribution of the Lorneville, Coldbrook and Saint John Groups near Saint John, showing locations referred to in text. Blank regions are underlain by other units.

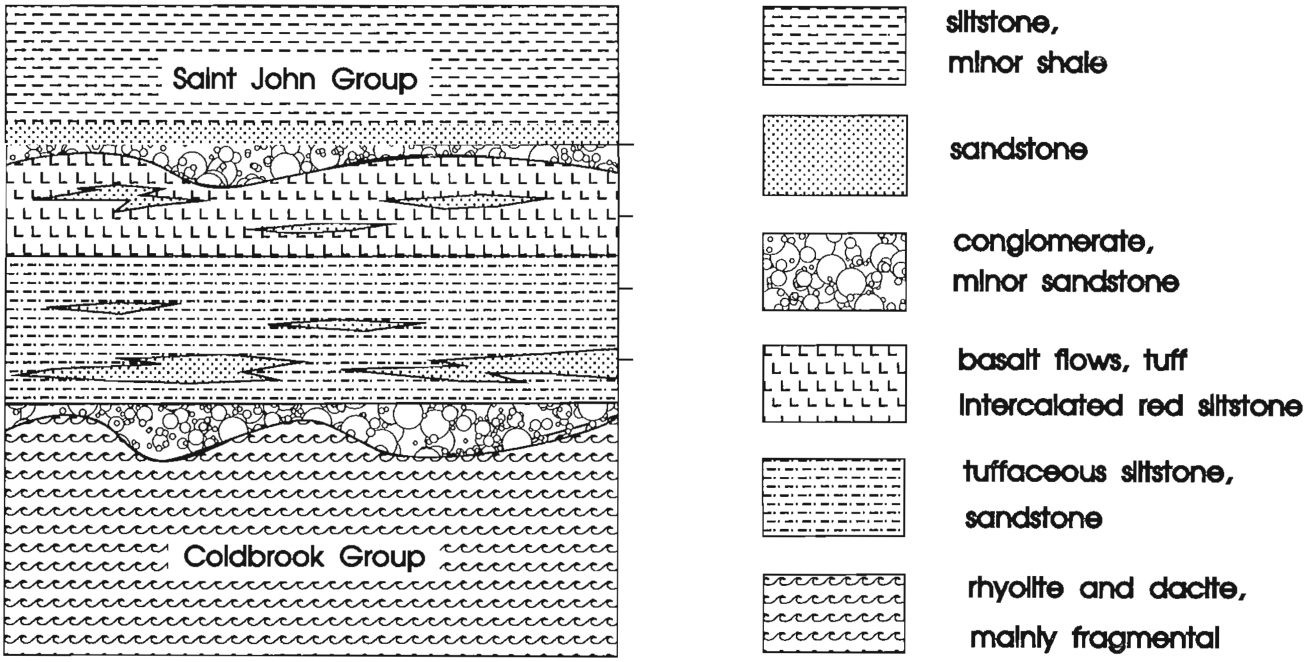

Figure 2. Diagrammatic stratigraphic column for "Lorneville beds".

Figure 3. Massive red conglomerate (L1), parking lot of Sobey's store, Grand Bay. (GSC negative 1991-036)

Figure 4. Cross-bedded pebble conglomerate (L1), fining upward to dark sandstone. Greenwich Hill (Lens cap 5 cm across, 1991-034B)

south it is truncated by a fault which juxtaposes the "Lorneville beds" against older, more deformed, volcanic rocks of the Coldbrook Group.

The top of the "Lorneville beds" can be seen in Ratcliffe and Hanford Brooks, where it consists of up to 55 m of red, massive conglomerate, with subordinate feldspathic sandstone lenses (Fig. 3, 4). Tanoli and Pickerill (1990) designated such rocks as facies RB1 of the Ratcliffe Brook Formation but the following considerations suggest that the conglomerate should be included with the underlying unit: (1) It was deposited sub-aerially whereas the rest of the Ratcliffe Brook Formation is marine (Tanoli and Pickerill, 1990); (2) it grades to or is interlayered with volcanogenic rocks, which are absent in Ratcliffe Brook Formation (Tanoli and Pickerill, 1990); (3) the conglomerate is not basal (as supposed by Tanoli and Pickerill (1990)), since it is consistently underlain by red and green siltstones, and may occur as lenses within such rocks and (4) there is a sharp upper contact between the conglomerates and overlying flaggy micaceous sandstone typical of the Ratcliffe Brook Formation. This conglomerate, here designated L1, can be distinguished from several other red conglomerates in this region by (a) intense crimson to purplish colour, (b) volcanogenic nature of cobbles, and rarity of granite and vein quartz cobbles, (c) presence of tuffs within the section and (d) rarity of shale or siltstone lenses (compare Tanoli et al., 1985).

The lower boundary of L1 against L2 may be gradational or abrupt, but conformable in either case. The interval L2 is the most varied part of the "Lorneville beds", but always contains a major volcanogenic component. It ranges from crimson to bright maroon chloritic mudstone with pale greenish lenses up to several metres long and 10 cm wide at Taylors Island, West Saint John and Hanford Brook, to apple green tuffs and pillowed to peperitic basalt with vesicular masses of fine-grained massive material in a jade-green volcanoclastic matrix (Fig. 5) in Lorneville, to red welded tuffs at Taylors Island and Beulah Camp (Fig. 6). The most typical lithology however is basalt flows and intercalated thin red siltstone, such as those exposed on Lorneville Point where at least five separate flows 15-30 m thick can be distinguished by characteristic amygdale and phenocryst contents. The flows are separated by 2 to 5 m of siltstone, either red or striped in red and pale buff on a centimetre scale. One flow contains an inclusion of blue grey marble 2 x 2 x 0.5 m in dimension surrounded by a 30 cm skarn rind, which resembles nearby late Proterozoic Green Head Group marbles (Currie, 1987). None of the flows exhibit pillows or peperitic phenomena. Similar basalts are exposed for almost 20 km along the Bay of Fundy coast at Taylors Island, Sheldon Point, Martello Tower, and the Saint John Dry Dock. Two thin basalt flows separated by red sandstone are present in the same stratigraphic interval south of Greenwich Hill along the Long Reach. The thickness of the volcanic section L2 ranges from almost 300 m at Lorneville Harbour to about 10 m at Beulah Camp. It appears to be absent in Hanford Brook.

Preliminary chemical analyses of L2 basalts (Table 1) show that they are tholeiites slightly more tholeiitic and within-plate in character than the Coldbrook Group on FeO*-MgO-Na_2O+K_2O and Zr/Y-Ti/Y plots (Fig. 7 A,B). On most trace element plots they fall in the same fields as Coldbrook Group rocks (Fig. 7 C,D,E,F), including such process-dependent plots as Sr-Nb. This similarity suggests that the Lorneville Group volcanics derived from the same source as the Coldbrook Group and crystallised under similar conditions. The extreme K_2O and Rb depletion (Fig. 7 G,H) is also seen in calc-alkaline, predominantly andesitic, plate margin volcanic rocks of the Coldbrook Group and in Silurian alkali basalts of the Long Reach Formation and Kingston complex. It probably results from Carboniferous disturbances (Currie and Eby, 1990).

The contact between volcanic rocks of L2 and red to green siltstone of L3 is gradational. The contact is arbitrarily placed at the first buff to pale purple, graded and cross-bedded sandstone lens. These lenses, up to 15 cm thick (Fig. 8) may form up to 15 percent of the upper part of the L3 and become more abundant toward the base. Many of these layers contain weathered feldspar fragments and abundant red or green chlorite, presumably volcanogenic. The bulk of this member consists of very even-bedded red to pale green siltstone, with bed thickness of 1-2 cm, giving the strata a striking striped appearance (Fig. 9). The L3 member exhibits a rich variety of primary sedimentary structure (Fig. 10), including numerous slumps and growth faults.

DESCRIPTION OF LOCALITIES

The upper contact of the "Lorneville beds" against the Saint John Group can be examined in Ratcliffe and Hanford Brooks, at the Saint John Dry Dock, Mystery Lake, Somerset Street and Lancaster Street in the city of Saint John, around

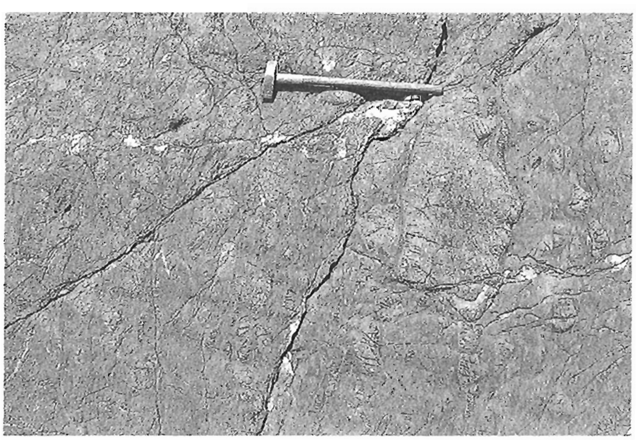

Figure 5. Peperitic basalt (L2), Lorneville Harbour. (Hammer 80 cm long, 1991-035D)

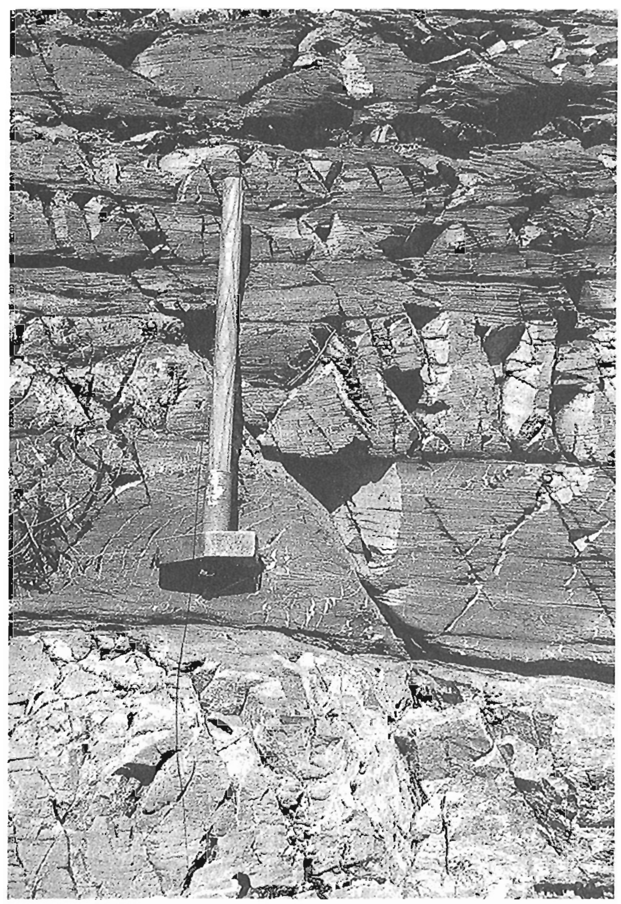

Figure 6. Finely laminated basaltic tuff resting on welded rhyolitic tuff (L2). Taylors Island. (1991-034M)

he Coleson Cove power station west of Lorneville, and at Devils Back, north of the Long Reach. The lower contact of the "Lorneville beds", against volcanic rocks, is exposed in Hanford Brook and at Devils Back north of the Long Reach. Descriptions of these localities follow.

In Ratcliffe Brook near the hamlet of Quaco Road, 8 m of red conglomerate underlie flaggy micaceous sandstone of the Ratcliffe Brook Formation. The lower part of the Ratcliffe Brook Formation commonly exhibits a dull, dark maroon colour, but in the brook the lowest beds are pale tan sandstone. About 15 cm of section is covered at the actual contact, but attitudes above and below the contact are conformable. The conglomerate consists of a dark crimson, chloritic matrix containing rounded to sub-angular clasts of rhyolite, porphyry, fine-grained quartzite and rare basalt from 5 to 15 cm in largest dimension (Fig. 3). The igneous rocks could be derived from nearby localities, but no nearby source is known for the quartzite. It somewhat resembles fine-grained quartzite in the Green Head Group around Lily Lake in the city of Saint John. The conglomerate lacks bedding, but rare 5-15 cm interbeds of white to pale purple sandstone exhibit well developed graded bedding. The lower contact is not exposed in Ratcliffe Brook.

A similar but much thicker section of conglomerate occurs in Hanford Brook, calculated to be 55 m thick. In both sections the contrast between the monotonous flaggy, micaceous sandstone of the Ratcliffe Brook Formation and the underlying massive conglomerate is striking. Below the red conglomerate in Hanford Brook, 86 m of evenly bedded, red and green, locally calcareous siltstone rest on andesitic to dacitic flows of the Coldbrook Group. Much of this interval consists of very regular beds 5-15 mm thick, but the basal 6 m is massive, buff conglomerate composed of rounded rhyolite clasts up to 30 cm across. The contact with the underlying volcanic rocks is sharp, but not clearly unconformable, although there is a striking change from stratified rocks in which primary features are easily discernable to volcanic rocks which are strongly altered and lack obvious primary features.

At the Saint John Dry Dock the contact between red feldspathic sandstone and conglomerate and the Ratcliffe Brook Formation is exposed in a rock cut at the rear of Willets Food Warehouse. A gradational contact between conglomerate and the underlying tuffaceous beds (L2) is well exposed in the nearby road cut along Bayside Drive. The contact between these tuffaceous beds and rather massive

Table 1. Chemical analyses of volcanics from the "Lorneville beds"

Sample	83042	83125	83117	83118	79066	82156	82107	80138	82064	82134	83221	80134
Grid	232086	244390	360223	360223	530230	078006	515157	328176	296147	477118	279118	262173
SiO2	54.10	46.80	48.00	47.55	77.10	78.00	46.15	47.55	49.40	47.25	45.45	68.60
TiO2	1.12	0.70	1.95	0.53	0.15	0.15	2.92	1.01	1.34	1.82	3.07	0.07
Al2O3	15.20	16.65	15.40	16.4	12.40	12.25	14.55	16.55	16.15	16.50	14.15	14.45
Fe2O3	2.90	3.35	4.05	2.65	1.60	0.15	7.05	2.80	2.45	4.30	7.50	0.50
FeO	5.65	4.80	7.65	6.15	0.10	0.65	8.40	6.00	7.35	6.95	8.75	4.00
MnO	0.16	0.29	0.22	0.15	0.02	0.01	0.55	0.16	0.21	0.35	0.34	0.11
MgO	6.30	7.90	5.95	9.3	0.20	0.25	5.45	6.80	6.65	7.40	6.90	2.10
CaO	4.71	9.09	8.69	9.56	0.28	0.12	7.43	12.04	7.49	7.78	4.72	2.34
Na2O	4.65	2.75	3.65	2.4	5.35	3.25	4.20	1.00	3.40	4.05	4.80	2.35
K2O	0.04	0.05	0.18	0.34	2.10	4.83	0.06	0.04	1.29	0.04	0.12	2.76
H2O	3.85	4.90	3.20	4.05	0.35	0.60	3.35	5.65	3.65	4.00	4.00	1.65
CO2	1.45	2.90	1.50	1.3	0.20	0.10	0.20	0.20	0.10	0.10	0.10	0.11
P2O5	0.29	0.09	0.25	0.09	0.00	0.01	0.34	0.23	0.63	0.14	0.31	0.14
Total	100.42	100.27	100.69	100.47	99.85	100.37	100.65	100.03	100.11	100.68	100.21	99.97
Trace element data in ppm												
Ba	196	84	168	142	421	429	139	82	421	82	167	479
Nb	11	3	9	5	27	18	13	12	8	7	16	17
Rb	0	0	0	2	66	140	0	0	16	0	0	125
Sr	505	375	265	240	31	28	350	155	450	380	165	195
Y	30	14	33	9	75	61	43	24	36	26	41	32
Zr	190	56	155	39	391	205	170	150	145	110	205	225
Co	26	139	40	42	2	1	49	25	31	37	38	14
Cr	76	350	140	445	7	4	29	74	150	235	69	41
Cu	13	190	76	25	1	8	110	21	58	10	160	26
La	24	5	8	5	80	44	10	21	22	3	9	28
Ni	33	78	46	140	15	11	46	39	71	79	68	37
V	180	220	270	205	2	0	425	295	225	295	385	69
Yb	2	1.3	2.5	0.8	7.3	5.3	3.5	2	2.5	2	2.6	2.9
Zn	99	99	101	63	12	24	210	105	145	160	280	71

83042 basalt tuff, Hardy Island
83125 basalt, Browns Flat
83117 basalt, Greenwich Hill
83118 basalt, Greenwich Hill
79066 rhyolite, Highway 111
82156 rhyolite, Musquash Harbour
82107 basalt, Emerson Creek
80138 basalt, Saint John Dry Dock
82064 basalt, Martello Tower
82134 basalt, West Beach
83221 basalt, Sheldon Point
80134 rhyolite, West Beach

Grid references to NTS sheets 21G/8 and 21H/4

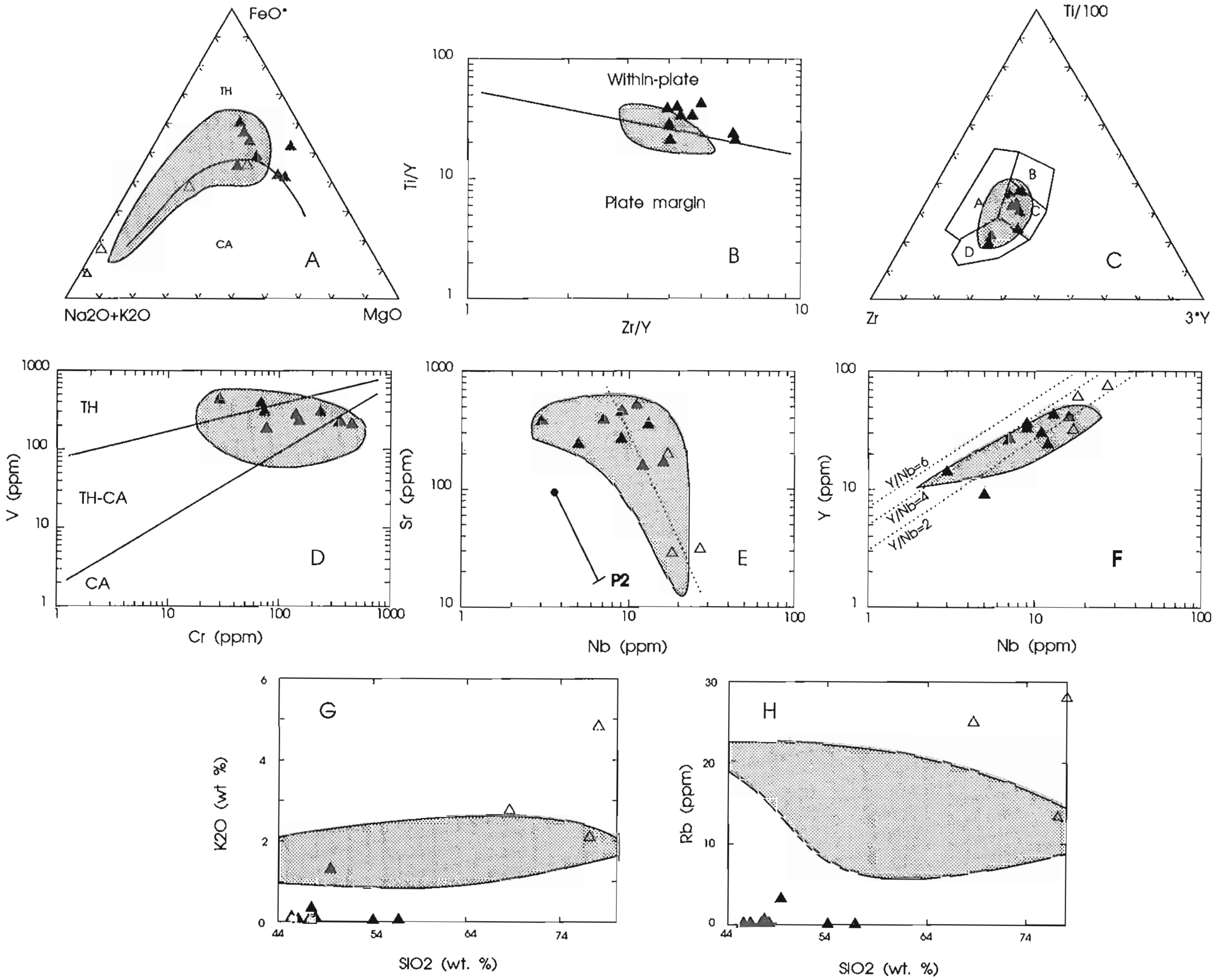

Figure 7. Geochemical discriminators for volcanic rocks of the "Lorneville beds". Analyses from the Coldbrook Group fall in the shaded region

(A) AFM plot for volcanic rocks of the "Lorneville beds". Divider between tholeiitic (TH) and calc-alkaline (CA) fields after Irvine and Baragar (1971).

(B) Zr/Y versus Ti/Y plot for basaltic rocks of the "Lorneville beds". Dividers from Pearce and Gale (1977).

(C) Ti/100-Zr-3Y plot for basaltic ($SiO_2$7%) rocks of the "Lorneville beds". Dividers after Pearce and Cann (1973). A - within plate tholeiites; B - MORB; C - MORB and volcanic arc basalts; D - volcanic arc basalts and within plate tholeiites

(D) V versus Cr (ppm) plot for basaltic rocks (7% SiO_2) of the "Lorneville beds". Dividers from Miyashiro and Shido (1975).

(E) Plot of Sr vs. Nb. Solid line shows trend of Rayleigh fractionation for plagioclase with D_{SR}=6. Length of the line corresponds to 20% crystals removed. The line is parallel to the apparent differentiation trend (dashed line) suggesting plagioclase fractionation caused most compositional variation

(F) Plot of Y vs. Nb. Y/Nb varies from mantle-like ratios of 6 to crustal-like ratios of (Taylor and McLennan (1985) for both basaltic and rhyolitic rocks, suggesting crustal contamination of both compositions.

(G) Plot of K_2O vs. SiO_2

(H) Plot of Rb vs. SiO_2

basalt to the east is faulted at this locality. The nature of the contact between the basalt and the juxtaposed Carboniferous Lancaster Formation was discussed by Currie (1988).

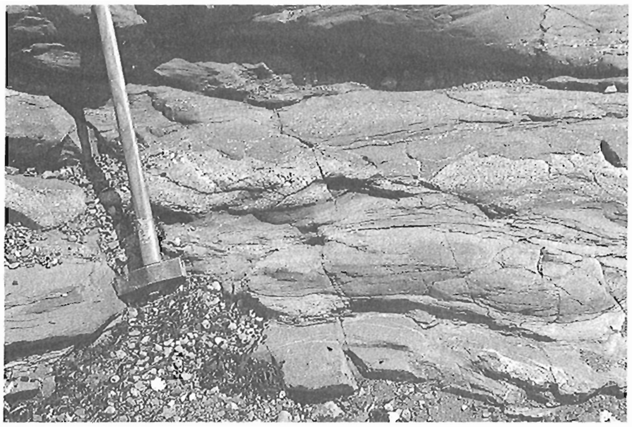

Figure 8. Interbedded green siltstone (below) and red siltstone (above), with lens of pebble conglomerate (L3), Taylors Island. Note channelling and crack-filling at the top of the conglomerate. (1991-035F)

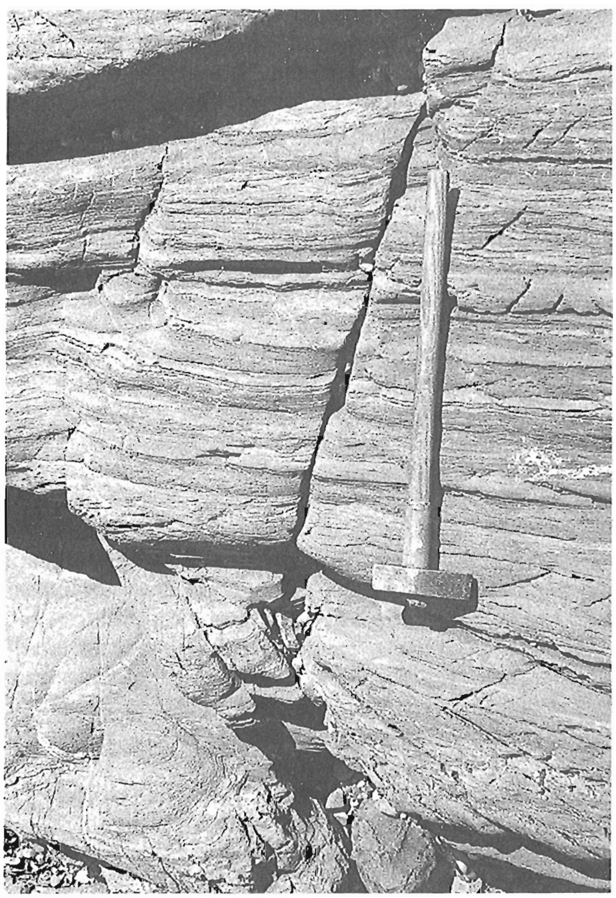

Figure 9. Finely laminated shale and siltstone (L3), Taylors Island. (1991-034Q)

Similar relations are exposed in West Saint John. On Lancaster Street the Ratcliffe Brook Formation is missing, and the Glen Falls Formation can be seen to rest directly on greenish salic tuffs and red granule conglomerate (L1, L2). Further south at Dufferin Street, thinly bedded green siltstone rests on massive basalt (L2). The same relations, now poorly exposed, but well exposed during road repairs in 1983-84, can also be observed a few hundred meters to the west in road cuts on Saint John Street.

The units described at the Dry Dock and in West Saint John, namely red conglomerate, grading to or containing tuffs which may be either mafic or salic, overlying thinly bedded siltstones, and massive basalts with intercalated red sandstone-siltstone layers, can be observed at Sheldon Point, Taylor's Island and along the beach at Lorneville Harbour. The importance of the last exposures lies in their continuity, which demonstrates that these units in fact form a continuous sequence whose stratigraphic relations are fixed by the Dry Dock and West Saint John exposures. The top of the sequence in Lorneville Harbour is covered by conglomerate of the Carboniferous Balls Lake Formation to the north and faulted against Precambrian Coldbrook Group volcanics to the south, but there is a marked difference in degree of alteration and deformation between the "Lorneville beds" and the older Coldbrook Group.

Relations between the Saint John Group and the underlying "Lorneville beds" can be observed 6 km southwest of Lorneville in rock cuttings around the Coleson Cove power station, and similar strata are exposed as far west as Musquash Head.

In the Forest Hills-Mystery Lake area, pebble quartzite of the Glen Falls Formation of the Saint John Group rests directly on red tuffaceous beds. The exposed contact south of Mystery Lake has been tectonically disturbed, but typical red conglomerate (L1) is present a few hundred meters further north. On the north side of the Saint John syncline on Somerset Street in the city of Saint John, at the locality

Figure 10. Soft-sediment deformation in thick-bedded shale and sandstone (L3), Lorneville Harbour. The fault truncating the siltstone bed in the centre does not offset the siltstone at hammer. Note large clast at base of the hammer. (1991-035E)

mentioned by Tanoli and Pickerill (1988), both the Saint John Group and underlying rocks have been disturbed by faulting. About 2 m of red conglomerate are present on the west side of the road cut, juxtaposed against Ratcliffe Brook Formation. Tuffaceous rocks of the "Lorneville beds" are exposed on the east side of the road, and along Burpee Avenue to the east.

A section of >10 m of conglomerate and red sandstone (Fig. 4) occurs at Greenwich Hill north of the Long Reach (Fig. 1) where it underlies Glen Falls orthoquartzite and flaggy sandstone of the Ratcliffe Brook Formation, and overlies a badly fractured and altered basaltic rock presumed correlative to the Coldbrook Group. The red conglomerate (L1) is well exposed in road cuts along Highway 102, and the contact with an outlier of the Saint John Group is exposed at two places in creeks to the south. The volcanic interval (L2) is spectacularly exposed on the shore of the Long Reach at Beulah Camp, but relations can only be inferred. Thinly layered siltstone and sandstone (L3) are well exposed in rock cuts along the abandoned railway line, which follows the shore of the Long Reach. This interval here contains two basalt flows 1-2 m thick. The lower contact is exposed in a rock cutting. As in Hanford Brook, there is a strong contrast in alteration and degree of fracturing, but no clear unconformity. Similar relations can be inferred along Campbell Road, about 10 km to the west although the actual contacts are covered. At this locality "Lorneville beds" rest on altered salic igneous rocks (Woodman Point porphyry).

In addition to these sections, whose identity is fixed by relations to easily recognized units, numerous fault slivers of red conglomerate probably assignable to the "Lorneville beds" are known. Two easily accessible examples occur in the parking lot of the Sobey's store in Grand Bay, and along Highway 177 near Lingley.

DISCUSSION

Some early authors such as Matthew (1889) and Bailey and Matthew (1872) suggested continuous gradation from volcanic rocks poor in sedimentary strata (presently termed Coldbrook Group) through mixed sedimentary and volcanic strata ("Lorneville beds") to fossiliferous sedimentary rocks (the present Saint John Group). However Hayes and Howell (1937) and Alcock (1938) elected for a twofold division with the Saint John Group resting unconformably on Coldbrook Group. This terminology left no place for "Lorneville beds" which were forced into one or the other of Saint John and Coldbrook groups, or swept up into poorly defined catchalls like the Mispec Group (compare Currie and Nance, 1983). Currie and Nance (1983) and Currie (1984) pointed out that separating out these strata simplified the stratigraphy of the Saint John area.

Tanoli et al. (1985) summarised methods of distinguishing these strata from the Saint John Group. One criterion proposed by Tanoli et al. (1985) has proved to be unreliable, namely presence of muscovite in sandstone, thought by them to be characteristic of the Ratcliffe Brook Formation. Muscovite-bearing sandstones occur locally in rocks of the "Lorneville beds" below the Ratcliffe Brook Formation, and in even older sandstones probably correlative to the Coldbrook Group (Currie, 1991), while some sandstones within the Ratcliffe Brook Formation do not contain muscovite. The source of the detrital muscovite is presently unknown.

Rocks of the "Lorneville beds" can be distinguished from the underlying Coldbrook Group by their relatively high proportion of sedimentary rocks, and strong red coloration. Sedimentary rocks within the Coldbrook Group are mainly thinly laminated, black and white cherty siltstones. Such rocks are not found in "Lorneville beds". Rocks of the Coldbrook Group tend to lack obvious primary structure, and fracture into small angular chips, whereas primary structure and foliation can commonly be discerned within the "Lorneville beds".

Alcock (1938), Rast et al. (1978) and some current workers considered the strata at Lorneville to be of Carboniferous age. Currie and Nance (1983) pointed out that known Carboniferous strata along the Bay of Fundy comprised Westphalian redbeds of the Balls Lake Formation, and dark lithic sandstone of the Lancaster Formation. The Balls Lake Formation, consists of lenses of conglomerate, rarely more than 30 cm thick, in a matrix of red shale. Mudstones of the "Lorneville beds" are commonly chloritic, whereas the Balls Lake Formation mudstones are not. The Balls Lake Formation contains abundant vein quartz and some granite cobbles, neither of which are present in the volcanic-dominated clasts of the "Lorneville beds". No volcanic rocks occur within the known Carboniferous rocks of the Saint John region, and no dark lithic sandstones like Lancaster Formation occur in the "Lorneville beds".

If the beds at Lorneville are of Carboniferous age, they must be of quite different (allochthonous) derivation from the Balls Lake and Lancaster Formations (compare Currie and Nance, 1983), and a major fault must separate this Carboniferous portion from the beds at Saint John Dry Dock and West Saint John which are firmly attached to the base of the lower Cambrian section. No mapping has yet identified such a fault, and the consistency of lithologic character and sedimentary structures, including facing, strongly suggest a single section of latest Precambrian age. However this hypothesis also faces objections, notably that "Lorneville beds" are less altered and deformed than the Saint John Group, and that the distinctive massive to pillowed basalt unit which is prominent along the Fundy coast does not appear inland. The first objection is not serious since it could depend on lithologic differences, including the presence of a stiff layer of basalt which shielded weaker lithologies from deformation. A variety of replies can be made to the second objection, notably that the lower portion of the "Lorneville beds" is rarely exposed inland due to faulting and other causes, and that where it is exposed at Greenwich Hill a basalt section, admittedly thin but with appropriate chemistry, is present. However such a section is not present in Hanford Brook. If the whole section is of latest Precambrian age, the basalt unit must thin rapidly inland.

East of Saint John, massive altered basalt stretches as a thin discontinuous band along the Bay of Fundy for 50 km or more, including such units as the West Beach Formation of Alcock (1938) but correlation of these rocks with the "Lorneville beds" is subject both to the objections noted above, and great uncertainty due to structural complexity. West and north of Saint John, Currie (1987) correlated voluminous salic volcanic rocks and intercalated red sandstone and conglomerate with "Lorneville beds", and recent dating (Zain Eldeen et al., 1991) shows that they are of latest Precambrian age (555 Ma).

STRATIGRAPHIC AND TECTONIC SIGNIFICANCE OF THE LORNEVILLE GROUP

The Saint John region exhibits an essentially continuous stratigraphic record across the paleontologically defined Cambrian-Precambrian boundary. The lower part of this record, the Coldbrook Group, formed in a volcanic arc with continental basement (Currie and Eby, 1990). The upper part (Saint John Group) formed during a marine transgression (Tanoli and Pickerill, 1988). The Lorneville Group represents a transition from one to the other during which conditions were terrestrial to littoral. This quasi-continuous stratigraphic record, including Coldbrook and Saint John Groups and "Lorneville beds" contrasts with the classic views of Alcock (1938) and Hayes and Howell (1937) who proposed a major discontinuity between Precambrian and Paleozoic strata. However, as in many proposed revisions, the spirit of the proposal was clearly defined in the writings of pioneer geologists such as Bailey and Matthew (1872) and Matthew (1889).

The Coldbrook-"Lorneville beds"-Saint John sequence occurs on both sides of the "Brookville terrane", as defined by Barr and White (1989), and as slivers along the bounding faults. A simple explanation of such relations is that the sequence originally also covered the terrane between the slivers, from which it has been removed by uplift and erosion.

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Geochemistry of late Proterozoic plutonic rocks from Flemish Cap, east of the Grand Banks of Newfoundland

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Abstract

Whole rock geochemical analyses of drill core samples of granodiorite from Flemish Cap show that they are typical subduction-related granitoid rocks. They closely resemble latest Proterozoic granitoid plutons of the Avalon Terrane of northern Nova Scotia and southeastern Newfoundland. A reconsideration of the existing radiometric age data is also consistent with such a correlation of the Flemish Cap granodiorite with the Avalon Terrane on land.

Résumé

Les analyses géochimiques de la roche entière, sur des échantillons de carottes de sondage prélevées dans le granodiorite du secteur de Bonnet Flamand, montrent qu'il s'agit d'une roche granitoïde typique dont la formation est liée à un épisode de subduction. Elles ressemblent fortement aux plutons granitoïdes de la toute fin du Protérozoïque, situés dans le terrane d'Avalon, dans le nord de la Nouvelle-Écosse et dans le sud-est de Terre-Neuve. Un nouvel examen des données existantes sur la datation radiométrique semble également confirmer cette corrélation du granodiorite de Flemish Cap (Bonnet Flamand) avec le terrane d'Avalon sur les terres émergées.

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INTRODUCTION

Bedrock outcrops on Flemish Cap described by Pelletier (1971) and King et al. (1985) appear on the basis of petrography to correlate with late Proterozoic Avalon zone rocks that outcrop in southeastern Newfoundland and northern Nova Scotia.

King et al. (1985) described the petrography of granodiorite, dacite and volcanic siltstone in drill core samples from bedrock outcrops, and interpreted the granodiorite as forming a single pluton at least 65 km in diameter. Pelletier (1971) obtained a K-Ar biotite age of 592 ± 20 Ma (recalculated to 615 ± 20 Ma using constants recommended by Steiger and Jager, 1977) from the granodiorite (P in Fig. 1). King et al (1985) dated zircons from granodiorite in core 15 (Fig. 1). They obtained highly discordant U-Pb spectra which they interpreted as representing an intrusion age in the range 750-830 Ma. The same authors obtained K-Ar ages of 657 ± 29 Ma on hornblende from core 15, 505 ± 28 Ma from core 12, and 619 ± 34 Ma from core 1 (Fig. 1). These dates were considered by King et al. (1985) to represent incomplete degassing during the superimposed subgreenschist metamorphism of uncertain age. Their proposed intrusion age 750-830 Ma is much older than those reported for petrographically similar Hadrynian rocks of eastern Newfoundland (Krogh et al., 1988) and northern Nova Scotia (Donohoe and Wallace, 1982).

The purpose of this note is to report the whole rock geochemistry of the Flemish Cap granodiorite cores and to compare this with possible correlative rocks onshore. The Holyrood plutonic series of Newfoundland (Strong and Minatidis, 1975, for which only summary geochemical plots are available) and the Jeffers Brook and Frog Lake diorites of northern Nova Scotia (Pe-Piper, 1988 and unpublished data) are used as comparative Avalon terrane rocks.

This study forms part of a broader investigation of the petrology, geochemistry and regional significance of basement rocks on the Scotian Shelf and the Grand Banks of Newfoundland, partly funded by EMR Research Agreements.

GEOCHEMISTRY

The samples analyzed (Table 1) are from the cores 2, 12B, 15, 18 and 20 of King et al. (1985). Major and trace elements were determined by X-ray fluorescence and REE by instrumental neutron activation analysis. Analytical methods are described by Pe-Piper and Piper (1989).

Petrographically, all the granodiorite samples are similar, as previously noted by King et al (1985). They are fine grained to coarse grained untectonized equigranular rocks, consisting of variable amounts of plagioclase, K-feldspar, quartz, hornblende, biotite and opaque oxide minerals. In some samples the quartz forms a graphic texture with the K-feldspar which may be common or only local.

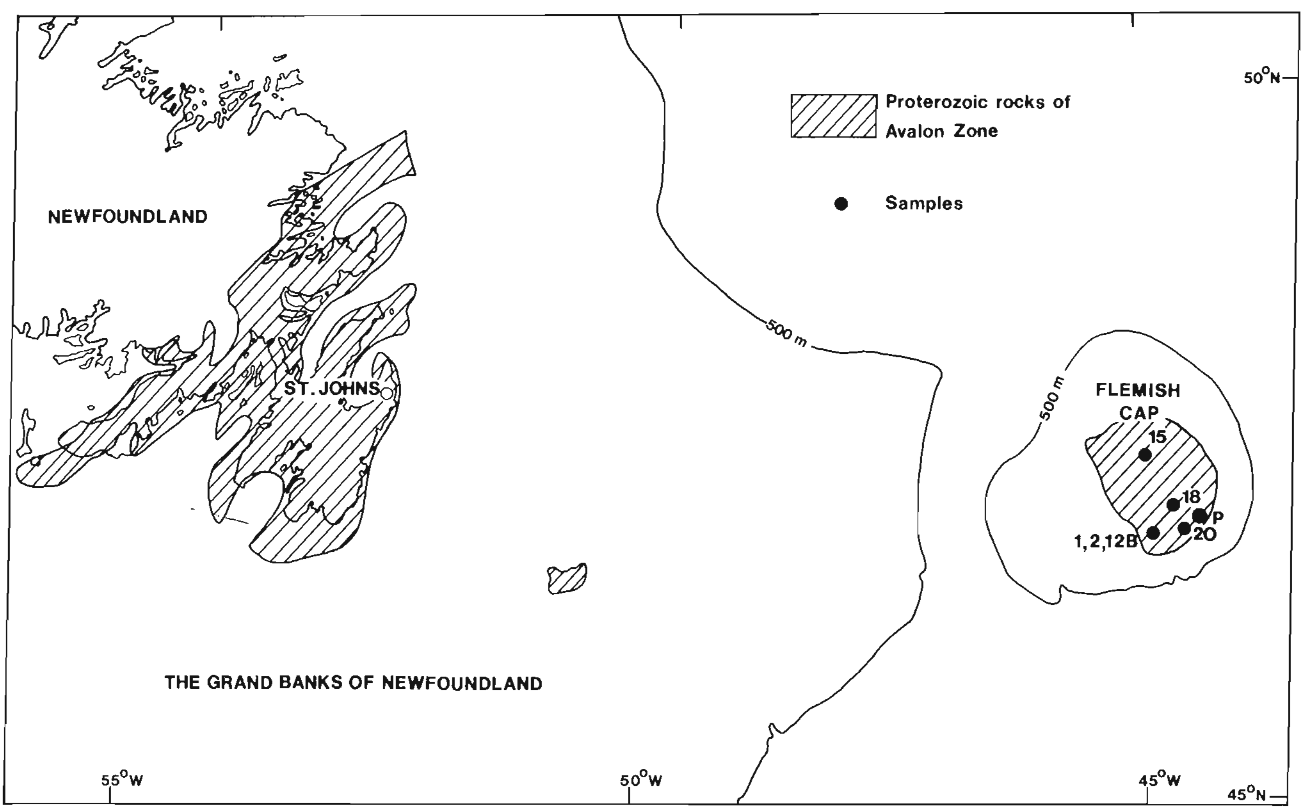

Figure 1. Location map for Flemish Cap showing distribution of Avalon Terrane late Proterozoic rocks (from Fader et al., 1989) and location of drill core samples (from King et al., 1985). P = sample site of Pelletier (1971).

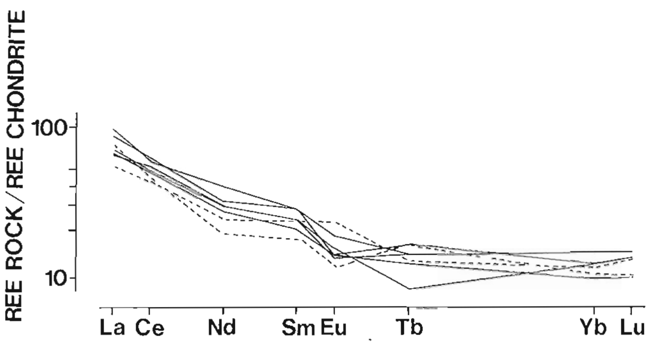

Figure 2. Chondrite-normalized REE patterns for the Flemish Cap granitoid rocks. Samples from the Jeffers Brook diorite of northern Nova Scotia are shown for comparison (dashed lines).

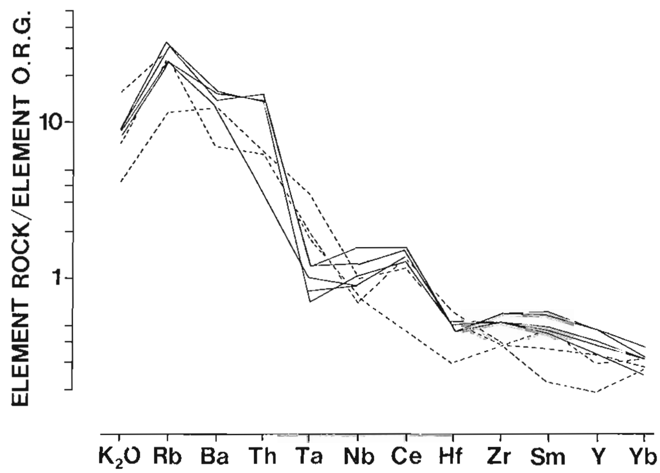

Figure 3. Ocean ridge granite (O.R.G.) normalized patterns of selected elements for granitoid rocks from Flemish Cap. Samples from the Jeffers Brook diorite of northern Nova Scotia are shown for comparison (dashed lines). O.R.G. normalizing values after Pearce et al. (1984).

Table 1. Whole rock analyses of granitoid rocks from Flemish Cap

Sample No.	18-76	18-8	18-53	15-66	2-8	18-85	12-151	20-25	12-43
Major elements (wt%) by XRF									
SiO_2	63.58	64.22	66.88	66.94	67.49	67.54	68.03	68.56	69.57
TiO_2	0.63	0.62	0.49	0.46	0.66	0.44	0.66	0.44	0.56
Al_2O_3	15.19	15.19	14.94	15.20	14.64	15.16	14.29	14.61	13.92
Fe_2O_{3t}	5.13	4.96	3.88	3.78	3.88	3.47	3.70	3.22	3.37
MnO	0.11	0.10	0.08	0.07	0.10	0.07	0.10	0.06	0.08
MgO	2.58	2.60	1.85	1.69	1.59	1.85	1.47	2.00	1.38
CaO	3.85	3.63	3.05	2.66	2.31	3.05	2.03	2.77	1.60
Na_2O	3.58	3.69	3.59	4.13	4.28	3.71	4.33	3.99	3.92
K_2O	3.33	3.15	3.71	3.46	3.44	3.65	3.53	3.16	3.74
P_2O_5	0.13	0.13	0.10	0.10	0.16	0.09	0.15	0.10	0.13
L.O.I.	1.10	0.90	0.80	0.70	0.80	0.60	0.90	0.50	0.90
Total	99.21	99.19	99.37	99.19	99.35	99.63	99.19	99.41	99.17
CIPW norms (wt%)									
Qz	15.70	16.71	20.66	19.44	20.31	20.97	21.22	22.82	25.58
Or	20.16	19.04	22.33	20.84	20.73	21.86	21.30	18.94	22.57
Ab	31.04	31.93	30.94	35.62	36.89	31.81	37.42	34.25	33.87
An	15.93	15.93	13.95	12.96	10.76	14.12	9.33	12.80	7.45
Hy	13.14	13.37	10.13	9.98	9.60	9.53	8.98	9.49	8.40
Di	2.54	1.54	0.83	0.07	0.07	0.69	0.14	0.68	-
Il	1.23	1.20	0.95	0.89	1.28	0.85	1.28	0.85	1.09
Ap	0.31	0.31	0.24	0.24	0.38	0.21	0.36	0.24	0.31
Cor	-	-	-	-	-	-	-	-	0.76
Trace elements (ppm) by XRF									
Ba	799	643	535	778	753	582	667	1004	742
Rb	102	93	103	98	128	98	123	82	132
Sr	343	333	269	293	247	275	218	429	181
Y	27	29	29	23	33	26	33	14	33
Zr	183	190	163	181	208	145	207	112	186
Nb	9	9	8	10	12	9	15	10	13
Th	13	3	12	11	11	9	12	11	12
Pb	2	-	6	13	10	-	5	-	18
Ga	19	18	16	16	19	16	19	17	16
Zn	58	57	48	48	83	39	59	30	53
Cu	-	3	3	-	-	-	-	-	-
Ni	12	12	8	11	10	9	8	15	8
V	95	89	59	46	38	52	41	55	36
Cr	40	41	37	40	33	36	33	44	39
REE and other trace elements (ppm) by INAA									
La	22.80	23.80	n.d.	22.90	31.80	n.d.	29.30	n.d.	n.d.
Ce	49.00	47.00	n.d.	44.00	52.00	n.d.	54.00	n.d.	n.d.
Nd	18.00	18.00	n.d.	17.00	25.00	n.d.	20.00	n.d.	n.d.
Sm	4.38	4.42	n.d.	3.95	5.23	n.d.	5.37	n.d.	n.d.
Eu	0.96	1.10	n.d.	0.96	1.03	n.d.	1.30	n.d.	n.d.
Tb	0.70	0.40	n.d.	0.60	0.80	n.d.	0.70	n.d.	n.d.
Yb	2.45	2.52	n.d.	1.96	2.45	n.d.	2.92	n.d.	n.d.
Lu	0.45	0.46	n.d.	0.34	0.45	n.d.	0.51	n.d.	n.d.
Cs	2.20	2.20	n.d.	1.30	5.30	n.d.	5.30	n.d.	n.d.
Hf	4.10	4.80	n.d.	4.60	4.10	n.d.	4.10	n.d.	n.d.
Sb	0.30	0.20	n.d.	0.10	0.40	n.d.	0.40	n.d.	n.d.
Sc	12.80	12.20	n.d.	8.00	10.60	n.d.	11.00	n.d.	n.d.
Ta	0.60	0.70	n.d.	0.50	0.80	n.d.	0.80	n.d.	n.d.
Th	9.50	9.30	n.d.	9.10	9.30	n.d.	9.30	n.d.	n.d.
U	3.50	3.90	n.d.	2.50	2.10	n.d.	2.10	n.d.	n.d.

The accessory minerals vary from core to core and include apatite, zircon, sphene and rutile. The rocks are variably altered which is shown distinctively by chloritization of biotite, sericitization of feldspar, hematitic stain and by minor amounts of epidote and calcite. Actinolitic overgrowths or patches on hornblende have also been seen. Cores 2 and 12B are more altered than the other cores.

The Flemish Cap plutonic rocks follow a trend from quartz monzodiorite through granodiorite to granite (chemical terminology after Streckeisen and LeMaitre, 1979). Both the quartz monzodiorites and granodiorites occur in core 18, which is the longest of the suite of cores. With increasing SiO_2 there is a decrease of Al_2O_3, Fe_2O_{3t}, MgO and V. All other major and trace elements are either almost constant (e.g. K_2O) or show a scatter.

The REE patterns (Fig. 2) show a moderate REE enrichment which increases with increasing SiO_2 content (up to around 100 times chondrite). Eu anomalies are either small or lacking.

Plots of elements normalized to abundance in ocean-ridge-granite (ORG) after Pearce et al. (1984) (Fig. 3) show strong enrichment in K, Rb, Ba and Th (except one sample) and relative enrichment in Ce compared with Nb, Zr, Sm and Y. Overall, the geochemical patterns of these granitoid rocks show element distributions similar to those of volcanic arc granites. On the Rb-(Y+Nb) discriminant diagrams of Pearce et al. (1984) (not illustrated) these rocks plot on the upper part of the volcanic arc field. They also plot in the volcanic arc field on the Rb/30 - Ta x 3 - Hf discriminant diagram of Harris et al. (1986) (not illustrated).

Covariation of pairs of incompatible elements indicates normal igneous fractional crystallization trends. Variation of Ba with Rb indicates K-feldspar and/or biotite fractionation in the evolution of the more felsic from the more mafic phases. Variation of Ba with Sr and Zr with Nb shows the importance once again of K-feldspar fractionation, together with hornblende and/or biotite.

DISCUSSION

In whole rock geochemistry, the granodiorites from Flemish Cap appear to be typical volcanic arc granitoids. Their element variation is similar to that shown in plots for the Holyrood granitoids by Minatidis and Strong (1985) and to granodiorites of late Proterozoic age from northern Nova Scotia, such as the Jeffers Brook pluton (Fig. 2, 3). There is no reason to suppose that the highly discordant zircon date of King et al. (1985) provides a basis for estimating intrusion age of the Flemish Cap granodiorite. Neglecting the date from altered sample 12B, the conventional K-Ar hornblende dates show a scatter and analytical error similar to that in similar dates from Avalon terrane plutons in northern Nova Scotia (e.g. as summarized by Donohoe and Wallace, 1982), where more precise $^{39}Ar/^{40}Ar$ hornblende dates suggest an intrusive age of 600-620 Ma (Keppie et al., 1990) that is also supported by U-Pb zircon dates (Doig et al., in press). The Holyrood granite has also yielded a U-Pb zircon age of about 620 Ma (Krogh et al, 1988). Thus both the geochemical data and the previous imprecise radiometric dating are consistent with a correlation of the Flemish Cap granodiorite with the latest Proterozoic subduction-related plutons of the Avalon terrane.

ACKNOWLEDGMENTS

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Uranium in Canada, 1991

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Ruzicka, V., 1992: Uranium in Canada, 1991; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 49-57.

Abstract

During 1991 Canada's uranium resources in quartz-pebble conglomerates decreased, but the resources in unconformity-associated deposits increased due to exploration of the P2 North and Sue deposits in the Athabasca Basin, Saskatchewan. The P2 North deposit represents a new sub-type in the class of the unconformity deposits. The Sue and the Eagle Point deposits indicate a consanguinity of the basement-hosted monometallic and the polymetallic deposits associated directly with the sub-Athabasca unconformity.

Résumé

En 1991, les ressources du Canada en uranium contenu dans les conglomérats à galets quartzeux ont diminué, mais les ressources présentes dans les gîtes associés à des discordances ont augmenté, grâce à la prospection des gîtes de P2 North et de Sue dans le bassin d'Athabasca en Saskatchewan. Le gisement de P2 North représente un nouveau sous-type dans la classe des gisements liés à une discordance. Le gisement de Sue et le gisement d'Eagle Point indiquent la consanguinité des gisements monométalliques contenus dans le socle et des gisements polymétalliques directement associés à la discordance sub-Athabasca.

INTRODUCTION

In 1991 the Canadian nuclear energy mineral resources underwent significant changes. On one hand, continuing exploration led to an increase of high grade uranium resources; on the other hand, the recent market conditions caused some reduction in recoverable low grade resources.

As of January 1991 the Canadian uranium resources, recoverable at prices up to $C100/kg U, amounted to 145.7 kilotonnes of uranium in the Reasonably Assured (Measured and Indicated resources) category and 149.1 kilotonnes in the Estimated Additional I (Inferred resources) category.

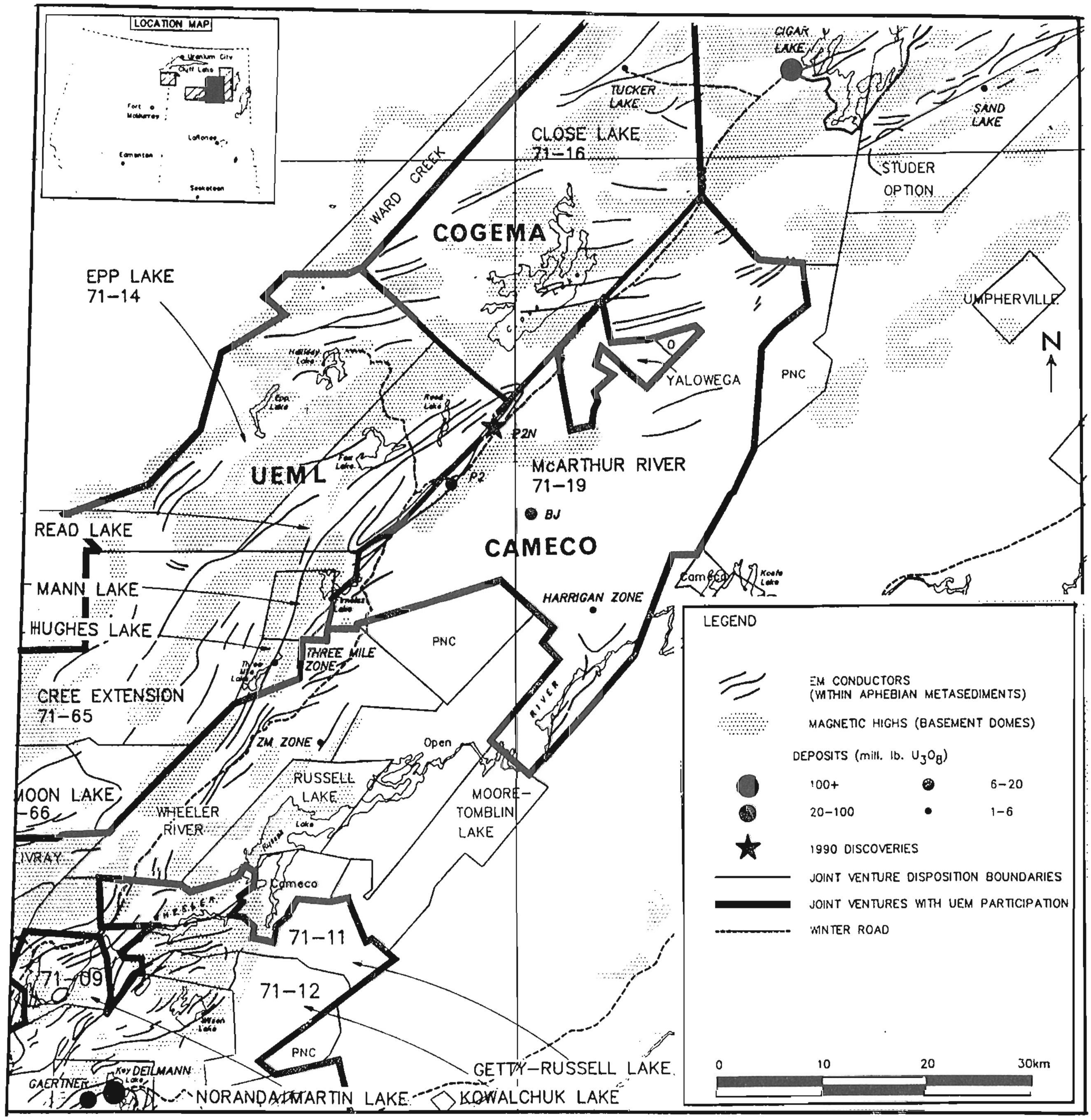

Figure 1. Claim blocks (including names from Mining Recorder's records) in the southeastern part of the Athabasca Basin, Saskatchewan. Operators of joint venture exploration projects for selected claim blocks: CAMECO - Canadian Mining and Energy Corporation, COGEMA - Cogema Canada Limited, UEML (UEM in the Legend) - Uranerz Exploration and Mining Limited. Modified after a disposition map, courtesy of Uranerz Exploration and Mining Limited.

Canada continued holding the world's leading position in production of uranium although its uranium output has been reduced from some 11 300 tonnes (1989) to about 8750 tonnes U (1990) and to about 8000 tonnes in 1991. The major portion of the production (about two thirds) is from the unconformity-associated deposits in the Athabasca Basin, Saskatchewan, and the rest from the quartz-pebble conglomerates at Elliot Lake, Ontario.

The exploration, with expenditures of about 45 million dollars in 1991, was focused particularly on areas favourable for occurrence of deposits associated with the Proterozoic unconformities in the Athabasca and the Thelon basin regions. Grass-roots exploration (reconnaissance) was conducted for the breccia-complex deposits in the Bear Structural Province. Advanced underground exploration took place at the Cigar Lake deposit.

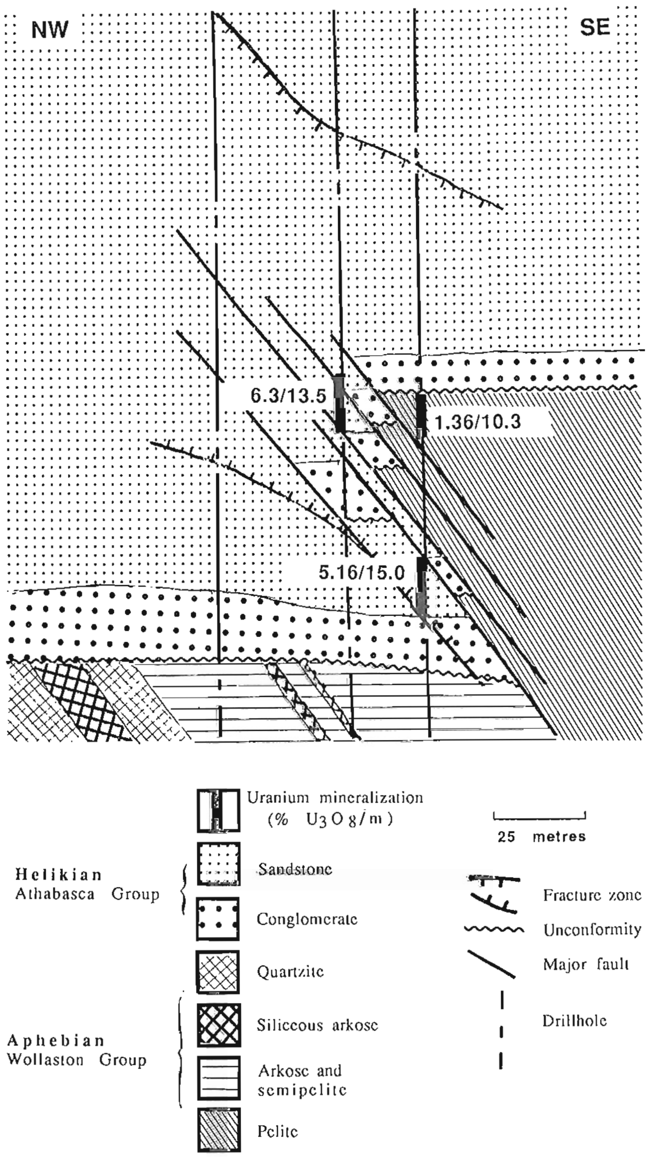

Figure 2. A schematic cross-section through the P2 North deposit, Saskatchewan. Modified after Marlatt et al. (in press).

The exploration in the Athabasca Basin region was concentrated: (i) in the Cigar Lake-Key Lake structural trend and its northeastern extension; (ii) in the Aphebian and Helikian rocks mantling the Collins Bay Granite Dome; and (iii) in the Carswell Structure and the Firebag and Clearwater structural domains. It was based on a conceptual metallogenic model postulating association of the deposits with the sub-Athabasca unconformity, Archean granitic domal structures, Lower Proterozoic euxinic rock horizons, and faults or fracture zones in the basement and cover rocks intersecting the unconformity.

The exploration in the Thelon Basin region was conducted in the Kiggavik structural trend, which extends from the Kiggavik deposit to the southwest. It was based on a conceptual metallogenic model postulating association of the deposits with the sub-Thelon unconformity.

In the Northwest Territories, the Geological Survey of Canada continued to contribute to the development of exploration strategy for polymetallic deposits in the Great Bear Magmatic Zone by a refined conceptual metallogenic model for the Zone, based on an analogy with the Australian Olympic Dam-type of copper-uranium-gold mineralization.

Mineral exploration of the McArthur River and Wolly blocks in the Athabasca Basin, and of the Kiggavik trend in the Thelon Basin region led to the increase of Canadian resources in high grade ores to such an extent that they more than compensated for the decrease of low grade resources in the Elliot Lake area, Ontario.

Advanced exploration of the Cigar Lake deposit continued with preparation of the geotechnical tests for establishing suitable mining methods for exploitation of the very high grade pitchblende orebody, which is surrounded by clay and unconsolidated sand.

Under the sponsorship of the Natural Sciences and Engineering Research Council (NSERC) and under the GSC liaison, two scientific university-industry projects continued: (i) "Fluid history as a tool for establishing conceptual genetic models for uranium, gold and base metal deposits in Athabasca Basin, La Ronge and Flin-Flon belts, Saskatchewan" was conducted in co-operation of the University of Saskatchewan and Cameco; and (ii) "Metallogeny of the Athabasca Basin based on geochronologic evidence" resumed in co-operation of the University of Alberta and Cameco.

In the course of mineral resource appraisal, key deposits and areas in Canada were selected for detailed studies. Some results of the field, laboratory, and office studies by the author are presented in this paper.

THE CIGAR LAKE - KEY LAKE STRUCTURAL TREND

A major portion of Canadian uranium resources occur in the southeastern part of the Athabasca Basin. The uranium mineralization is controlled by the sub-Athabasca

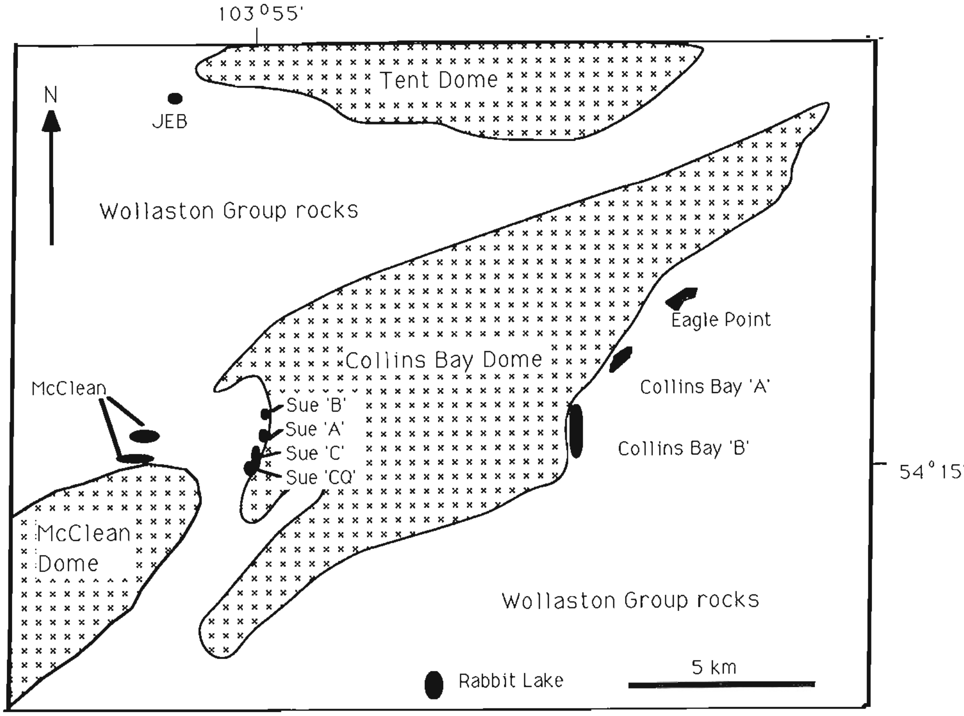

Figure 3. Assumed outline of the paleosurface of the pre-Helikian rocks in the vicinity of the Collins Bay Dome. Modified after Ey et al., (in press).

unconformity and by prominent dislocations, such as Key Lake Fault, Rabbit Lake Fault, Cigar Lake Fault, Collins Bay Thrust, and the P2 Fault.

The P2 Fault is a part of a major northeasterly-striking Cigar Lake-Key Lake structural trend, where two major uranium-polymetallic deposits, the Cigar Lake and the Key Lake, occur. The P2 Fault is a well developed thrust fault on the Cameco's McArthur River block, where it extends along its northwestern boundary (Fig. 1). The thrust fault strikes between 45° and 50°, dips 45° southeast, and displaces the unconformity vertically by 25 to 50 m. The fault controls the location of two deposits: the P2 North, and the P2 Main. Geology and exploration history of the McArthur River block have been recently described by Marlatt et al. (in press). The author had the opportunity to examine the drill core from the deposits on the property and made following observations.

P2 North deposit

The P2 North deposit is located about 70 km northeast of the Key Lake Mine. It contains predominantly monometallic uranium (pitchblende) mineralization. It occurs in the Athabasca Group rocks, just above the sub-Athabasca unconformity and in the footwall of the P2 thrust fault (Fig. 2). The mineralized zone has been traced for 1850 m along the strike by vertical drillholes. It is on average 30 m wide and 7 m thick, but locally more than 50 m wide and its vertical thickness is as much as 46 m. The main orebody occurs from about 500 to about 600 m below the surface. It was tested by vertical drillholes. Marlatt et al. (in press) estimated that the deposit contains in excess of 78 000 tonnes of uranium metal in ores grading 3.4% U. The orebody consists of massive pitchblende and very minor amounts of

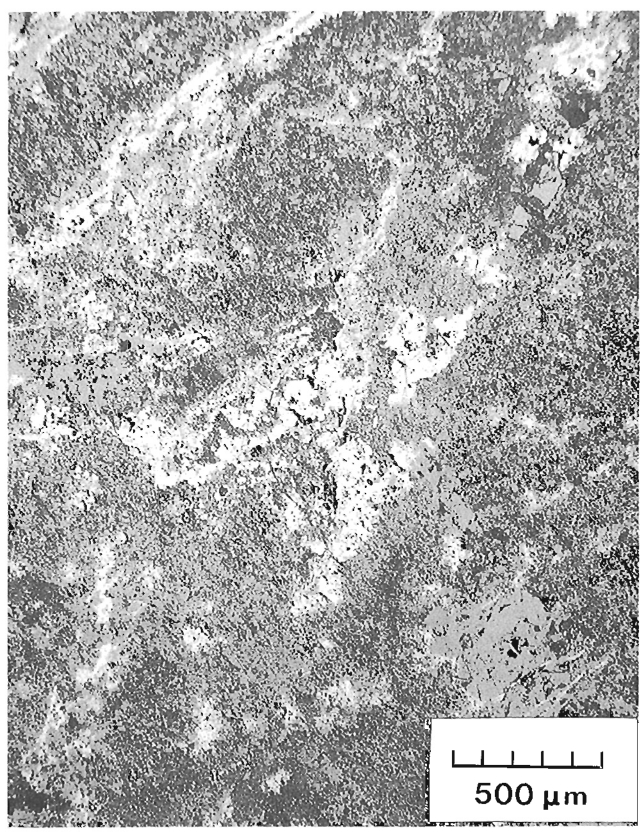

Figure 4. Patches of pitchblende (white) in micaceous matrix (granular gray) surrounding quartz (gray) in pegmatite intercalated with graphitic schist; Eagle Point deposit, Saskatchewan. Drillhole ES 239/119.7 m. Photomicro-graph, reflected light.

galena, pyrite, and chalcopyrite. The basement rocks in the footwall of the orebody consist of quartzite interbedded with garnetiferous and cordieritic gneisses, with a few metres of chloritic and hematitic regolith. The overthrusted basement rocks consist of Aphebian graphitic and sericitic schists, quartzites, and minor amounts of pegmatites and calc-silicates. The basement rocks are unconformably overlain by conglomerate and sandstone of the Helikian Athabasca Group. The host rocks are strongly silicified, but otherwise only relatively weakly altered by illite, chlorite, kaolinite, hematite, limonite, siderite, and dravite. Except for the silicification, the alteration of the deposit is restricted to a narrow aureole around the orebody. The mineralization exhibits two main U-Pb ages: an older age of 1514 ± 18 Ma and a younger age of 1327 ± 8 Ma (Cumming and Krstic, 1991). The older date represents the oldest known mineralization among the deposits associated with the sub-Athabasca unconformity.

P2 Main deposit

The P2 Main deposit (also known as the P2 South Zone) is located about 5 km to the soutwest of the P2 North deposit. The deposit occurs at an intersection of the P2 Fault and a subvertical mylonite zone, 15 to 55 m above the sub-Athabasca unconformity, i.e. from about 520 to about 560 m below the present surface. The principal ore-forming mineral is pitchblende, which occurs in altered arkose and sandstone. The basement rocks include graphitic-pyritic metapelite and calc-silicate rocks. The uranium mineralization of the P2 Main Zone was preceeded by extensive silicification of the Athabasca Group rocks and by tourmalinization (dravitization) of the host rocks. The uranium mineralization is accompanied by kaolinization and hematitization. Limonitization of the host rocks took place during the post-ore stage. The mineralization shows features of redeposition.

Both the P2 North and P2 Main deposits represent a new sub-type of the unconformity-associated deposits: (a) in spite of their occurrence in the sandstone immediately above the unconformity, their mineralization is monometallic; (b) their alteration halo is restricted to a relatively narrow zone; (c) their mineralization is structurally rather than lithologically controlled.

DEPOSITS ASSOCIATED WITH THE COLLINS BAY DOME

The Collins Bay Dome is a northeasterly elongated Archean pluton made up of medium- to coarse-grained biotite granite, foliated granite, and granite gneiss, flanked by Aphebian

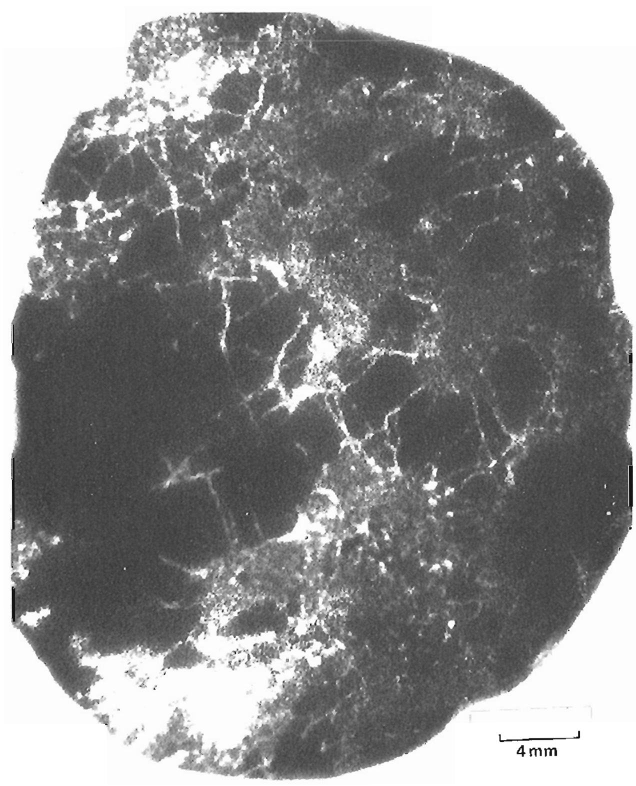

Figure 5. Autoradiograph (positive image) of pitchblende (white) in pegmatite intercalated with graphitic schist. Eagle Point deposit, Saskatchewan; drillhole ES 239/119.7 m.

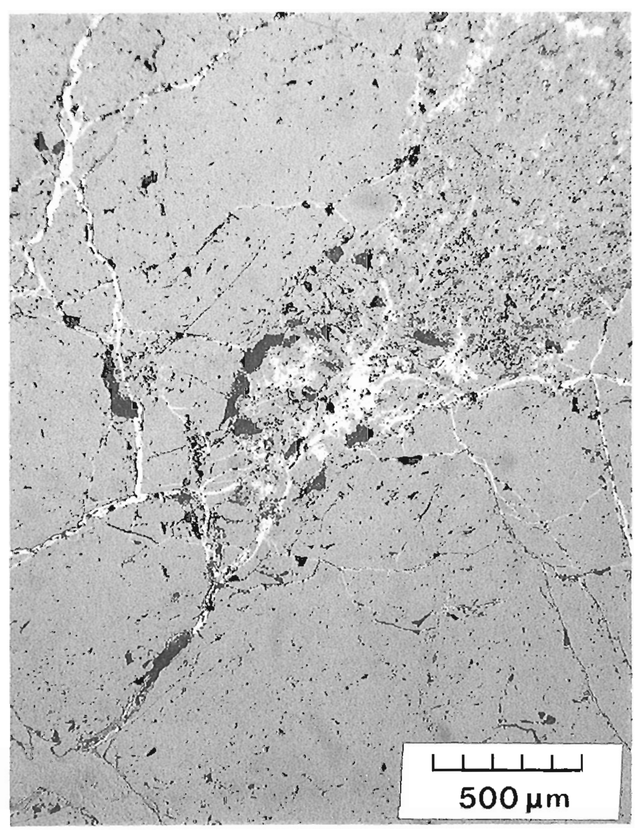

Figure 6. Fractures in massive quartz in pegmatite (gray) invaded by pitchblende (white). Eagle Point deposit, Saskatchewan. Drillhole ES 239/119.7 m. Photomicrograph, reflected light.

metasedimentary rocks, including layers of graphitic gneiss. The pluton is surrounded by faults, such as the Collins Bay Fault, the Eagle Point Fault, the Rabbit Lake Fault, and the Sue Fault that control uranium and uranium-polymetallic mineralization. The Collins Bay 'A', 'B', and 'D' deposits are associated with the Collins Bay Thrust, which extends along the southeastern margin of the Dome. The Eagle Point deposit, which occurs at the northeastern part of the Dome, is controlled by the Collins Bay Thrust and Eagle Point Fault. The Rabbit Lake deposit is structurally controlled by the Rabbit Lake Fault, which occurs near the southeastern margin of the Dome. The Sue Zone, which comprises several deposits, is a part of the north-trending Sue Fault, and terminates the southwestern part of the Dome.

The Sue deposits

A conceptual model, which postulates association of uranium deposits with granitic domes (Ruzicka, 1989) has been reconfirmed in the case of the Sue Zone on the Wolly block (Ruzicka, 1990). The zone is a part of the Sue Fault, a structural trend, which commences at the McClean deposit, and continues for about 2 km to the east, then swings around the Collins Bay Dome and continues in the roughly southern direction for about 3 km (Fig. 3). It is mineralized within its sub-meridional part adjacent to the southwestern margin of the Collins Bay Dome. The trend has a distinct VLF (very low frequency) response, which indicates presence of conductive horizons, represented by a layer of graphitic gneiss within the Aphebian sequence. As of January 1, 1991 the following deposits, associated with the zone, have been identified (from north to south): Sue B, Sue A, Sue C and Sue CQ (Ey et al., in press).

The writer examined the drill core from all the Sue deposits in the summer of 1991 and made following observations:

The Sue 'B' deposit is located about 2.5 km east of the McClean deposit with its top in the Athabasca Group sandstone, about 10 m below the present surface (just below the overburden). The mineralizaton occurs in two lenses within the clay-altered sandstone down to the unconformity, which is about 75 m below the surface. The strongly argillized basement rocks consist of pelites, feldspathic gneiss, and pegmatoids. The deposit contains polymetallic mineralization consisting of uranium oxides and silicates, nickel and cobalt arsenides, and sulpharsenides (altered in the oxidized drill-core to annabergite and erythrite), and sulphides of iron, lead and molybdenum (the latter altered to ilsemannite). Ey et al. (in press) reported following average

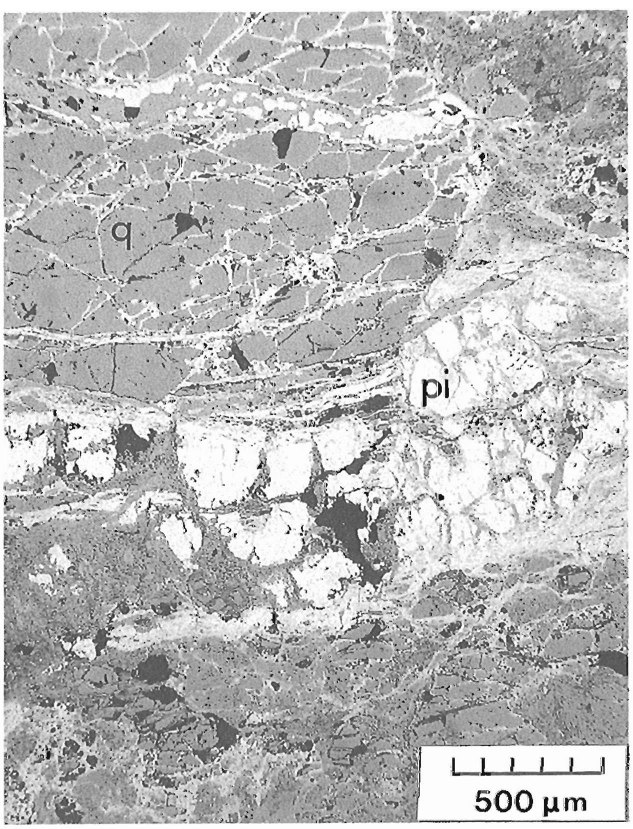

Figure 7. Pitchblende with a very fine crystal structure (pi) forms a patch at the fractured massive quartz (q) in pegmatite. Eagle Point deposit, Saskatchewan, drillhole ES 239/132.0 m. Photomicrograph, reflected light.

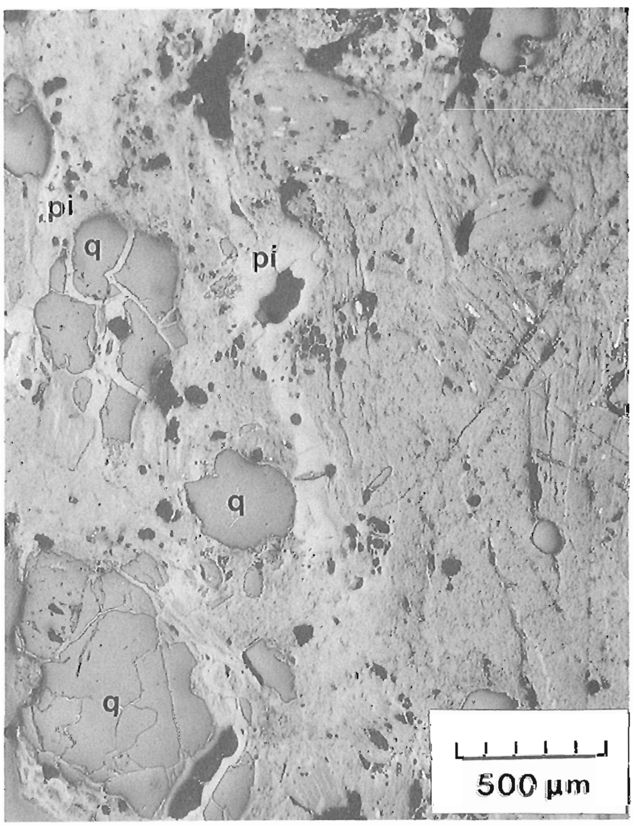

Figure 8. Pitchblende (pi) surrounding and invading corroded and fractured quartz grains (q) in clay-altered paragneiss. Eagle Point deposit, Saskatchewan, drillhole ES 294/117.8 m. Photomicrograph, reflected light. XRF analysis of this specimen is in column 3 of Table 1.

contents of metallic constituents accompanying uranium in an unspecified number of samples from the Sue 'B' deposits: Ni - 1.67%, Co - 0.02%, As - 1.86%, Pb - 0.06%, Mo - 0.01%, Cu - 0.05%, and V - 0.13%.

The Sue 'A' deposit is located about 250 m south of the Sue 'B' and occurs in chloritized, limonitized, and hematitized sandstone at intersections of the Sue Fault with northeasterly-trending faults. The top of the orebody is about 60 m below the present surface (i.e. about 50 m below the overburden). The highest grade ores occur just above the unconformity, which is about 75 m below the surface. The basement rocks in the footwall of the ore body consist of strongly argillized felsic and graphitic gneisses. Pitchblende and nickeline are the principal ore minerals. According to Ey et al. (in press), the Sue 'A' deposit contains non-radioactive metals accompanying uranium: 3.57% Ni, 0.13% Co, 4.46% As, 0.28% V, 0.10% Pb, and 0.03% Cu.

The Sue 'C' deposit is located about 100 m south of the Sue 'B' deposit and occurs only below the unconformity in fractured, argillized and limonitized gneiss at a depth from 90 m down to about 150 m from the present surface. The Athabasca sandstone above the unconformity is strongly kaolinized. The mineralization is monometallic consisting mainly of massive pitchblende, which occurs in several lenses, plunges to south, and transgresses to the Sue "CQ" deposit.

The Sue 'CQ' deposit is located in extension of the Sue 'C' deposit and occurs entirely in the basement rocks, particularly in altered graphitic schist at the contacts with Aphebian quartzite and gneiss. Its principal uranium mineral is disseminated coffinite and locally high grade pods of pitchblende. The uranium mineralization is accompanied by nickel, lead, and vanadium minerals.

The Sue deposits are an excellent example of the consanguinity of the sandstone- and basement-hosted mineralization (i.e. of the polymetallic and monometallic mineralization respectively).

Eagle Point deposit

The Eagle Point deposit occurs in Aphebian rocks of the Wollaston Group near the eastern margin of the Collins Bay Dome (Fig. 3). Geology of the deposit has been described by Andrade (1989). The principal uranium ore mineral is pitchblende; coffinite, boltwoodite and, uranophane occur in smaller amounts. The pitchblende occurs in several

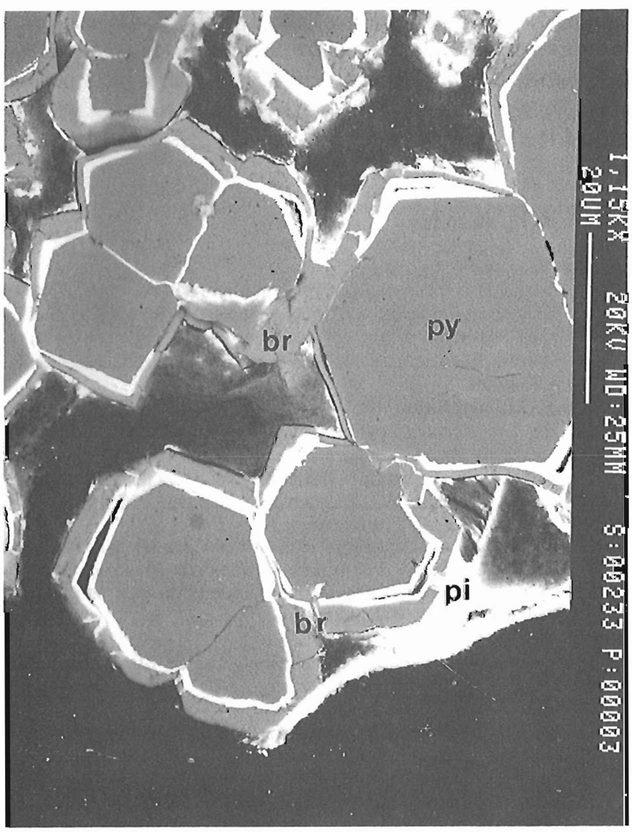

Figure 9. BSE (Back Scattered Electron) image of pyrite (py) surrounded by pitchblende (pi) and bravoite (br). Eagle Point deposit, Saskatchewan, drillhole ES 294/117.8 m. Photomicrograph, reflected light.

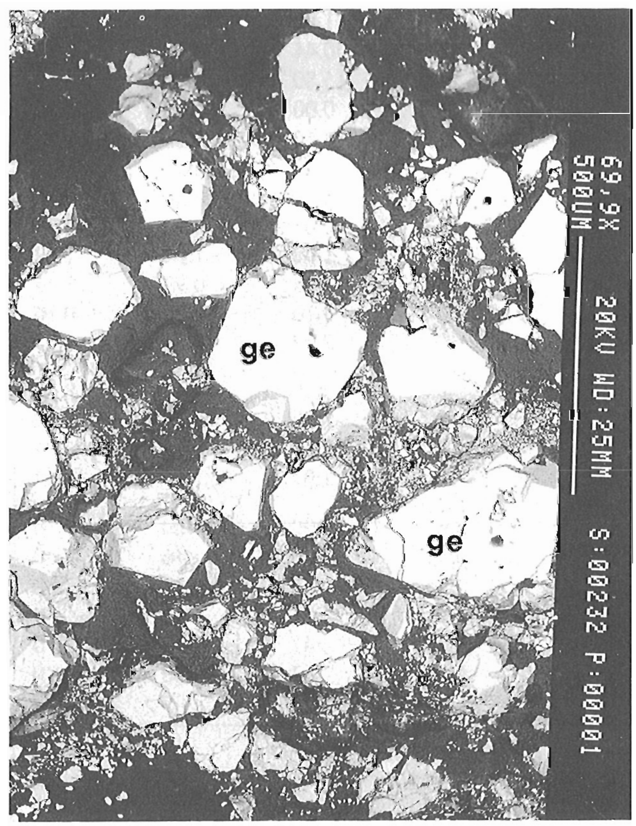

Figure 10. Granular, subhedral, gersdorffite $P2_13$ (ge) - a low-tem-perature polymorph of gersdorffite - in a granular quartz-rich matrix. Eagle Point deposit, Saskatchewan, drillhole ES 293/450.8 m. Photomicrograph, reflected light. XRF analysis of this specimen is in column 2 of Table 1.

generations (Ruzicka and Littlejohn, 1982; Ruzicka, 1986; Andrade, 1989). The mineralization is commonly associated with zones of alteration, which follow faults, bedding-planes, and foliation and cause corrosion or complete replacement of the rock-forming minerals. The pitchblende, intimately associated with micaceous matrix (Fig. 4), locally surrounds quartz grains and invades fractures in massive quartz in pegmatoids without any accompanying gangue minerals (Fig. 5, 6). The pitchblende saturation of hairline fractures in the quartz attests to very fine crystalline structure of the mineral and is considered by Frondel (1958) as "a consequence of rapid crystallization at relatively low temperatures or a result from the crystallization of gel masses". Pitchblende of the same crystal structure also forms a patch at the fractured quartz (Fig. 7). Locally, the pitchblende, associated with illite even replaces relics of quartz grains in the quartzite host (Fig. 8). The illite matrix contains scattered grains of pyrite, pyrrhotite, chalcopyrite, and bravoite. According to Ramdohr (1969) "bravoite always seems to be formed at low temperatures and almost always from descendent solutions, from percolating ground waters, or in euxenic sedimentary environments". The presence of bravoite, which locally surrounds pyrite and pitchblende (Fig. 9) thus attests to a low-temperature crystallization process (its stability is below 135°C, Kullerud, 1961). Furthermore, the low-temperature conditions are confirmed by an XRD (X-ray dffiation) identification of a granular, subhedral gersdorffite $P2_13$, which represents a low-temperature polymorph of this mineral (Fig. 10). (This specimen has provided a new standard pattern for the mineral XRD data base and has been deposited in the National Mineral Collection for reference.) The mineralization is associated with strong desilicification (see XRD analysis of sample 3 (mineralized) in comparison with samples 1 and 2 (nonmineralized) in Table 1) of the host rock and with anomalous concentration of rubidium (2868 ppm, Table 1).

These features of the Eagle Point uranium deposit thus indicate its formation under low-temperature conditions and support a conceptual model postulating consanguinity of the basement-hosted monometallic mineralization with the polymetallic mineralization directly associated with the unconformity (Ruzicka, 1984).

Table 1. (XRF) X-ray fluorescence analyses of selected samples from the Eagle Point deposit, Saskatchewan

Constituent	Unit	Sample 1	Sample 2	Sample 3
SiO_2	%	61.80	70.80	35.00
TiO_2	%	0.44	0.37	0.34
Al_2O_3	%	15.80	12.70	12.30
Cr_2O_3	%	0.00	0.00	0.00
Fe_2O_{3T}	%	7.50	2.50	2.00
Fe_2O_3	%	n.d.	n.d.	n.d.
FeO	%	n.d.	n.d.	n.d.
MnO	%	0.05	0.00	0.07
MgO	%	1.96	1.24	1.68
CaO	%	2.38	0.92	1.46
Na_2O	%	3.70	0.10	0.10
K_2O	%	2.02	3.86	2.17
H_2O_T	%	n.d.	n.d.	8.70
CO_{2T}	%	2.70	7.10	0.20
P_2O_5	%	0.21	0.10	0.21
S	%	1.97	1.69	0.07
Ba	ppm	309	420	0
Nb	ppm	30	46	0
Rb	ppm	93	159	2868
Sr	ppm	248	43	0
Y	ppm	47	39	0
Zr	ppm	152	178	0
U	ppm	8.5	4.9	224000

Sample 1: DH ES 293/271.3 m
Sample 2: DH ES 293/450.8 m
Sample 3: DH ES 294/117.8 m
n.d. no data

N.B. H_2O_T, CO_2 and S analyzed by rapid chemical methods; U by neutron activation.
Analyses in GSC by C. Veys; U by Saskatchewan Research Council.

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Recherches sur le pergélisol dans la région de Blanc-Sablon, Québec

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Résumé

Les palses de Blanc-Sablon, localité située au sud de l'isotherme annuelle de l'air de 0°C, font l'objet de recherches particulières depuis 1989. Le pergélisol, qui y atteint une épaisseur maximale de 10 m, est en dégradation et semble relictuel. Il s'agit probablement d'un pergélisol hérité d'une période froide antérieure, vraisemblablement le Petit Âge glaciaire. Il se serait conservé jusqu'à aujourd'hui grâce au climat particulier de cette région du golfe du Saint-Laurent; ce climat est caractérisé par des hivers peu rigoureux et des étés frais, peu ensoleillés et brumeux. Cette contribution rend compte des mesures en cours et offre quelques données préliminaires.

Abstract

Investigations of palsa in the Blanc-Sablon area, which is located to the south of the 0°C annual air temperature isotherm, have been underway since 1989. Permafrost, which can measure up to 10 m in thickness, is currently degrading. It is most likely a relict feature inherited from a colder period, most likely the Little Ice Age. Permafrost has been preserved in a few peat bogs because of the particular climatic conditions occurring in this area of the gulf of St. Lawrence. The climate is characterized by relatively mild winters and cool summers with little sunshine and frequent fog. This paper reports on field work in progress and presents preliminary data.

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INTRODUCTION

Dans la vallée de la rivière Blanc-Sablon, sur la basse Côte-Nord du golfe du Saint-Laurent (51°20' N, 57°10' W), il existe quelques îlots de pergélisol. À première vue, on ne s'étonne guère d'en trouver à cet endroit dépourvu d'arbres et d'aspect plutôt sévère que Hare (1959), à la suite de Hustich (1949), a cartographié comme une zone de toundra côtière, expression reprise dans une étude écologique récente (Ducruc et coll., 1985).

Mais en réalité, la région de Blanc-Sablon, de par sa proximité à la mer, connaît un climat moins sévère qu'on peut l'imaginer à distance. En effet, les hivers, bien que longs ne sont pas plus rigoureux qu'à Québec (fig.1). La température moyenne en janvier (le mois le plus froid aux deux endroits) est comparable: -11,8°contre -10,9°C. Par contre, les étés sont courts et frais, peu ensoleillés et brumeux. En comparaison avec Québec, la température moyenne du mois le plus chaud (août) n'est que de 12,1° contre 19,2° à Québec, en juillet. L'amplitude moyenne se révèle beaucoup plus faible que celle d'autres stations plus continentales, sises à la même latitude. D'après les données d'Environnement Canada pour la période 1968-1990. la température moyenne annuelle à Blanc-Sablon (station de l'aéroport) est de 0,6°C; à Québec, elle est de 4,5°, alors qu'à Mistassini et à Fort-Rupert (Jamésie), à peu près à la même latitude que Blanc-Sablon, la moyenne annuelle de température de l'air est inférieure à -1°C.

La présence de pergélisol à Blanc-Sablon, soit au sud de la limite connue du pergélisol discontinu ou sporadique (Brown, 1978) est loin d'être banale. Au contraire, elle mérite qu'on s'y intéresse, en particulier dans le contexte des changements à l'échelle planétaire et du réchauffement du climat suite à l'augmentation du CO_2 dans l'atmosphère.

PROBLÉMATIQUE

D'après la plupart des auteurs, la limite méridionale du pergélisol actif dans l'hémisphère Nord est sise au nord de l'isotherme annuelle de l'air de -1°C (Dionne, 1984). Au Canada, il existe très peu de sites de pergélisol au sud de cette isotherme. Les rares sites connus se trouvent au Manitoba dans des milieux humides et tourbeux, dans lesquels on observe des buttes de tourbe gelée plutôt que de véritables palses, c'est-à-dire des palses avec un noyau minéral recouvert d'une enveloppe de tourbe. En réalité, les palses typiques apparaissent surtout au nord de l'isotherme annuelle de -1°C. Au Québec, le type de palse trouvé à Blanc-Sablon (palses non boisées) ne se retrouve que beaucoup plus au nord, soit la région immédiate de Kuujjuarapik où la température moyenne annuelle de l'air est inférieure à -4°C (Dionne, 1984).

Dans l'est du Canada, Brown (1976) a signalé l'existence de palses sur la côte du Labrador, notamment à Carthwright (à 47 m d'altitude), où la température moyenne annuelle de l'air est de 0,2°C. L'hypothèse d'un pergélisol relictuel a été évoquée pour les sites sis au sud de l'isotherme annuelle de l'air de 0°C (Brown, 1979, p. 285).

La découverte de palses à Blanc-Sablon remonte à 1978. D'abord observées sur photographies aériennes, elles ont été visitées au cours de l'été de 1979 (Dionne, 1980). Par la suite, elles ont fait l'objet d'un essai sur la limite méridionale des palses dans l'hémisphère Nord (Dionne, 1984).

La présence de palses au sud de l'isotherme annuelle de l'air de 0°C étant inusitée voire même exceptionnelle, on a cherché à comprendre leur raison d'être à Blanc-Sablon. La conclusion préliminaire a été à l'effet qu'il s'agissait d'un pergélisol relictuel hérité d'une période antérieure plus

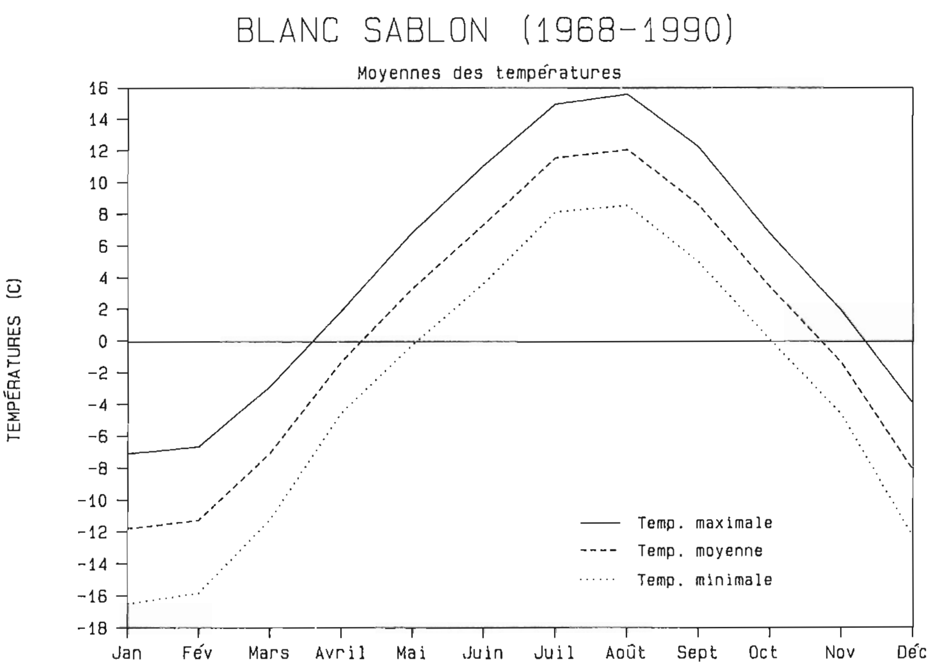

Figure 1

Moyennes annuelles des températures pour la période 1968-1990, station de Blanc-Sablon.

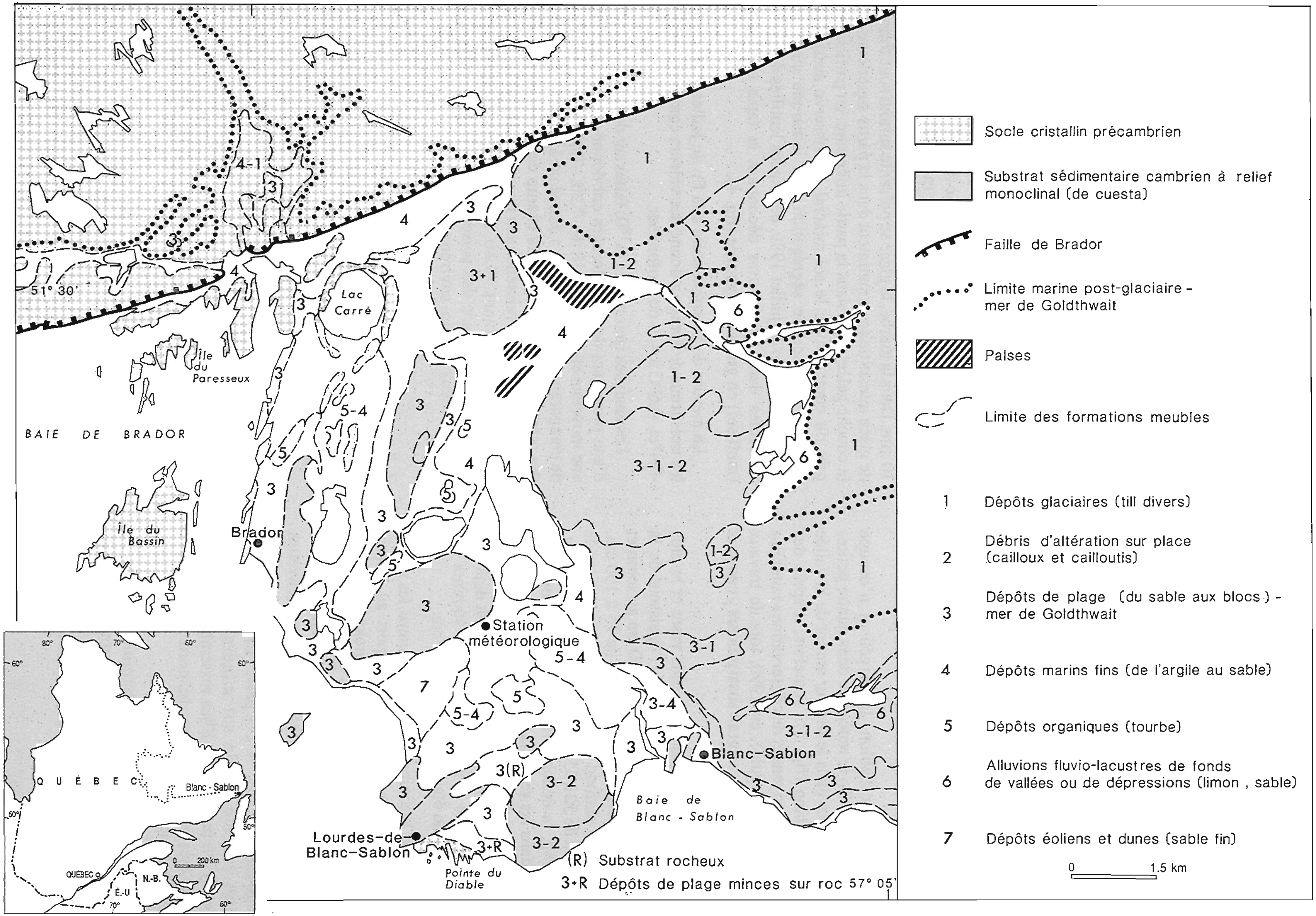

Figure 2. Croquis géomorphologique et carte de localisation des palses, région de Baie-Sablon, Basse Côte Nord du Saint-Laurent, Québec.

froide, soit le Petit Âge glaciaire. Malheureusement, cette conclusion ne reposait pas sur des mesures précises recueillies sur le terrain et portait flanc à la critique.

RECHERCHES EN COURS

Les recherches en cours ont précisément pour but de recueillir des données de terrain. Dans un premier temps, il était essentiel de connaître l'étendue et l'épaisseur du pergélisol et de préciser la nature du matériel ou du substrat touché. Ce travail a été effectué à l'été 1989. Dans un deuxième temps, il fallait obtenir des données sur la température au sol et dans le sol et suivre son évolution au cours d'une année au moins. À cet effet, un cable à thermistances a été installé dans une grosse palse en août 1990.

DONNÉES PRÉLIMINAIRES

1. Répartition, étendue et épaisseur du pergélisol

Dans la région de Blanc-Sablon, le pergélisol n'occupe qu'une infime superficie. On le rencontre dans quatre petites cuvettes tourbeuses, sises dans le secteur amont de la vallée de la rivière Blanc-Sablon (fig. 2). Toutefois, d'après des témoignages oraux des citoyens et des observations faites par l'un des auteurs (M. K. Seguin), en 1990, lors de travaux routiers et d'excavations, il y en aurait aussi ailleurs y compris dans le substrat rocheux.

Dans le champ de palses étudié, le plus vaste et le plus septentrional, sis à une altitude de 60 m, l'épaisseur de la tourbe varie d'une palse à l'autre allant de 50 cm au sommet à plus de 3 m d'épaisseur, la normale étant de 75 à 100 cm. À la fin août 1979, le plafond du pergélisol, au sommet de plusieurs palses, était de 50 à 55 cm de profondeur seulement. À la fin juillet 1989, la couche dégelée était de l'ordre de 35 cm seulement dans les secteurs couverts de lichens, alors que dans les aires de déflation (tourbe érodée), secteurs légèrement en creux et plus humides, elle atteignait déjà 50 à 55 cm. Le 28 septembre 1991, le mollisol au sommet des palses à tapis lichénique était aussi de 50 à 55 cm.

Dans le site étudié, le moins dégradé des quatre, les palses ont entre 1 et 1,5 m de hauteur par rapport au terrain avoisinant; mais au droit des cuvettes thermokarstiques, la hauteur relative des palses peut atteindre 2,5 m. Ce site est caractérisé par un complexe de buttes relativement étroites et allongées ainsi que par de nombreuses dépressions (cuvettes thermokarstiques) de forme circulaire, de dimensions très variées et d'âge différent (fig. 3). Les palses sont constituées de tourbe, en surface, et de sol minéral (sable fin limono-argileux), en profondeur. Elles ressemblent donc au type classique très répandu dans le nord du Québec, notamment en Hudsonie (Dionne, 1978).

L'épaisseur du pergélisol a été déterminée à l'aide de méthodes géophysiques (Seguin, 1992) comprenant 1) des sondages de résistivité électrique (SR); 2) des sondages-profilages de polarisation provoquée (PP); et 3) des profils géothermiques.

En général, le pergélisol est peu épais: 4 à 5 m en moyenne. Par endroits, cependant, dans les plus grosses palses, il atteint 7 à 8 m, au maximum 10 m. D'après le forage effectué pour installer le cable à thermistances, il dépasserait même 10 m par endroits. L'épaisseur a été vérifiée à l'aide de forages, dont trois de 4 m, ce qui a permis de recueillir des échantillons aux fins de diverses analyses, (granulométrie, teneur en eau et en glace de ségrégation, composition isotopique 0^{18}). L'épaisseur du pergélisol le long des trois transects principaux varie beaucoup. Le plancher du pergélisol est curieusement très ondulé. Le pergélisol se rencontre à la fois dans la tourbe et le sol minéral, mais la majeure partie est dans le minéral.

La teneur en eau et en glace a été mesurée au moyen d'un échantillonnage systématique des carottes provenant des trois forages. Elle est naturellement plus élevée dans la tourbe. Dans le minéral, la teneur en eau varie de 30 à 88 %. La glace de ségrégation est abondante dans le minéral; elle forme des lentilles de 1 à 3 mm, en général, mais qui atteignent parfois 10 mm. Dans la tourbe, la glace se présente surtout sous forme de glace interstitielle, notamment dans la partie supérieure de la couche gelée; mais dans la partie inférieure, en particulier près du contact avec le sol minéral, on observe aussi des lentilles de glace de ségrégation. On remarque aussi la présence, à la fin de juillet 1989, de minces lentilles de glace verticales dans les fissures découpant le tapis tourbeux (mollisol) en grands polygones (fig. 4).

La mesure de la température à la fin de juillet 1989 dans les trois trous de forage a donné des valeurs comprises entre -0,4°et -0,6°C à 4 m de profondeur.

Les analyses isotopiques ($\Delta^{18}0$) faites sur des échantillons d'eau provenant des trois forages ont donné respectivement une moyenne en oxygène -18 de -7,30, -7,83 et -7,33. À titre de comparaison, l'eau de mer a donné un $\Delta^{18}0$ de -2,86, celle des mares thermokarstiques du champ de palses entre -2,06 et -4,45; les eaux de ruissellement dans la vallée de la rivière Blanc-Sablon donnent -8,80 à -11,03. La composition isotopique de la glace des palses diffère donc de celle des divers milieux aqueux environnants actuels.

2. Le couvert neigeux

Les données météorologiques indiquent que la région de Blanc-Sablon est bien arrosée. En effet, il y tombe, en moyenne, 1140 mm de précipitations totales par année. La neige totalise 474 cm, dont 75 % tombe entre décembre et mars, principalement en janvier et décembre, les deux mois de l'année où les précipitations sont les plus fortes. Le champ de palses étudié étant localisé au fond et à la tête de la vallée de la rivière Blanc-Sablon, qui forme un large amphithéâtre avec des versants relativement raides, la neige poussée par les vents d'ouest y est piégée.

Les observations faites à la fin mars 1991 lorsque les températures journalières atteignaient entre -14°et -16°C, indiquent une accumulation préférentielle de la neige sur la moitié inférieure des versants (zones couvertes de krummolz), dans les ravins, dans la rainure de la rivière et dans les petites dépressions (les mares thermokarstiques). À

Figure 3. Vue aérienne oblique à basse altitude d'une partie du champ de palses étudié. À remarquer l'importance des formes thermokarstiques soulignant l'état avancé de la dégradation du pergélisol. Photo prise le 20 août, 1979. La situation a peu changée depuis.

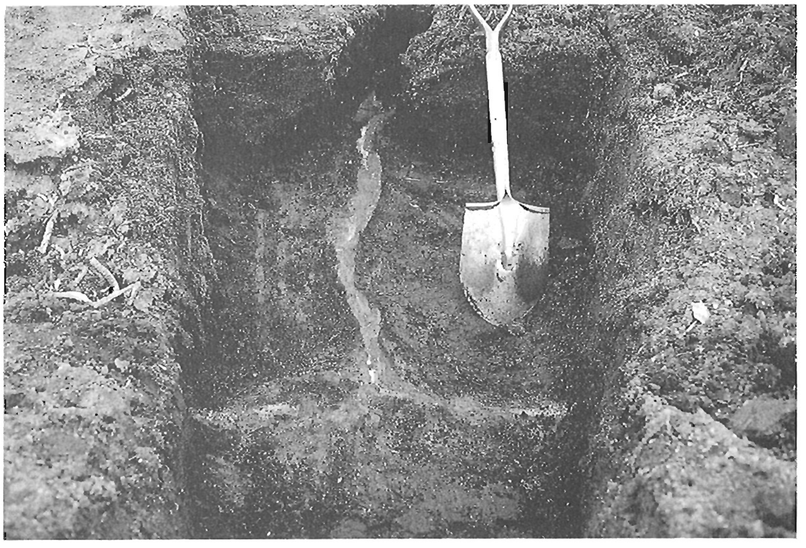

Figure 4. Fissure de gel annuel remplie de glace. Ces fissures forment des polygones à la surface des palses aux endroits dénudés (érodés). Les coins de glace ne pénètrent pas dans le pergélisol. Photo prise à la fin juillet, 1989.

Figure 5. Vue aérienne oblique d'une partie du champ de palses étudié. Les dépressions sont comblées de neige alors que le sommet des palses est dénudé. Photo prise le 22 mars, 1991, après une période de temps doux et de pluie.

ces endroits, l'épaisseur de la neige atteignait ou dépassait 300 cm le 21 mars 1991. Dans le champ de palses proprement dit, les buttes étaient presque entièrement (90 %) enterrées; seule l'extrémité sommitale émergeait de quelques centimètres au-dessus de la couverture neigeuse. Ce jour-là le sommet était d'ailleurs couvert d'une mince pellicule de glace et de neige durcie. Deux jours plus tard, le sommet des principales crêtes était nettement visible (fig. 5).

Il convient, toutefois, de mentionner que quelques jours de pluie et des températures journalières au-dessus de la moyenne avaient prévalu la semaine précédant les observations. L'épaisseur du couvert neigeux au sommet des palses en hiver (janvier-février), soit durant la période la plus froide, n'a pas été mesurée. De même, la densité de la neige (paramètre important pour l'isolation) dans les mares thermokarstiques n'a pas été mesurée parce que la neige, sous une croûte durcie de 3 à 4 cm d'épaisseur, était gorgée d'eau, donc impropre à des données valables.

Il faut souligner en passant que l'enneigement à Blanc-Sablon est de loin supérieur à celui prévalant dans les régions où on trouve des palses classiques (soit des palses du type de Blanc-Sablon) puisqu'il est généralement inférieur à 200 cm; ceci fait de Blanc-Sablon un site assez singulier.En conséquence, de plus amples observations sur les caractéristiques du couvert neigeux paraissent nécessaires.

3. *Cable à thermistances*

Un cable à douze thermistances a été installé à la mi-août 1990, au sommet d'une des plus grosses palses afin 1) de mesurer en continu la température de l'air, celle au sol et celle du pergélisol à différentes profondeurs; 2) de connaître le bilan thermique local et 3) de pouvoir, par la suite, déterminer dans quel sens évolue le pergélisol à Blanc-Sablon, projet qui requerra quelques années d'observations.

Des douze thermistances, une enregistre la température de l'air à 1,5 m au-dessus du sol, une autre la température à la surface du sol mais à l'abri, une autre la température du mollisol (50 cm) et les neuf autres la température du pergélisol à tous les mètres de profondeur. Les thermistances (de type Yellow Spring - modèle 44033 pour le sol et Fen-Wals Electronics - UVB-31J-1 pour l'air) sont couplées à une mémoire solide (logger CR-10 de la Campbell Scientific avec mémoire SM-192K à 12 canaux), qui enregistre la température aux quatre heures, sauf la température de l'air qu'elle enregistre à toutes les heures.

Les données préliminaires obtenues en mars 1991 indiquaient une faible différence avec la température de l'air à la station de l'aéroport pour la période allant de septembre 1990 à mars 1991. Le 23 mars, la température du sol à 50 cm de profondeur était de -4,2°C; elle était de -2,7° à 1 m et de -0,25° à 10 m de profondeur (fig. 6).

4. *Nature et âge du tapis tourbeux*

Une coupe d'environ 3 m de hauteur a été pratiquée sur le versant d'une mare de thermokarst qui s'est formée au centre d'une grosse palse. La tourbe atteignait 2,5 m d'épaisseur. Un échantillon de débris organiques recueilli à 30 cm sous le dépôt de tourbe proprement dit, soit dans les sédiments lacustres, a donné un âge radiométrique de 8830 ± 100 BP (GSC-4946)[1]. Cette date confirme l'émersion rapide des terres après la submersion goldthwaitienne survenue vers 12 ka (De Boutray et Hillaire-Marcel, 1977; Grant, 1972, 1986; Bigras et Dubois, 1987). Il y a environ 9000 ans, le niveau de 60 m était déjà émergé et un lac résiduel peu profond occupait une dépression sise derrière une moraine frontale dans le secteur amont de la vallée de la rivière Blanc-Sablon.

Le dépôt de tourbe est composé de trois unités principales dont deux de sphaignes relativement peu décomposées et séparées par une couche plus mince de tourbe brun foncé à noirâtre, bien décomposée et riche en macro-restes (tiges et racines d'arbustes). Une série de datations ^{14}C faites sur des

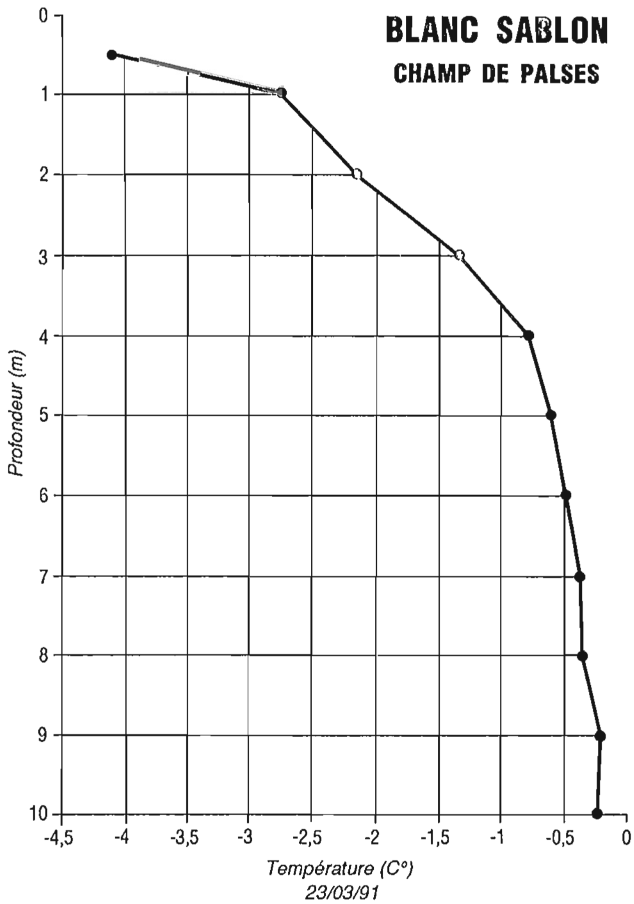

Figure 6. Température du sol à différentes profondeurs le 23 mars, 1991.

[1]Âge non corrigé

échantillons pris à la base, au milieu et au sommet de chaque unité montre une accumulation progressive et continue de la matière organique dans le bassin entre il y a 4000 et 8400 ans. La base de la tourbe est, en effet, datée à 8370 ± 100 BP (GSC-4952), alors que le sommet, caractérisé par la présence d'une couche noire (horizon de feu) ainsi que de subfossiles de krummolz, est daté à 3950 ± 60 BP (GSC-5014) et 4010 ± 60 BP (GSC-5010).

La surface de la tourbière est présentement recouverte d'un tapis de lichen dominé par les espèces Vaccinium sp., Empetrum sp. et Ledum sp., qui a l'aspect d'un humus ligneux et sec d'une dizaine de centimètres d'épaisseur. Cette couche indique un changement marqué des conditions hydriques locales correspondant probablement à un soulèvement de la surface suite à la formation du pergélisol. La base de cette couche date d'environ 500 ans (480 ± 50 BP; GSC-4991), soit à peu près le début du Petit Âge glaciaire (Grove, 1988). Il existe donc une lacune stratigraphique importante entre le tapis de lichen superficiel et le dépôt tourbeux sous-jacent. Pour l'instant, cette discontinuité n'est pas expliquée. Un nouvel échantillonnage des premiers 60 cm du tapis tourbeux a été fait en septembre 1991 afin de vérifier si la lacune stratigraphique constatée dans la coupe de référence existait aussi ailleurs.

CONCLUSION

Les recherches en cours, entreprises à l'été de 1989, ont déjà fourni des données appuyant l'hypothèse d'un pergélisol relique hérité du Petit Âge glaciaire (Dionne, 1983, 1984). Le pergélisol aurait été maintenu grâce aux conditions climatiques particulières de la région de Blanc-Sablon, climat caractérisé par des hivers peu rigoureux et des étés frais, peu ensoleillés et brumeux. L'analyse complète des données de la température de l'air et du sol mesurées au cours de la première année devrait confirmer si le pergélisol est en équilibre avec les conditions climatiques modernes.

Dans la perspective des changements à l'échelle planétaire, le cas de Blanc-Sablon présente un intérêt indéniable et pourrait servir aux spécialistes dans l'élaboration de modèles. Si le pergélisol est effectivement relictuel et hérité du Petit Âge glaciaire, on aurait là un bon exemple d'une évolution très lente à la suite d'une amélioration climatique relativement importante. Ainsi, les effets catastrophiques dus au réchauffement causé par l'effet de serre, tels que prédits par certains modèles, paraissent plutôt alarmistes. Si, au contraire, le pergélisol est moderne ou actuel, on a là une excellente démonstration du fait que la température de l'air ne constitue pas le seul facteur. D'autres paramètres, comme l'ensoleillement, l'évaporation, l'albédo, le drainage, et autres, doivent aussi être considérés.

Quoiqu'il en soit, étant donné l'état actuel des connaissances, il arrive rarement que, dans l'hémisphère Nord, le pergélisol se développe au sud de l'isotherme annuel de l'air de -1°C.

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The use of the characteristics of native gold as an exploration tool: an overview with emphasis on the Soviet contribution

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Abstract

The author visited gold fields of the eastern USSR in the summer of 1991 and had discussions with the local gold specialists. The Soviet workers, who for the past 40 years have studied the fineness, trace element composition, and morphological character of placer gold have concluded that these properties are generally inherited directly from the bedrock source. "New" gold is not important in the formation of placers. They have found that these features are a useful tool both for the interpretation of the history of gold in the surficial environment (e.g., the recognition of gold deformed under eolian conditions) and for the location and classification of the bedrock source of the gold (e.g., mesothermal gold quartz veins of different mineralogies). Their work is in general agreement with the present author's published findings.

Résumé

L'auteur a visité pendant l'été 1991 les districts aurifères de l'est de l'URSS et a discuté avec les spécialistes des gîtes aurifères de la région. Les chercheurs soviétiques, qui depuis 40 ans étudient la finesse, la composition en éléments à l'état de traces et le caractère morphologique de l'or alluvionnaire, concluent que ces propriétés sont en général directement héritées de la roche en place d'origine. L'or «nouveau» ne joue pas un rôle important dans la formation des placers. Les chercheurs ont noté que ces détails constituent un outil de valeur à la fois pour interpréter l'évolution de l'or dans un milieu de surface (par ex. pour reconnaître l'or déformé par l'action du vent) et pour situer et classer la roche en place qui est à l'origine de la minéralisation aurifère (par ex. les filons mésothermaux de quartz aurifère de minéralogie diverse). Les résultats de leurs travaux concordent généralement avec les résultats publiés par l'auteur du présent article.

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INTRODUCTION

In the summer of 1991 the author visited some of the gold producing areas of the eastern USSR and discussed Soviet work on the characteristics of gold with the Soviet researchers. This brief report describes some of the Soviet work. A summary of the author's work is provided to enable the Soviet work to be placed into a Canadian context.

The author and Dr. K.C. McTaggart have been studying the characteristics of lode and placer gold in British Columbia and its use in exploration for the past 10 years (Knight and McTaggart, 1986, 1989, 1990; Nelson et al., 1990). The author, S. Morison, and Dr. J. Mortensen have been studying gold in the Yukon for the last 5 years. In order to study the relationship between the characteristics of gold, the author and his associates compiled a database containing the characteristics of gold from British Columbia and the Yukon. It consists of some 6000 microprobe analyses of gold particles (Au, Ag, Cu, Hg; with an average total of about 99.8 wt. %; standard deviation of about 0.3; detection limit for Hg of 0.065 wt. % and Cu 0.025 wt. %) and observations of the morphology and internal structure of these particles (Knight and McTaggart, 1986, 1989, 1990; Nelson et al., 1990, and unpublished data).

The author and his associates concluded that the gold in lode and placer gold deposits vary widely in fineness (i.e. gold content expressed as ppt.), Hg and Cu content, and in the morphology of gold particles. These variations were used to relate the placer gold to its bedrock source and to classify the lode source type. For example, in the Coquihalla area Knight and McTaggart (1990) relate Cu-rich placer gold to a lode containing Cu-rich gold. Such gold occurs typically in altered ultramafic rocks.

Because of the dependence of the USSR on gold for foreign exchange, the Soviets have, for many years, conducted intensive studies of the characteristics of placer gold and their usefulness in finding the bedrock sources of the gold. Consequently in the USSR there is a large body of literature and many scientists with expertise on this subject (Petrovskaya, 1973, 1974; Sergeenko and Samusikov, 1975; Nesterenko, 1977; Skryabin, 1978; Nikolaeva, 1979, 1983, 1990; Yablokova, 1980; Samusikov, 1981; Samusikov and Petrova, 1983; Murzin and Malygin, 1987; Filippov and Nikiforova, 1988, 1990; Nikiforova and Filippov, 1990; Savva and Preis, 1990; see also the bibliography in Boyle, 1979). Most authors writing in English do not realize the extent of the Soviet investigations. The language barrier has certainly been a factor in this lack of familiarity. But in addition the Soviets have traditionally placed strict limits on the release of 'strategic' information such as location maps, map scales, and almost all information dealing with gold deposits and production. Many of the Soviet papers are therefore incomplete by western standards, (e.g., Fedchuck et al., 1983) or simply unavailable. However it is clear from what has been published that they have considerable research and exploration expertise and experience in this field.

With the ongoing restructuring of the Soviet Union the restrictions on certain types of information are being lifted. An extended visit in the summer of 1991 to the Yakutsk Institute of Geology in Yakutsk (Fig. 1) and a brief visit to the North Eastern Integrated Scientific Institute in Magadan (both institutes of the Academy of Sciences of the USSR) provided an opportunity to fill in some of the gaps in the Soviet literature and to learn first hand about the Soviet work. Nevertheless, because much of the work is done for expeditions (regional mineral exploration groups) and because the restructuring is not complete, some information is still not available.

ALDAN DISTRICT AND STANOVOY RIDGE

In the unglaciated Aldan shield region (Southern Yakutia-Sakha ASSR) the gold district is centred around the town of Aldan. The Soviets have identified at least 4 types of lode gold mineralization there. All are related to syenitic Mesozoic plutons, dykes, and plugs emplaced to a moderate to shallow depth. The intrusions penetrate Archean rocks which are unconformably overlain by Cambrian and Jurassic carbonate rocks. The gold deposits are generally related to the contact between the Archean shield rocks and the overlying carbonates. As expected, skarn-related deposits dominate although quartz veins are important. The most famous of the skarn deposits is the Kuranach deposit some 30 km north of Aldan (e.g., Boyle, 1979; Kochetkov et al., 1979)

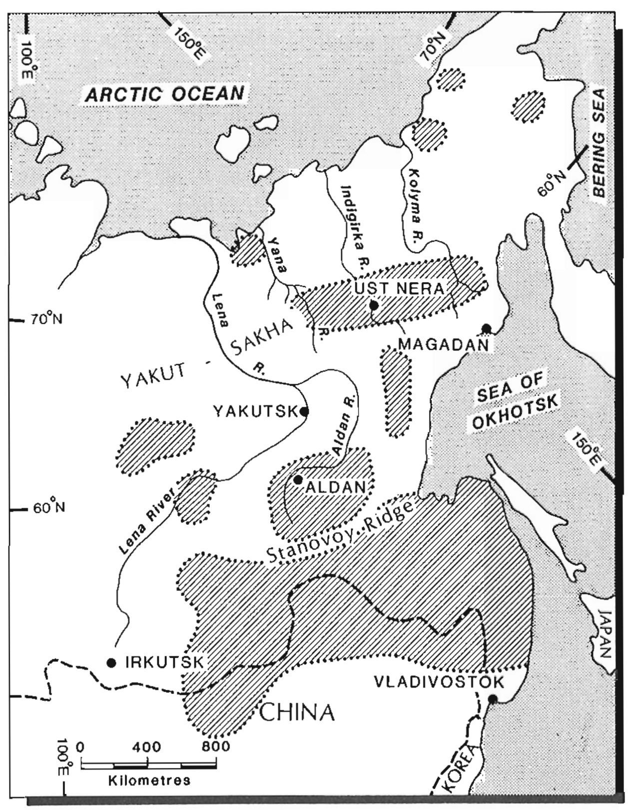

Figure 1. Location map of places visited in the eastern Soviet Union. Shaded areas are the principal placer gold producing areas of the northeastern USSR (after Boyle, 1979 with revisions by V. Samusikov, pers. comm., 1991).

where the mineralization is concentrated in paleoregoliths and karsts (perhaps of Jurassic age) in the carbonate sequence. In the unoxidized ore, free gold and gold telluride minerals occur, while in the oxidized ore, which makes up the majority of the ore, only native gold is commonly found (A. Kim, pers. comm., 1991). More than half the gold is in the <100μm size fraction. Within the district in general, the gold particle sizes in the lode deposits vary from small nuggets to "invisible", with a significant proportion of the gold falling in the <100μm range.

Despite a long history of placer mining, placer gold is still the most important gold source in the district. Today about 10 dredges and numerous small operations are active. There is general acceptance in the Soviet Union that there is a direct connection between bedrock deposits and placers and that the characteristics of gold in the placer deposits have been inherited from the lode (e.g., Petrovskaya, 1973). From a discussion with one of the exploration (expedition) geologists working in the region it appears that the characteristics of gold are routinely studied in their exploration program (see also Kochetkov et al., 1979). The most common source of data on the characteristics of gold in the surficial environment comes from the placer gold recovered during mining. Although most of the gold particles in a particular lode may be small (<100μm) most of the creeks draining that lode have developed placers. The local geologists feel that even if the placers are economically insignificant they are still useful in locating economically significant bedrock sources. The characteristics of gold panned from the soils nearer to the sources and away from a creek are used to try to identify possible sources.

Academic studies in the area have broadened the understanding of the systematic relationships between the characteristics of gold in the surficial environment and its source. A discussion with A. Kim (also Kim, 1975; Kim and Lantsev, 1980) indicated that the morphology, size, trace element content, and inclusions are the most useful of the characteristics of gold in the Aldan region. For example, the oxidation of Au tellurides and maldonite (AuBi) present in the Kuranach deposit has resulted in the formation of 'gold sponge' i.e. porus gold masses, in the oxidation zone. The presence of this type of gold in the placers and soils provides a strong indication of the nearness of to this type of source. Care should be taken not to confuse this natural sponge with gold sponge formed by amalgamation of gold with mercury lost during previous, old placer mining operations.

To the immediate south of the Aldan area, in the Stanavoy ridge region, where Archean greenstones dominate, the fineness, trace element content, inclusions in, and attachments to the gold are the most significant characteristics (A. Kim, pers. comm., 1991). These characteristics have led to the recognition of 4 possible lode source types (Table 1) although it appears that no major deposits have yet been found.

UST NERA DISTRICT

The Ust Nera (Upper Indigirka River) gold placer district of Northeast Yakutia-Sakha ASSR, is underlain by siltstones, sandstones, and argillites of Triassic-Jurassic age. Many of the argillites are graphite- and pyrite-rich. The regional grade of metamorphism is low. The area is intruded by upper Jurassic-lower Cretaceous granite plutons which have contact aureoles about 300 m wide. Most of the mineralization is concentrated in quartz veins which are localized in the siltstone members. The distribution of mineralization on the geological maps appears to be, at least in part, peripheral to the granite intrusions. Despite the fact that some of the veins attain grades of 32 g/t in gold, only a few of the veins are being mined and most of the gold production is from the placers. Although the veins are oxidized to 100 m, the local geologists think that the oxidation, with the possible exception of one or two locations, has not played a significant part in the mobilization and reprecipitation of the gold. The characteristics of the gold found in the placers are therefore inherited directly from the characteristics of the gold in the source lode. The shape of a gold particles changes while the composition remains constant with distance of transport from the lode source. The mining geologists believe that the direct relationship between the characteristics of placer gold and the gold in the bedrock sources will enable them to find lode sources in the future. Records and samples of gold are being kept to enable them to take advantage of this possibility. Because of the richness of the placers, most of the present work on the relationship between the placers and the lodes is of an academic nature.

One example of the type of study which has worked particularly well in the Ust Nera area is that of Skryabin (1978, 1989a,b). He postulates that there is a simple relationship, within the mineralization area, between the composition of gold, the location of the lode source (quartz vein), and the mineralogy of the gold-quartz vein. When a contour map of fineness is superimposed on a geological map it is seen that the isopleths define highs around the granitic intrusions (Skryabin, 1978). Furthermore there is a

Table 1. Type of lode gold occurrence expected in the Stanovoy area inferred from the characteristics of gold and gangue minerals attached to gold which was recovered from placers and soils (after A. Kim, pers. comm., 1991)

Inferred Au lode type	fineness	attached gangue minerals
Mo, W	780 - 900	quartz, muscovite
Ag	500 - 700	quartz, kaolinite
Polysulphide	800 - 900	quartz, carbonate
Metamorphic	>950	quartz, amphibole, sericite, albite, chlorite

systematic relationship between the mineralogy of the gold-quartz veins and the fineness of the gold contained in it, and consequently between the vein mineralogy and its distance from the granitic body (Table 2).

The classification of the lode types of Table 2 apparently has been used successfully by the exploration geologists to find antimony-rich gold quartz veins by using the composition of the placer gold.

V. Filippov (unpub. data), has indicated that in one area of the Stanovoy ridge district the contoured fineness values have helped in the recognition of the significance of granitic sills found in drill core and may also provide information about the direction of movement of the mineralizing fluids.

EOLIAN GOLD

Gold with an unusual morphology (toroids and hollow spheroids), have been reported from the Aldan shield and the Anabar shield areas. Early reports include those of Yablokova (1972) and Shpunt (1974). Recent work on these gold particles in the basin between these two shield areas by Filippov and Nikiforova (1988, 1990), has shown that the morphology of this gold is the result of eolian deformation. During the Quaternary, the Aldan and Anabar shields and the intervening basin were the site of large cold deserts. The 70 m or so of loess in the Yakutsk area, some 500 km to the east of these areas, represents material removed from the Aldan and Anabar shields by eolian erosion. During the process of erosion it is postulated that residual gold concentrations were formed. This gold was subject to severe eolian deformation which experiments have shown (Filippov and Nikiforova, 1988) was clearly capable of forming the observed toroidal and spheroidal shapes. These experiments also showed that alluvial processes could not have formed this type of gold. A brief field visit to an occurrence of this spheroidal gold about 40 km west of Yakutsk, revealed a most unusual setting. The plateau area has very low local relief (<100 m) and the rivers crossing it are sinuous in the extreme. The Precambrian to Cretaceous rocks which form the basement are unconformably overlain by up to 3 m of Quaternary sediments. A significant proportion of the sediments is eolian sands. The surface and river bars are almost devoid of particles larger than about sand size. In the one instance where there was an occurrence of alluvial pebbles on a river bar (mostly of well rounded quartz to 5 cm) it was possible to pan gold in the sub-millimetre size range. The majority of the gold is in the <.2 mm size range and spheroidal in shape (Fig. 2). Because of the poor exposure the relationship between the gold occurrence and the stratigraphy is unclear. The local geologists report that the gold is concentrated in the lower sections of the Quaternary sequence. After approximately 10 km of transport in the alluvial environment the particles lose much of their spheroidal shape. In the Tymanski district of the northern Urals, Soviet geologists have applied this information about the deformation of gold under eolian conditions to gold recovered from a Devonian deposit hosted by unconsolidated sediments. Based on the morphology of the gold and the sedimentology of the host, they concluded that the deposit is eolian in origin, i.e. a paleo eolian placer.

The recognition and understanding of this type of gold is significant because it is likely that similar conditions existed during the last ice age in North America, specifically in the Yukon and Western Alberta. In Canada the conditions of formation for this type of gold are only well preserved in the Yukon but the recognition of gold with this morphology in other parts of Canada could help in the interpretation of the conditions under which it formed and the origin of the gold anomalies associated with it. For example, some of the gold described from Western Alberta by Giusti and Smith (1984), may belong in this category.

Table 2. Relationship between the fineness of gold and the quartz vein mineralogy for the Indigirka River area (Skryabin (1978, 1989a,b) and V. Samusikov (pers. comm., 1991))

Fineness	Mineralogy of the quartz vein
650 - 750	gold, wolframite
750 - 875	gold, scheelite, arsenopyrite
875 - 925	gold, arsenopyrite, pyrite
925 - 950	gold, sulphide, antimonide
950 - 999	gold, antimonide

NUGGETS AND NEW GOLD

Soviet geologists have considered the standard arguments about the origin of nuggets. They do not find it necessary to advocate mechanisms of precipitation or agglomeration to

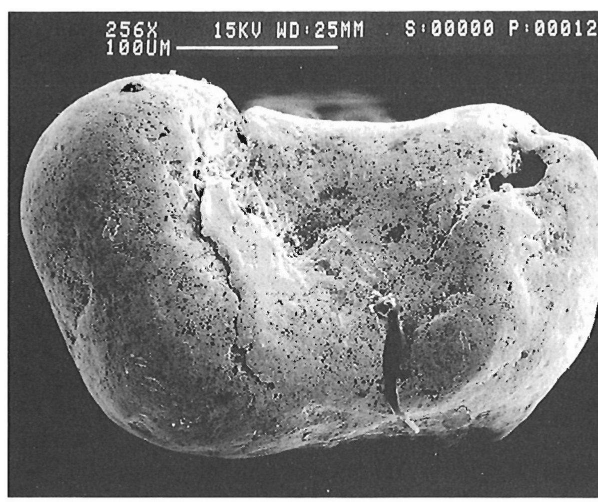

Figure 2. Secondary electron photograph of a gold particle deformed under eolian conditions. Some of the characteristics of eolian gold illustrated by this particle include: a very smooth surface, a suture and a hole in the upper righthand corner. The hole and suture suggest that this particle is hollow. Particle collected 40 km east of Yakutsk.

form nuggets. They believe that placer nuggets are large gold masses which have been eroded from the lode where they formed under hypogene conditions. The report by Samusikov (1990) which concluded that the composition of large placer nuggets represents the conditions of their growth in the lode and the samples of large gold masses from various lodes in the eastern USSR seen by the author (e.g., 5 masses >2 X 2 X 2 cm) are two pieces of evidence in support of their conclusion.

No discussion of the characteristics of gold would be complete without a mention of "new" gold, i.e. gold precipitated in the surficial environment rather than derived directly from the lode. There is general agreement among Soviet geologists that "new" gold is unimportant in the formation of placers (Yablokova, 1980). Knight and McTaggart (1990) reached a similar conclusion for the placer areas studied in British Columbia and the Yukon. Knight and McTaggart (1990) also concluded that "new" gold forms in restricted environments, is usually limited to a size fraction <100μm, commonly has a spongy or platy morphology, and has a high fineness. Soviet geologists (Nikolaeva, 1958; Pitul'ko, 1976; Yablokova, 1980; Nesterenko et al., 1985) generally agree with these conclusions. Spongy gold formed by the breakdown (oxidation) of gold minerals is one exception because the size of the spongy gold particle is determined by the original size of the gold mineral. (Spongy gold can be formed by the removal of mercury from a manmade gold-mercury amalgam and is not considered here).

Despite this consensus there is no general agreement amongst the Soviets (as in the West) on the exact significance and mode of formation of "new" gold in the surficial environment. From discussions with the Soviet geologists, it appears that in large part this disagreement is the result of uncertainty about the characteristics which define "new" gold, uncertainty about the characteristics indicating dissolution of gold, and uncertainty about the common distance of transport of gold in solution from its solution to precipitation point. V. Filippov (pers. comm., 1991) provided an example of the importance of resolving these uncertainties. He noted that in the Stanovoy ridge area the application and interpretation of soil geochemistry is complicated. The geologists in the area believe that these complications are the result of the dissolution in that area of most of the very small particles of gold. The case for dissolution is supported by the recognition of dissolution textures on the surface of the remaining larger gold particles. He notes, however, that no sites for the deposition of this gold have been found and postulates that the gold is either taken up in the vegetation or removed to the sea.

Nearly all the placer gold samples from British Columbia and the Yukon that the author studied included particles with a partial to complete rim of high fineness (>970) gold with Hg and Cu below the detection limit. The rim varies from submicron to 20 m in thickness. It was concluded that the rims were formed by the removal of Ag, Hg, and Cu and not by the precipitation of gold (Knight and McTaggart, 1990); i.e. rims do not represent "new" gold. There is general agreement amongst the Soviet geologists with this conclusion.

CONCLUSIONS

An extended visit with gold specialists in Yakutia-Sakha ASSR and a brief visit to the Northeast region of Russia SSR have shown that a considerable body of knowledge and experience about the characteristics of gold is being put to use in exploration.

Several of the conclusions about the origin of the characteristics of gold in Canada reached by the writer and his colleagues are in agreement with those reached by the Soviet geologists in the areas visited. Specifically, the chemical composition of placer gold is generally inherited from the lode source and that the composition of the gold reflects the lode type. In addition there seems to be general agreement on the low importance of "new" gold" in the formation of placers, the origin of nuggets in lode deposits and the formation of pure gold rims on alluvial gold particles by the removal of Ag, Cu, and Hg from the surface of the particle.

The conclusion of the writer and his colleagues that these characteristics can be useful in exploration is not new to the Soviets. They have been applied by the Soviet geologists for many years to locate bedrock sources, identify lode types and recognize gold deformed in the eolian environment.

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Teskey, D.J., Stone, P.E., Kiss, F., Dostaler, F., Anderson, K., Tod, J., Knappers, W., Jobin, D., and Gibb, R.A., 1992: Aeromagnetic survey program of the Geological Survey of Canada, 1991-92; in *Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 73-76.*

Abstract

During 1991-92, the GSC collected 183 022 line kilometres of aeromagnetic data. Of these, 73 202 line kilometres were flown in Alberta during the second phase of a three year cost-sharing government/industry survey, 51 250 line kilometres were flown in Saskatchewan and Manitoba in the first phase of a four year cost-sharing federal government/provincial government/industry survey, 4 000 line kilometres were flown over the Lincoln Sea in the final phase of a three year joint Institute of Aerospace Research (IAR)/Defence Research Establishment Pacific(DREP)/Geological Survey of Canada project, and 12 000 line kilometres were flown over the polar margin off Axel Heiberg and Ellesmere islands as the first phase of a second joint IAR/DREP/GSC project. The remaining data were collected under the federal-provincial Mineral Development Agreement program. Aeromagnetic gradiometer, total field and EM data were collected in four provinces - Nova Scotia (7 170 line kilometres), New Brunswick (10 600 line kilometres), Manitoba (13 800 line kilometres), and Saskatchewan (11 000 line kilometres). The GSC, under an agreement with the Canadian International Development Agency (CIDA), continued to monitor the scientific and technical aspects of a contract survey in Zimbabwe. The flying phase of this survey (144 824 line kilometres) was completed early in 1991.

Résumé

En 1991-1992, la CGC a recueilli des données sur 183 022 kilomètres de lignes de levés aéromagnétiques. De ce total, 73 202 kilomètres de levés aéroportés ont été réalisés en Alberta lors de la seconde phase d'un projet de levé de trois ans, dont le gouvernement fédéral et l'industrie se partagent le coût, 51 250 kilomètres de levés aéroportés ont été réalisés en Saskatchewan et au Manitoba lors de la première phase d'un projet de levé de quatre ans dont le gouvernement fédéral, le gouvernement provincial et l'industrie se partagent le coût, 4 000 kilomètres de levés aéroportés ont été réalisés au-dessus de la mer de Lincoln lors de la phase finale d'un projet conjointement entrepris pendant une période de trois ans par l'Institut de la recherche aérospatiale (IRA), le Centre de recherches pour la Défense-région pacifique (CRDP) et la Commission géologique du Canada (CGC), et 12 000 kilomètres de levés ont été réalisés au-dessus de la marge polaire au large de l'île Axel Heiberg et de l'île d'Ellesmere dans le cadre de la première phase d'un second projet conjointement entrepris par l'IRA, le CRDP et la CGC. On a recueilli les données restantes dans le contexte du programme précisé lors de l'Entente fédérale-provinciale d'exploitation minérale. On a recueilli, en employant un gradiomètre aéromagnétique, des données sur le champ total et des données ÉM dans quatre provinces - la Nouvelle-Écosse (7 170 kilomètres de lignes de levés), le Nouveau-Brunswick (10 600 kilomètres de lignes de levés), le Manitoba (13 800 kilomètres de lignes de levés), et la Saskatchewan (11 000 kilomètres de lignes de levés). La CGC, conformément à une entente conclue avec l'Agence canadienne de développement international (ACDI), a continué à surveiller les détails scientifiques et techniques d'un levé réalisé à forfait au Zimbabwe. La phase aéroportée de ce levé (144 824 kilomètres de lignes de levés) a été complétée au début de 1991.

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[1] Contribution to the Canada-Nova Scotia Cooperation Agreement on Mineral Development 1990-92; the Canada-New Brunswick Cooperation Agreement on Mineral Development, 1990-95; the Canada-Manitoba Partnership Agreement on Mineral Development, 1990-95; the Canada-Saskatchewan Partnership Agreement on Mineral Development, 1990-95.

INTRODUCTION

The aeromagnetic survey program of the Geological Survey of Canada continued in 1991-92 with activity in four existing projects - southern Alberta, the Arctic Continental Margin, Zimbabwe (Teskey et al., 1991), and Northern Yukon and on five new projects which include a federal government/provincial government/industry cost-sharing survey in southern Saskatchewan and Manitoba and four aeromagnetic gradiometer surveys flown under Mineral Development Agreements in Nova Scotia, New Brunswick, Manitoba, and Saskatchewan. Survey activity for 1991 is summarized in Figure 1 and in Table 1.

SOUTHERN ALBERTA

Phase one of the southern Alberta survey, flown for a consortium of oil and mineral companies and the GSC on a cost-sharing basis (Teskey et al., 1991), was delivered to the consortium members in January 1991 and will be released to the public in 1996. Phase two of this project was flown in the summer and fall of 1991 with delivery of data to the consortium members planned for late 1991 or early 1992. These surveys will contribute to current oil exploration, to the GSC's Western Canada Basin Initiative program and to Lithoprobe investigations in Alberta.

ARCTIC CONTINENTAL MARGIN

The Lincoln Sea survey (Teskey et al., 1991) funded by the Institute for Aerospace Research (IAR), the Defence Research Establishment Pacific (DREP) and the GSC was completed in April 1991 with the addition of 4000 line kilometres. A preliminary interpretation report has been published (Nelson et al., 1992). A second area north of Axel Heiberg and Ellesmere islands (Fig. 1) was also flown (12 000 line kilometres) as phase one of a larger survey which is expected to take three years to complete. A progress report on this survey area has also been published (Forsyth et al., 1992).

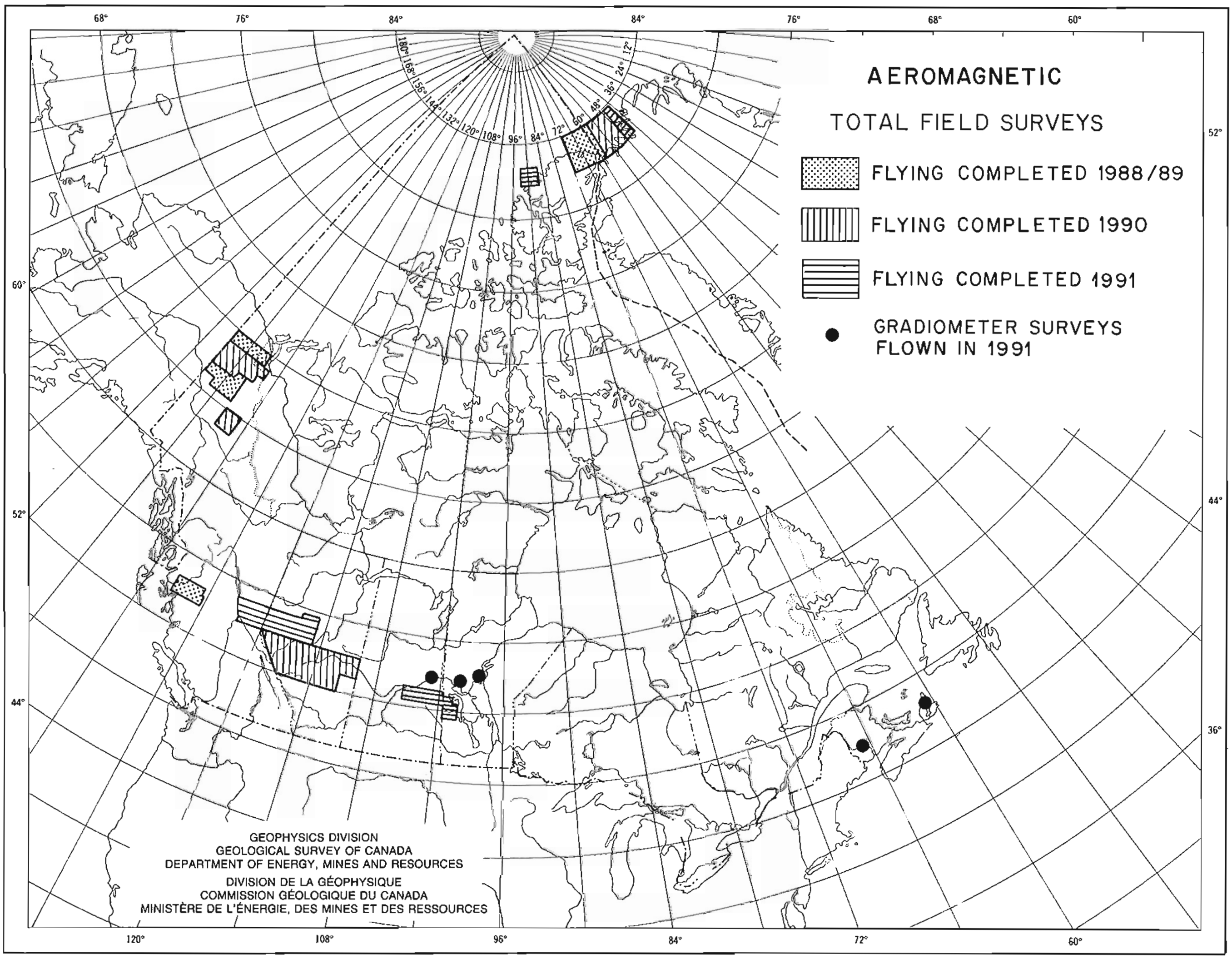

Figure 1. Aeromagnetic surveys in progress, 1991-1992.

ZIMBABWE

Flying of the Zimbabwe aeromagnetic survey (144 824 line kilometres) for which the GSC is acting as "scientific authority" under an agreement with CIDA was completed in January 1991. Compilation is continuing and, although the original target date of late 1991 will not be met, it is expected that final delivery of digital data and maps to CIDA and Zimbabwe will take place in the spring of 1992.

NORTHERN YUKON

Data and maps for the portion of the survey flown in the period 1988-90 under the initial contract (Teskey et al., 1991) were published in June 1991, while compilation for the area flown in 1990-91 under a second contract has progressed during 1991 with release of digital data and map products scheduled for mid-1992.

SOUTHERN SASKATCHEWAN-MANITOBA

Phase one of a four year cost-sharing GSC/Saskatchewan Energy and Mines/industry survey was initiated in southern Saskatchewan in 1991. Three companies pursuing kimberlite exploration are participating in the project and will receive proprietary use of the data for three years before public release of the data by the GSC. A second project, the aeromagnetic survey of southern Manitoba, is being flown

Table 1. Aeromagnetic surveys

Survey	Type	Line Km	Line Spacing	Altitude (m)
Yukon (1988-90)	Aeromagnetic Total Field	28 840	2 km (southern portion) 3 km (northern portion)	2 135 Baro. 2 745 Baro.
Yukon (1990-91)	Aeromagnetic Total Field	35 550	2 km (southern portion) 3 km (northern portion)	300 Radar and 2 135 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
Lincoln Sea (1991-92)	Aeromagnetic Total Field	4 000	4 km	300 Radar
Axel Heiberg (1991-92)	Aeromagnetic Total Field	12 000	4 km	300 Radar
S. Alberta (1991-92)	Aeromagnetic Total Field	73 202	1.6 km	Various Baro. Altitudes
S. Saskatchewan/ Manitoba	Aeromagnetic Total Field	51 250	800 m	150 Radar
Saskatchewan	Gradiometer	11 000	300 m	150 Radar
Manitoba	Gradiometer	13 800	300 m	150 Radar
Cape Breton Island	Gradiometer	7 170	300 m	150 Radar
New Brunswick	Gradiometer	10 600	300 m	150 Radar
Zimbabwe (1990-91)	Aeromagnetic Total Field	144 824	1 km	300 Radar

under the same contract and is sponsored by the GSC and Manitoba Mines and Energy. As well as contributing to the search for kimberlites, the survey will contribute to basement geological mapping and the search for other minerals such as potash.

AEROMAGNETIC GRADIOMETER SURVEYS

Four gradiometer surveys intended to assist detailed geological mapping in areas of specific interest for mineral exploration were carried out in the Cape Breton Highlands of Nova Scotia, the southwest Miramichi Belt of New Brunswick and over the Phanerozoic/Shield margin of northern Manitoba and Saskatchewan under federal-provincial Mineral Development Agreements. The latter two surveys also constitute contributions to the GSC's NATMAP mapping project.

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Études préliminaires de la géologie du Quaternaire des régions de Big Bald Mountain et de Serpentine Lake, Nouveau-Brunswick[1]

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Doiron, A., 1992 : Études préliminaires de la géologie du Quaternaire des régions de Big Bald Mountain et de Serpentine Lake, Nouveau-Brunswick; dans Recherches en cours, Partie D; Commission géologique du Canada, Étude 92-1D, p. 77-82.

Résumé

Les stries glaciaires observées suggèrent que les glaces tardi-wisconsiniennes qui provenaient de l'ouest depuis le centre de dispersion des monts Notre-Dame ou de la ligne de partage glaciaire du nord du Maine étaient déviées vers le nord-est dans la vallée de Curventon-Bathurst par les glaces s'écoulant vers le nord depuis le centre de dispersion de Gaspereau et vers l'ouest depuis le centre de dispersion d'Escuminac. Des sédiments non glaciaires intercalés entre deux tills sont exposés dans les vallées de la rivière South Sevogle et du ruisseau Mullin. Ces coupes montrent l'existence d'une réavancée vers le nord-est du front glaciaire qui pourrait être reliée à la détérioration climatique du Dryas récent. Cette hypothèse reste à démontrer.

Abstract

Observed glacial striations suggest that Late Wisconsinian glaciers flowing from the west and originating from a dispersion centre located in Notre Dame Mountains or the ice divide in northern Maine veered to the northeast in Curventon-Bathurst after the contact with ice flowing from the Gaspereau dispersion centre northwards, and from the Escuminac dispersion centre westwards. Nonglacial sediments interstratified between two tills outcrop in South Seavogle Valley and in Mullin Creek Valley. These sections demonstrate a readvance towards the northeast of the glacier front which might be associated to the climatic deterioration during the Latest Dryas. This hypothesis is unproven to date.

[1] Contribution à l'Entente de coopération Canada - Nouveau-Brunswick sur l'exploitation minérale, 1990-1995.

INTRODUCTION

Au cours de l'été 1991, l'échantillonnage des tills de surface, mené aux fins d'analyses géochimiques et d'étude de provenance et les travaux de cartographie des formations en surface se sont poursuivis durant une seconde année dans les régions de Big Bald Mountain (SNRC 21 O/1) et de Serpentine Lake (SNRC 21 O/2) dans la partie nord et est des hautes-terres de Miramichi, au Nouveau-Brunswick (fig. 1).

Des études portant sur la géochimie des tills de cette région ont été effectuées antérieurement par Fyffe et Pronk (1985) et surtout par Lamothe (1988, 1989, 1990a, 1990b, 1990c, 1990d, 1990e, sous presse). Les travaux de ce dernier s'étendaient sur l'ensemble de la zone tectonostratigraphique de Miramichi reconnue pour son potentiel minéral (Zn, Pb, Cu, Ag). La présente étude géochimique s'inscrit dans la foulée des travaux de Lamothe et porte sur un secteur peu étudié jusqu'à maintenant principalement à cause des difficultés d'accès.

Gauthier (1979, 1980, 1982, 1983) et Rampton et al. (1984) ont effectué des travaux de cartographie des formations en surface du secteur présentement étudié. Le rapport de Gauthier (1982) traite du nord du Nouveau-Brunswick au nord du 47° parallèle alors que celui de Rampton et al. (1984) constitue une synthèse des connaissances portant sur la géologie du Quaternaire de l'ensemble du Nouveau-Brunswick.

LES MICROFORMES D'ÉROSION GLACIAIRE

Les stries et les queues de rat levées dans le cadre de ce projet (fig. 2) présentent dans leur ensemble des orientations similaires à celles relevées par les auteurs antérieurs (Gauthier, 1982, 1983; Rampton et al., 1984; Fyffe et Pronk, 1985; Paradis et al., 1986; Lamothe, sous presse). Dans la partie ouest, les microformes d'érosion sont orientées généralement vers le sud-est et vers l'est. Dans la partie est, les microformes se redressent progressivement vers le nord-est, parallèlement à la bordure est des hautes terres et en suivant l'axe de la vallée de Curventon-Bathurst.

Quelques rares sites de stries montrant un écoulement vers le nord (N354°, N010°) pourraient refléter une incursion glaciaire précoce vers le nord issue soit du centre de dispersion de Gaspereau au Wisconsinien supérieur ou soit d'un centre de dispersion semblable au Gaspereau mais plus ancien (fig. 3). En effet, Lamothe (sous presse) signale que plus au sud, dans la région de Hayesville, des cannelures orientées selon un axe nord-sud sont recoupées par des

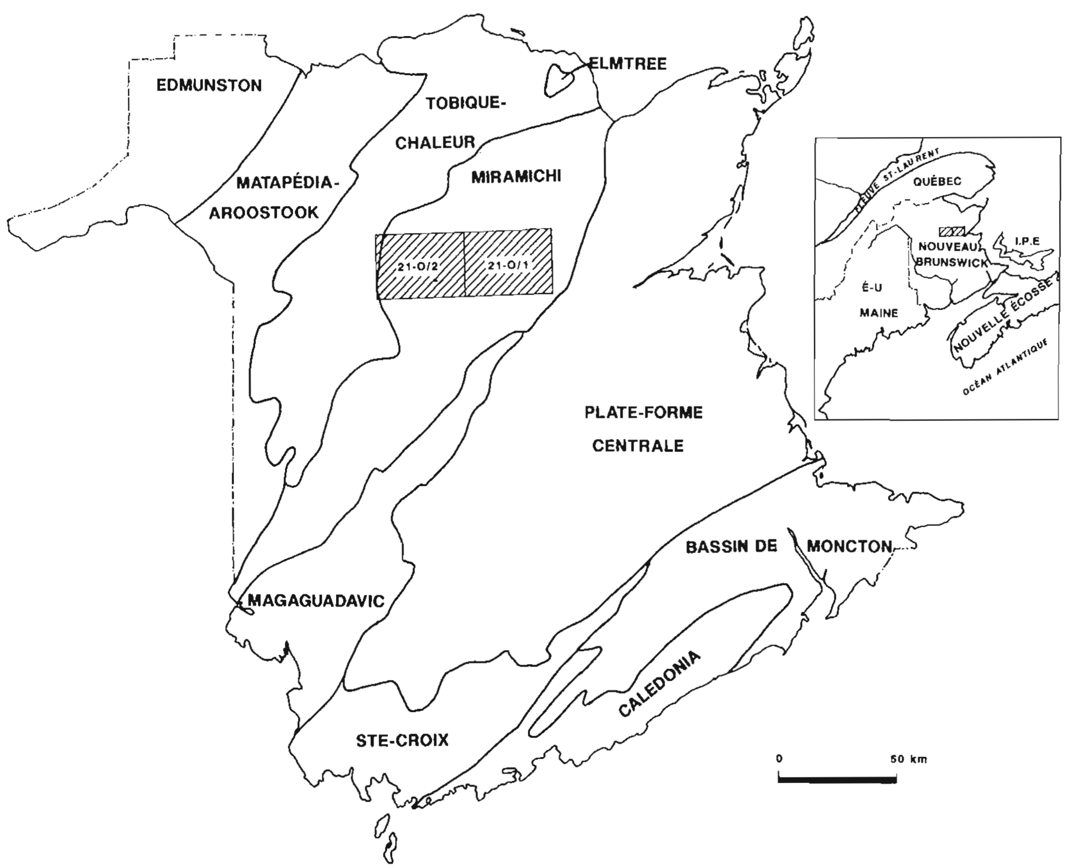

Figure 1. Carte de localisation montrant les zones tectonostratigraphiques du Nouveau-Brunswick (d'après Ruitenberg et al., 1977). La région étudiée est hachurée.

microformes plus récentes orientées vers le nord-est. Il en déduit que l'écoulement vers le nord serait antérieur au Wisconsinien supérieur.

Dans la partie est de la zone étudiée, il est probable qu'au Wisconsinien supérieur, pendant la phase de Chinecto (Rampton et al., 1984; Gauthier, 1983), les glaces qui s'écoulaient vers le sud-est et l'est depuis les centres de dispersion des monts Notre-Dame ou depuis la ligne de partage glaciaire du nord du Maine étaient progressivement déviées vers le nord-est dans la vallée de Curventon-Bathurst suite à leur convergence avec les glaces provenant à la fois du sud depuis le centre de dispersion Gaspereau et de l'est depuis le centre de dispersion d'Escuminac.

Dans la partie nord-est de la région étudiée, de nombreux recoupements de stries montrent que l'écoulement glaciaire qui s'effectuait vers le nord-est (N040°) s'est progressivement réorienté vers l'est (N095°). Il est possible que pendant la phase plus tardive de Bantalor, le retrait progressif des glaces provenant des centres de dispersion d'Escuminac et de Gaspereau ait permis aux glaces provenant de l'ouest de s'écouler librement vers l'est. Ces glaces pouvaient provenir des centres de dispersion des monts Notre-Dame ou de la ligne de partage glaciaire du nord du Maine ou encore, pour la partie est du secteur étudié, de centres de dispersion locaux situés dans la bordure est des hautes terres de Miramichi (Rampton et al., 1984; Rappol, 1989; Pronk et al., 1989; Lamothe, sous presse).

LITHOSTRATIGRAPHIE

À la bordure sud-est de la région étudiée, des sédiments non glaciaires intercalés entre deux tills ont été observés dans deux coupes naturelles situées en bordure de la rivière South Sevogle et du ruisseau Mullin (fig. 4). L'intérêt de ces coupes s'explique par l'apparente rareté des coupes à tills multiples dans la région.

Description préliminaire des unités lithostratigraphiques

Coupe de la rivière South Sevogle

Les unités lithostratigraphiques de cette coupe se décrivent ainsi depuis la base (fig. 4) :

Unité 1 : till inférieur

La base de l'unité n'est pas exposée. La partie visible atteint une épaisseur d'au moins 3,2 m. Il s'agit d'un till massif et compact à matrice de silt sablo-argileux. Les cailloux occupent environ 15% du volume du till et leur taille varie de quelques millimètres à 20 cm de diamètre. Plusieurs sont striés et leur arrondi varie d'anguleux à subarrondi. La nature pétrographique des cailloux se résume principalement à des phyllades cambro-ordoviciennes, des volcaniques et des granites. Le till humide est de couleur gris olive (5Y 4/2). Une fabrique de till effectuée à 1,6 m sous la limite supérieure de l'unité montre des pics d'azimut et de plongée vers le N107°/8° pour les axes A et vers le N153°/70° pour les axes C. Une seconde fabrique de till localisée à 30 cm sous la limite supérieure de l'unité montre des pics d'azimut et de plongée vers le N118°/8° pour les axes A et vers le N233°/67° pour les axes C. Une interprétation préliminaire de ces fabriques les associerait à une glace en régime d'extension s'écoulant vers le sud-est. Cette hypothèse reste à être confirmée, entre autres, à l'aide des comptages pétrographiques qui seront effectués sur les cailloux. Il existe en effet dans le secteur quelques sites où les stries sont orientées vers le sud-est mais les autres indiquent majoritairement un écoulement vers le nord-est.

Unité 2 : sédiments fluvio-glaciaires

Le contact entre cette unité et le till de base est franc. Il s'agit de sables et de graviers stratifiés. L'épaisseur des lits et leurs caractéristiques granulométriques sont très variables. Il s'agit d'une séquence fluviatile probablement

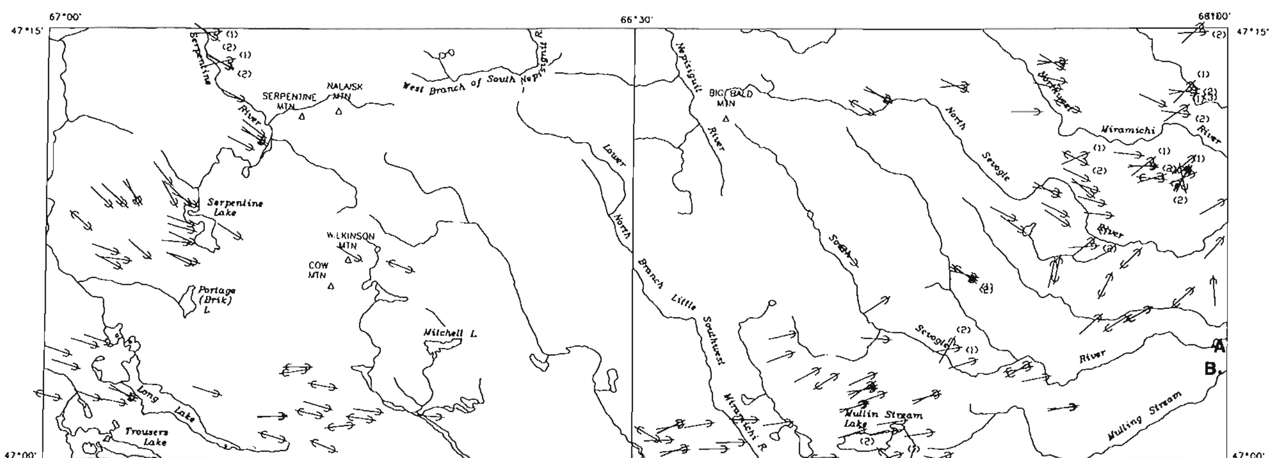

Figure 2. Microformes d'érosion glaciaire levées au cours de ce projet; (1), (2) et (3) : chronologie relative des diverses familles de stries se recoupant en un même site; 1 est plus ancien que 2, etc.. A: localisation de la coupe de la rivière South Sevogle; B: localisation de la coupe du ruisseau Mullin.

fluvio-glaciaire. Dans quelques horizons, les particules sont fortement cimentées par des oxydes de manganèse noirs ou des oxydes de fer rouges. Il est probable que ces zones oxydées sont reliées à des paléo-surfaces hydrostatiques qui ont été ultérieurement rabattues suite à l'encaissement du cours d'eau.

Unité 3 : rythmites glacio-lacustres et turbidites

Le contact avec l'unité sous-jacente est subhorizontal, net et sans lacune apparente. La puissance des lits grossiers varie de 2 cm à 6 cm tandis que celle des lits fins est comprise entre 0,3 cm et 1,3 cm. Au sommet de la séquence, les strates en contact avec le till n'ont pas subi de déformations glaciotectoniques apparentes.

Au centre de cette séquence est inséré un lit de 60 cm d'épaisseur constitué de silt argileux et de graviers atteignant un diamètre de 1,5 cm. Ces sédiments sont très peu triés, presque diamictiques, et la stratification subhorizontale est à peine perceptible. Il s'agit possiblement de turbidites mises en place dans un milieu glacio-lacustre proximal dans un contexte de réavancée glaciaire.

Unité 4 : till supérieur

Le contact entre cette unité et les sédiments sous-jacents est souligné par un pavage de blocs dont la taille est d'environ 30 cm. Le till se présente comme un diamicton compact mais

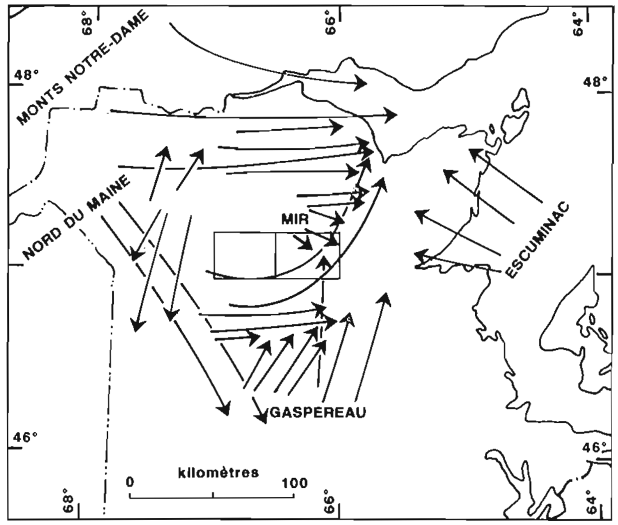

Figure 3. Directions d'écoulement glaciaire au Pléistocène dans le nord et le centre du Nouveau-Brunswick (modifié d'après Rampton et al., 1984 et Lamothe, sous presse).

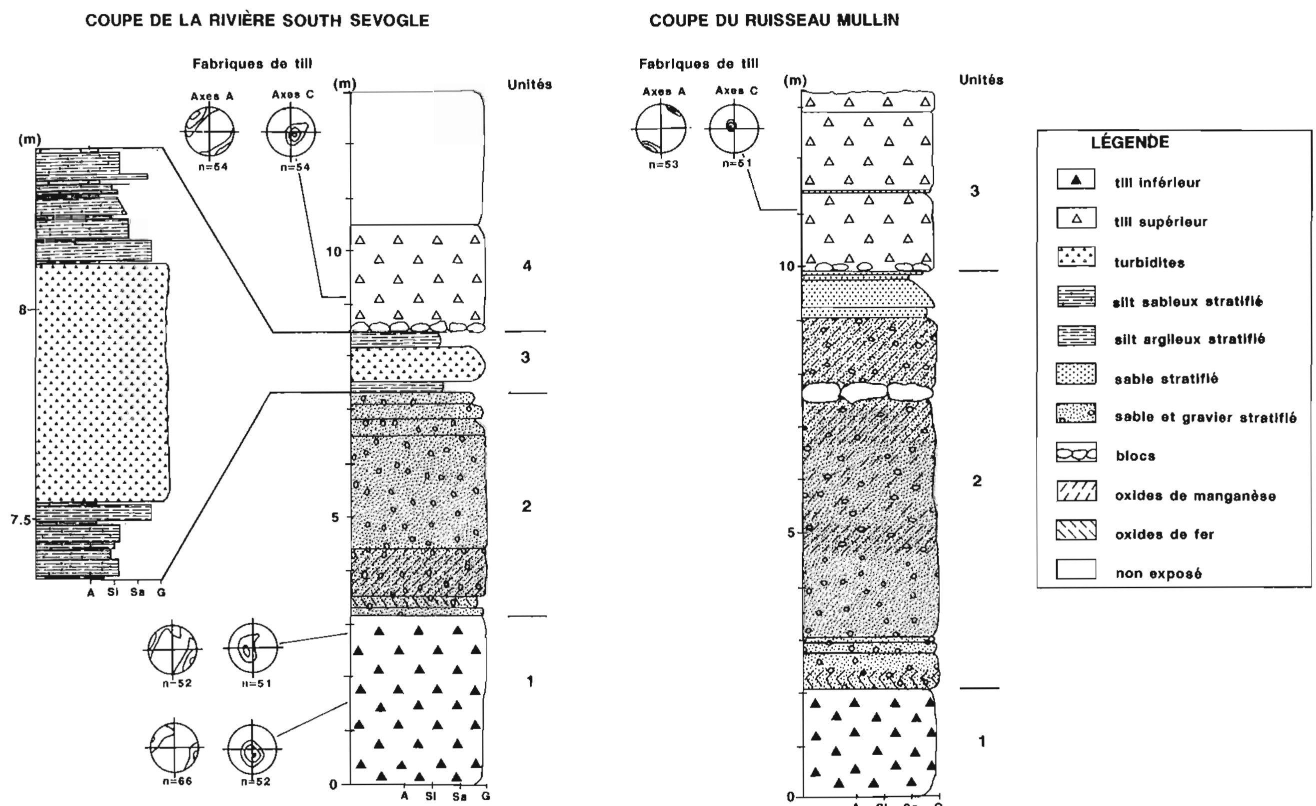

Figure 4. Coupes lithostratigraphiques de la rivière South Sevogle et du ruisseau Mullin.

à l'intérieur duquel de petites vacuoles tapissées d'une fine pellicule d'argile sont fréquentes. La matrice est sablo-silteuse. Le till humide est de couleur brun olive clair (2.5Y 5/4). Le till exposé atteint une puissance de 2 m et sa partie supérieure n'est pas exposée. Une fabrique de till effectuée à 80 cm au dessus de la base de l'unité montre un pic vers le N315°/15° pour l'axe A et vers le N315°/71° pour l'axe C. Prise isolément, cette fabrique est peu significative mais, en y associant les familles de stries connues localement, elle pourrait être associée à une glace s'écoulant vers le sud-est en régime d'extension ou vers le nord-est en régime de compression. Si cette unité est corrélative du till supérieur à la coupe du ruisseau Mullin décrit ci-dessous, sa fabrique reflèterait alors un écoulement vers le nord-est.

Coupe du ruisseau Mullin

Unité 1 : till inférieur

La base de l'unité est enfouie et la partie visible atteint une épaisseur de 2,5 m. Il s'agit d'un diamicton compact et massif à matrice silto-argileuse. Humide, il est de couleur gris olive (5Y 4/2). La majorité des cailloux sont anguleux, souvent altérés et proviennent des phyllades grises cambro-ordoviciennes d'origine locale. Sont également présents quelques granites roses et amphibolites arrondis ou subarrondis. Il est possible que les phyllades étaient altérées avant même leur incorporation dans le till. Ce diamicton est très différent, de par sa nature pétrographique, de l'unité 1 de la coupe de la rivière South Sevogle. Il s'agit probablement d'un faciès local du till provenant de l'incorporation de colluvions.

Unité 2 : sédiments fluvio-glaciaires

Le contact avec l'unité sous-jacente est franc et subhorizontal. Cette unité est composée de sables et de graviers stratifiés. L'épaisseur des lits et les distributions granulométriques sont très variables. L'ensemble a une puissance de 5,7 m. Il s'agit d'une séquence fluviatile, probablement fluvio-glaciaire. Des cailloux dont le grand axe se redresse vers l'est-nord-est suggèrent un écoulement dans cette direction. Il existe quelques niveaux fortement cimentés par des oxydes de manganèse noirs ou par des oxydes de fer rouges. Tout comme pour la coupe précédente, il est probable que ces oxydes soient liés à des paléo-surfaces hydrostatiques.

Unité 3 : till supérieur

Le contact avec l'unité inférieure est marqué par un pavage de blocs. La base de ceux-ci est en contact avec les sables et graviers alors que leur partie supérieure est dans le till, est striée et présente des queues de rats indiquant un écoulement vers le nord-est (N054°, N040°). Le till est compact et massif et sa matrice est silto-argileuse. La couleur du till humide passe de gris olive (5Y 4/2) à la base à brun olive clair (2.5Y 5/4) vers le sommet. Les cailloux sont subarrondis et subanguleux et ils représentent environ 10% à 15% du volume du till. Ils sont principalement constitués de phyllades cambro-ordoviciennes avec une faible proportion de granite. Un lit de sable grossier de 0,5 cm à 3 cm d'épaisseur est inséré au centre de l'unité. Le sable y est cimenté par des oxydes de manganèse noirs. Une fabrique de till effectuée à 80 cm au dessus de la base de l'unité montre un pic vers le N216°/4° pour l'axe A et vers le N315°/71° pour l'axe C. En jumelant cette fabrique aux queues de rats présentes sur les blocs du pavage, il semble évident que ce till est associé à une glace s'écoulant vers le nord-est. Cette glace serait possiblement responsable des formes fuselées orientées vers le nord-est dans la région (Gauthier, 1982; Rampton et al., 1984, Lamothe, sous presse)

INTERPRÉTATION PALÉOGÉOGRAPHIQUE ET CONCLUSION

Pour les deux coupes étudiées, les sables et graviers fluvioglaciaires insérés entre les deux tills sont associés à un retrait glaciaire. Étant donné l'absence de matériel datable, le laps de temps séparant les deux événements glaciaires ne peut être évalué présentement mais il pourrait être relativement court. Il n'y a pas d'horizons de paléosols développés dans la partie supérieure de cette unité mais il est vrai que la réavancée glaciaire ait pu les décaper. Les horizons oxydés pour leur part ne sont pas associés à des paléosols et ils se sont probablement développés au cours de l'Holocène.

Les rythmites glacio-lacustres se sont probablement déposées dans un lac proglaciaire barré par une langue glaciaire lors de la réavancée.

Rampton et al. (1984) estiment que le retrait glaciaire final s'est effectué dans ce secteur il y a environ 13 000 ans. Par contre, près de Hayesville à 80 km au sud-est des coupes décrites ci-haut, Lamothe (sous presse) signale la présence, sous le till de surface, de silts laminés contenant de la matière organique. Ces débris organiques comprennent des fragments de plante et d'insectes et ont été datés à 11 500 ± 150 ans BP. Lamothe en déduit que la déglaciation finale des hautes terres de Miramichi serait plus récente que ne l'estimaient Rampton et al.(1984) et suggère que le till de surface de la coupe d'Hayesville pourrait refléter la détérioration climatique du Dryas récent qui est enregistrée géologiquement à plusieurs endroits dans les Maritimes (Mott et al., 1986; Stea et Mott, 1990). Le till supérieur de la rivière South Sevogle et celui du ruisseau Mullin pourraient être corrélatifs du till de surface de la coupe d'Hayesville et avoir été mis en place au Dryas récent. Cette hypothèse est intéressante, puisqu'elle indiquerait une réavancée glaciaire d'échelle régionale, mais celle-ci reste à démontrer.

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Projet 900008 de la Commission géologique du Canada

Contexte géologique du gîte polymétallique de Champagne, Appalaches du Québec

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Résumé

Les lithologies de la région avoisinant l'amas polymétallique de Champagne appartiennent à trois des quatre formations constituant le groupe de Magog. Ainsi, on y retrouve, de la base vers le sommet, les formations de Frontière, d'Etchemin et de Beauceville. Les argilites noires de la formation de Beauceville encaissent le gîte de Champagne.

L'évolution du groupe de Magog, depuis les mudslates rouges puis vertes de l'Etchemin aux argilites noires du Beauceville, témoigne du caractère réducteur grandissant du milieu sédimentaire. Par ailleurs, une activité volcanique importante se manifeste au moment où la formation de Beauceville se constitue. On y observe des lits de volcanoclastite intercalés dans les argilites noires. Les premières manifestations d'activité hydrothermale apparaissent dès le dépôt de la formation d'Etchemin. En effet, des fragments de sulfures massifs sont observés dans une unité de volcanoclastite. L'activité hydrothermale reprend et culmine au niveau de la formation de Beauceville et génère l'amas sulfuré de Champagne.

Abstract

The lithologies around the Champagne polymetallic deposit belong to three of the four formations which constitute the Magog Group, i.e. from the base to the top, the Frontière, Etchemin, and Beauceville formations. The black argillites of the Beauceville Formation host the Champagne deposit.

The evolution of the Magog Group from red and green mudslates of the Etchemin to black argillites of the Beauceville testify to an increasingly reducing sedimentary environment. Also, important volcanic activity occurred during the deposition of the Beauceville Formation. Volcanoclastite beds are intercalated in the black argillites. The first manifestations of hydrothermal activity appear during the deposition of the Etchemin Formation. Fragments of massive sulphide are observed in a volcaniclastite bed. Hydrothermal activity started anew and culminated during the deposition of the Beauceville Formation and resulted in the formation of the Champagne massive sulphide deposit.

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INTRODUCTION

Depuis 1823, date de la première découverte d'or alluvionnaire dans les Appalaches du Québec (Douglas, 1864; Gauthier et al., 1989), de nombreux travaux d'exploration pour ce métal y ont été entrepris. Le gîte polymétallique de Champagne fut mis à jour en 1952 à la suite de la découverte d'un bloc erratique minéralisé. En 1953, Panet Metals Corporation Limited estime, par des travaux de géophysique et de forage, les réserves du gîte de Champagne 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 faits. Depuis 1985, Golden Hopes Mines Limited effectue de nouveaux travaux d'exploration sous la direction de R.E. Schaaf et Associés. Les études antérieures ont pour la plupart été faites sur le gîte de Champagne. Dans le présent travail, nous nous proposons d'examiner les conditions géologiques qui ont présidé lors du dépôt des roches du groupe de Magog et, plus particulièrement celles qui ont présidé lors de la formation de l'amas sulfuré de Champagne.

Le secteur à l'étude se situe à 105 kilomètres au sud-est de la ville de Québec (fig. 1). Le périmètre cartographié s'étend sur 19 km^2 autour du gîte de Champagne. Le gîte de Champagne est un corps stratiforme de sulfures massifs encaissé dans les argilites noires de la formation de Beauceville, laquelle appartient au groupe de Magog. Les sédiments du groupe de Magog sont interprétés comme étant des dépôts de bassin d'avant-arc euxinique d'âge Ordovicien moyen (Llandeilien-Caradocien). Ce bassin se serait développé entre des complexes ophiolitiques obductés à l'ouest et un arc volcanique au sud-est (Laurent, 1987; Cousineau, 1988; Desbiens, 1988; Bossé et al., 1991).

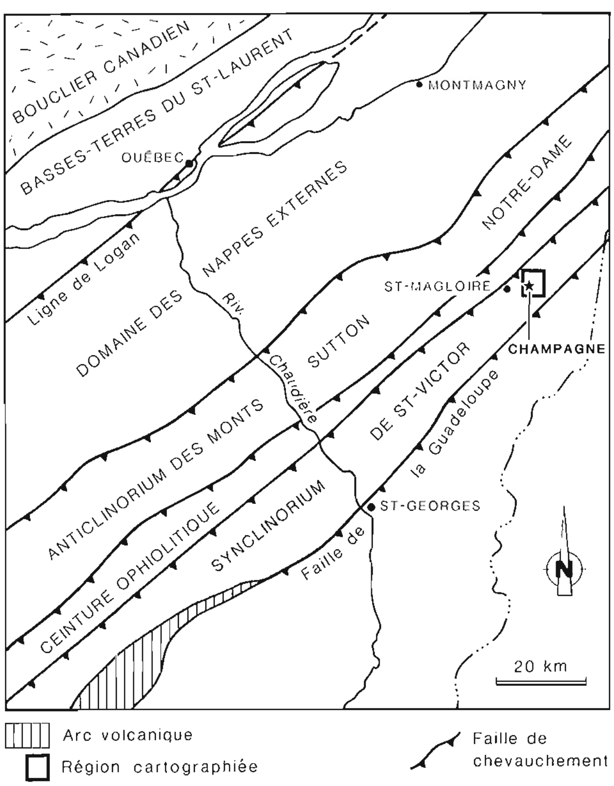

Figure 1. Carte de localisation du secteur cartographié, du gîte de Champagne et du synclinorium de Saint-Victor dans les Appalaches du Québec.

CONTEXTE STRATIGRAPHIQUE

Cousineau (1986, 1988) a subdivisé le groupe de Magog en quatre unités lithostratigraphiques, qui sont de la base au sommet, les formations de Frontière, d'Etchemin, de Beauceville et de St-Victor. Dans la région cartographiée, seulement trois des quatres formations sont présentes, soit celles de Frontière, d'Etchemin et de Beauceville.

Les contacts géologiques entre les différentes formations n'ont pas été observés sur le terrain. Pour cette raison, les contacts géologiques de la figure 2 sont soit tracés à partir de levés électromagnétiques aéroportés (Jobin et al., 1984; MERQ, 1987), soit tracés à équidistance des derniers affleurements de chacune des deux formations respectives. Pour Cousineau (1986, 1988) qui a pu observer les limites des formations ailleurs dans la région, il s'agirait de contacts concordants subséquemment repris par des failles de chevauchement orientées vers le nord-est.

Une coupe schématique des unités lithologiques identifiée dans la région cartographiée est présentée à la figure 3. L'épaisseur des différentes unités y est apparente. Les épaisseurs vraies sont approximativement les suivantes: Frontière, 390 mètres; Etchemin, 1160 mètres; Beauceville, 550 mètres.

La formation de Frontière est constituée de grès de couleur gris moyen à patine blanchâtre interstratifiés de mudslate gris moyen à gris foncé. Les grès à grain moyen forment généralement des bancs massifs et homogènes de 10 à 20 centimètres d'épaisseur. A quelques kilomètres au sud-ouest de la région cartographiée, des lits obliques, des laminations parallèles, des granoclassements normaux, des lits tronqués et des contacts discordants sont observés dans une séquence de grès, de mudstone et d'argilite appartenant à la formation de Frontière. Une polarité normale vers le sud-est y a été observée. A l'instar de Cousineau (1986), nous avons constaté qu'il s'agit de wackes lithiques composés de grains de quartz (30%), de fragments de roches (15%) et de feldspaths (plagioclase et microcline, 40%). Les grains sont subanguleux à subarrondis. On retrouve des fragments de volcanite felsique constitués d'un agrégat de quartz, de feldspath et de séricite en proportions variables et des fragments de roches constitués de matériel détritique fin (argiles et mudstones). Les grains sont jointifs et on retrouve environ 10 à 15% de matrice. Cette matrice contient presque toujours un peu de chlorite et d'autres matériaux détritiques fins. L'examen en lame-minces polies montre la présence en traces de grains de chromite et de leucoxène. Les mudslates gris forment généralement des interlits de quelques millimètres d'épaisseur régulièrement interstratifiés avec les grès.

A la base de la formation d'Etchemin, une alternance de lits de mudslate et de grès, tous deux siliceux, gris moyen verdâtre à patine beige, est observée. Les mudslates et les grès forment des bancs massifs de 10 à 15 centimètres d'épaisseur.

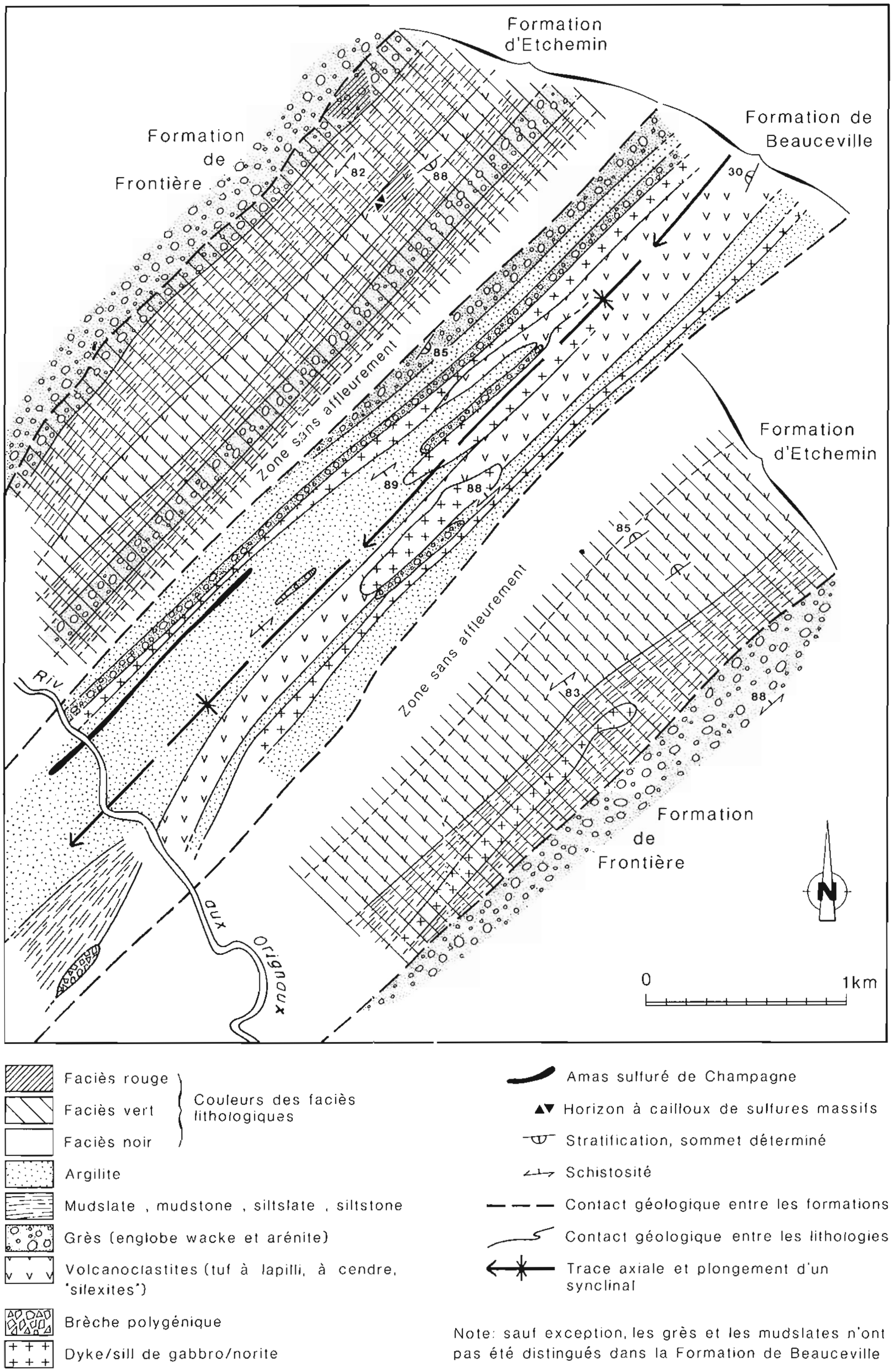

Figure 2. Carte géologique simplifiée du périmètre cartographié autour du gîte de Champagne. Subdivisions stratigraphiques du groupe de Magog, selon Cousineau (1986, 1988).

Un granoclassement normal indique une polarité vers le sud-est. Ces mudslates et ces grès présentent localement un stockwerk constitué de quartz blanc saccharoïdal, de sphalérite, de galène et de pyrrhotite. Près de la base de la formation d'Etchemin, on observe localement des bancs concordants décimétriques de mudslate violacé. Ces mudslates violacés constituent un bon marqueur stratigraphique. Une séquence de tufs à lapillis interlités avec des tufs à cendres de couleur gris moyen verdâtre à patine beige surmonte stratigraphiquement les mudslates et les grès. Cette séquence tufacée est silicifiée et possède une altération noire en surface. Elle contient une bande discontinue de pyrite massive de cinq centimètres d'épaisseur ainsi qu'un niveau à nodules de pyrite massive pouvant atteindre 15 cm d'épaisseur. On y retrouve aussi un mince horizon de 3 à 4 cm de mudslate rouge. Toute cette unité de volcanoclastite peut contenir jusqu'à 2% de pyrite disséminée dans les plans de la schistosité. On remarque par endroits des laminations parallèles. Ces tufs sont surmontés d'une alternance de mudslate et de grès immatures siliceux à patine blanchâtre. On y retrouve des clastes subarrondis à subanguleux dont la taille peut atteindre 7 centimètres de diamètre.

La formation de Beauceville débute, du coté sud-est du synclinal, par un banc de brèche polygénique (fig. 4) minéralisée en chalcopyrite et en pyrrhotite. La brèche est non organisée et consiste en fragments de mudslate et d'argilite laminés ou massifs (1mm à 1 m de diamètre). La matrice silicifiée est constituée de fragments de roche volcanique et de mudslate, ainsi que de cristaux de feldspaths et de quartz qui baignent dans une pâte riche en quartz, chlorite, feldspath et micas. Du coté nord-ouest du synclinal, la formation de Beauceville débute par un conglomérat verdâtre polymicte à matrice grèseuse passant graduellement à des wackes arkosiques gris moyen à patine gris pâle. Selon Burzynsky (communication personnelle, 1990), ce faciès conglomératique est observé à plusieurs endroits près du contact avec la formation d'Etchemin sous-jacente. Les wackes passent abruptement à des mudslates siliceux. C'est dans cette unité que l'on retrouve un grand nombre de figures sédimentaires telles que des granoclassements normaux, des glissements ("slumps"), des structures en flamme et des laminations parallèles, ondulantes, obliques et entrecroisées (fig. 5). Toutes ces structures indiquent une polarité vers le sud-est. Un mince horizon d'argilite et de mudslate noirs est coincé entre ces mudstones et un tuf felsique à lapillis. Ces tufs contiennent des cristaux de feldspaths subanguleux et des minéralisations disséminées de pyrite, de pyrrhotite et de chalcopyrite. Latéralement, les tufs deviennent des silexites. Ces dernières sont constituées d'aiguilles de minéraux opaques dans une matrice aphanitique siliceuse. Un mudstone surmonte les tufs. Celui-ci passe ensuite aux argilites noires typiques de cette formation (fig. 6). Les argilites noires à patine grise forment des séquences de plusieurs mètres d'épaisseur dans lesquelles s'intercalent des niveaux d'argilite graphiteuse. La pyrite contenue dans ces argilites (jusqu'à 5%) se présente soit sous forme de cubes disséminés, soit en de fins lits de pyrite framboïdale. L'argilite se débitent en lamelles de 0.5 à 2 cm d'épaisseur. On note un horizon où l'argilite est lessivée et minéralisée en pyrite et en sphalérite. Elle prend alors une teinte gris moyen bleuté en cassure fraîche. La formation de Beauceville se termine par une unité de grès immature gris foncé à noir dont les fragments (2 à 3%) sont principalement constitués d'argilite et de mudstone siliceux.

Roches intrusives

Des filons-couches et des dykes de norite recoupent tous les sédiments du groupe de Magog et plus particulièrement les sédiments de la formation de Beauceville. L'âge de ces intrusifs est indéterminé. Les roches sédimentaires entourant ces intrusifs présentent toutes des cornéennes silicifiées ("hornfels"). L'étude pétrographique démontre que ces intrusifs sont composés de plagioclase (80%), orthopyroxène (10%), titanite (7%) et quartz (3%). La chlorite et la calcite

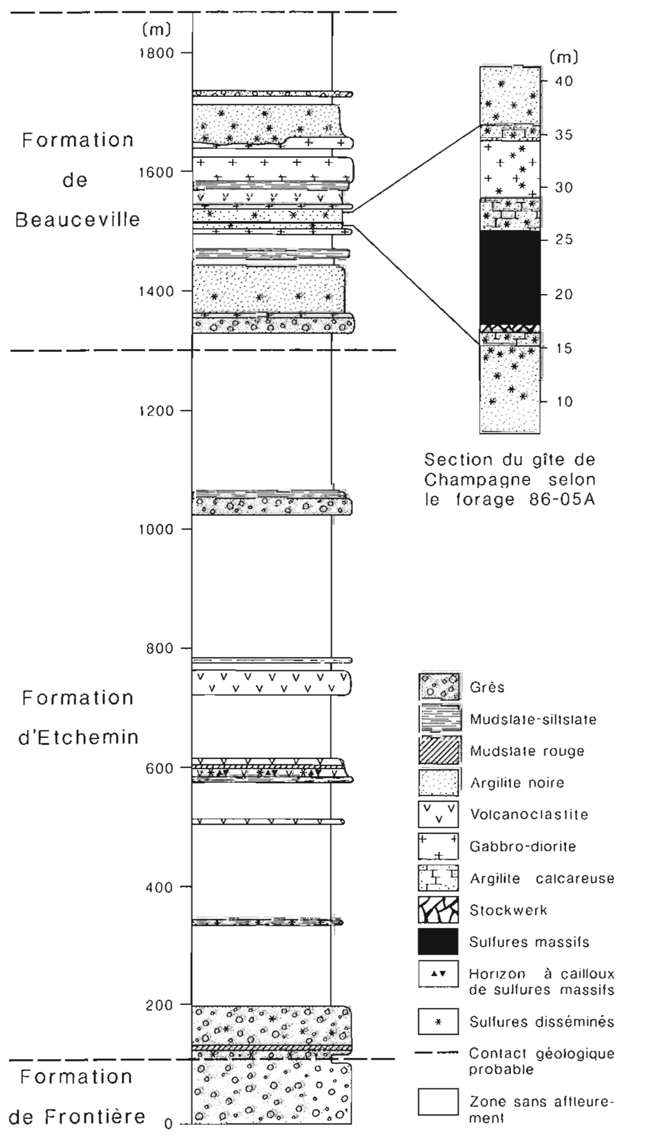

Figure 3. Coupe schématique du groupe de Magog dans la région cartographiée. Subdivisions stratigraphiques du groupe de Magog, selon Cousineau (1986, 1988).

constituent les principaux produits d'altération. Les roches intrusives sont principalement minéralisées en pyrrhotite (<5%). Sur les cartes aéromagnétiques, les filons-couches et les dykes forment une série de crêtes magnétiques.

CONTEXTE STRUCTURAL

Les déformations régionales sont associées à l'orogenèse acadienne. Une schistosité pénétrative de direction NE-SW associée à cette orogenèse affecte toutes les unités du groupe de Magog (fig. 7). A quelques rares endroits, une schistosité antérieure à cette dernière fut observée. La schistosité pénétrative a partiellement remobilisé la minéralisation, ainsi la minéralisation suit la stratification mais s'étire selon les plans de clivage dominants.

VARIATIONS LOCALES DES FACIÈS VOLCANO-SÉDIMENTAIRES ET HYDROTHERMAUX

Les travaux menés à l'échelle régionale par Laurent (1987), Cousineau (1986, 1988), Desbiens (1988) et Godue (1988), pour ne citer que ces auteurs, démontrent que le contexte géodynamique du groupe de Magog en est un de bassin d'avant-arc euxique coincé entre des lambeaux ophiolitiques de la formation de Saint-Daniel au nord-ouest et des copeaux de l'arc volcanique d'Ascot-Weedon au sud-est.

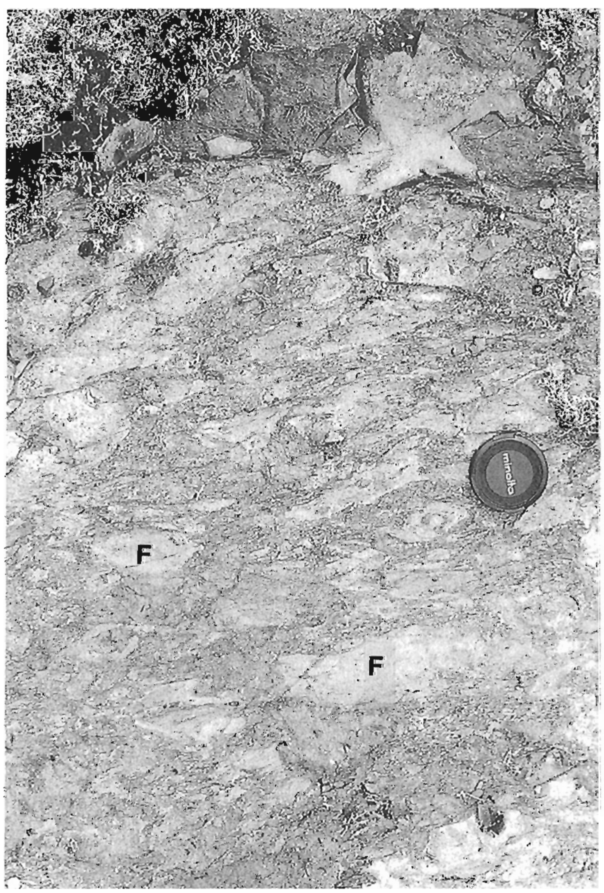

Figure 4. Brèche polygénique située à la base de la formation de Beauceville. F = fragment.

L'examen détaillé de l'empilement volcano-sédimentaire des environs du gîte de Champagne, nous permet de préciser davantage les caractères physiques de ce bassin à l'approche d'un amas sulfuré. Nos principales observations se résument comme suit:

(1) Variations verticales des faciès

Une certaine cyclicité dans la variation verticale des faciès sédimentaires existe à travers les formations du groupe de Magog. Ainsi, dans la formation de Frontière, on observe une interstratification régulière de lits de grès et de mudslate; dans la formation d'Etchemin, on observe une alternance de lits de mudslate, de grès et de volcanoclastite; et dans la formation de Beauceville, la séquence d'argilite est interstratifiée avec des mudslates et des volcanoclastites. Sur environ 50 mètres, la séquence à la base du Beauceville (conglomérat-wacke-mudslate-argilite) est remarquablement bien rythmée et les pulsions sédimentaires y sont rapides et régulières, c'est-à-dire, que les lits de quelques centimètres d'épaisseur s'empilent les uns sur les autres pour former une séquence d'environ 5 mètres d'épaisseur qui se répète plusieurs fois. La séquence débute avec un conglomérat de base qui passe graduellement à un wacke arkosique. Ce banc est suivi d'un lit de mudslate siliceux montrant de nombreuses structures sédimentaires typiques de turbidites classiques. Finalement, la séquence se termine avec un mince horizon d'argilite et de mudslate noirs. Toute cette séquence qui se répète plusieurs fois, possède les divisions classiques a/c/e de Bouma. Dans la formation d'Etchemin, les lithologies forment des bandes plus épaisses que leurs équivalents lithologiques de la formation de Beauceville. Ceci dénote des changements de faciès plus lents au niveau de l'Etchemin.

Les formations d'Etchemin et de Beauceville montrent un changement prononcé au niveau de leur potentiel d'oxydo-réduction. Ainsi, les faciès rouges (mudslates violacés)

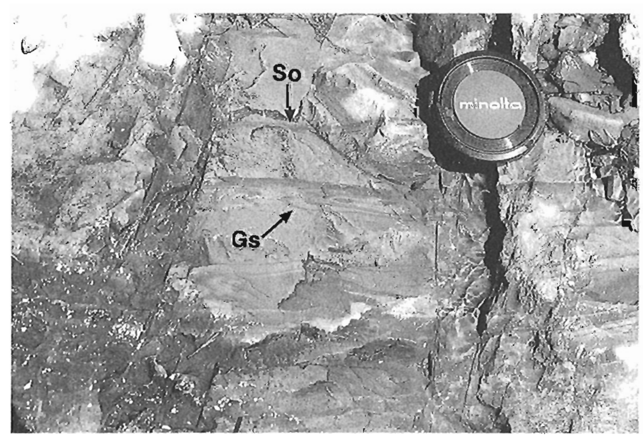

Figure 5. Glissements synsédimentaires (Gs) dans les mudstones siliceux.

ne sont présents que près de la base de la formation Etchemin. Ils s'estompent pour faire place à des faciès verts, réduits. Cette tendance réductrice se fait plus prononcé en montant dans la séquence stratigraphique. En effet, les faciès silicoclastiques situés à la base de la formation de Beauceville passent rapidement à des argilites, des mudslates et des volcanoclastites noirâtres. Les argilites noires forment jusqu'à 65% de la Formation de Beauceville. De minces niveaux d'argilite graphiteuse s'intercalent dans la séquence d'argilite noire. Les faciès noirâtres ainsi que les niveaux graphiteux (présence de matière organique) dénotent un milieu de sédimentation calme, sapropélique, où l'apport en éléments terrigènes est plus faible que dans les autres formations du groupe de Magog (Cousineau, 1988). Ce passage à des faciès sapropéliques correspond possiblement à un approfondissement du bassin sédimentaire caractérisé par un apport terrigène nettement plus modeste (Jébrak et Gauthier, 1991).

Les faciès volcanoclastiques sont présents aussi bien dans la formation d'Etchemin que dans celle du Beauceville. Cependant, ils sont plus abondants et surtout plus grossiers dans la formation de Beauceville, comme en témoigne un affleurement de brèche polygénique dans les environs du gîte de Champagne et dont les clastes atteignent la taille de bombes (fig. 4).

La formation d'Etchemin montre les premiers signes d'activité hydrothermale. Il s'agit d'un mince lit de sulfures massifs syngénétiques et de fragments de sulfures contenus dans un horizon de tuf. L'activité hydrothermale reprend à la base de la formation de Beauceville où elle se manifeste sous la forme d'une brèche minéralisée. L'activité hydrothermale semble culminer plus haut dans la séquence stratigraphique, par exemple, dans la formation de Beauceville, avec le dépôt de l'amas sulfuré de Champagne. Le tarissement de cette venue hydrothermale est souligné par l'apparition d'un mudstone noir à nodules de carbonates dans les quelques mètres qui surplombent l'amas sulfuré. Une zonalité métallifère souligne dans le détail cette transition (Gauthier et al., 1989; Bossé et al., 1991). Ainsi, on remarque un enrichissement graduel en or, en argent et en arsenic de la base vers le sommet de l'amas sulfuré, ainsi qu'un enrichissement en barium dans le mudstone à nodules de carbonates.

(2) Variations latérales des faciès

La variation latérale des faciès la plus remarquable dans les environs du gîte de Champagne est le passage d'une brèche polygénique grossière à des tufs fins. La densité des affleurements ne nous permet pas d'observer la transition entre ces deux faciès volcanoclastiques. Cependant, la continuité des faciès sédimentaires situés de par et d'autre de ce niveau, corrobore l'hypothèse selon laquelle il s'agit d'une variation granulométrique rapide à l'intérieur d'un même horizon stratigraphique.

DISCUSSION

Le bassin sédimentaire dans lequel s'est déposé le groupe de Magog était un milieu propice à la formation et à la préservation d'amas sulfurés exhalatifs, communément appelés SEDEX. Cependant, on doit noter ici que le contexte géodynamique de ce bassin diffère sensiblement de celui des gîtes SEDEX "classiques" tels que Rammelsberg et Meggen en Allemagne et ceux du bassin de Selwyn au Yukon (Gustafson et Williams, 1981; Large, 1983; Abbott et Turner, 1991). En effet, ces derniers sont situés dans des bassins épicratoniques à intracratoniques, tandis que celui de Magog repose sur un fond océanique. Le contexte géologique et géodynamique du bassin de Magog pourrait expliquer, dans une certaine mesure, les caractéristiques métallifères (par exemple, riche en Au, Ag et Cu) de l'amas sulfuré de Champagne.

Un de nos objectifs principaux dans l'étude de la région du gîte de Champagne, est de vérifier si les variations verticales et latérales de faciès peuvent annoncer et expliquer, dans une certaine mesure, la présence d'amas sulfurés tels que celui de Champagne. Les observations que nous avons faites jusqu'à ce jour, suggèrent que oui. Les paramètres suivants sont considérés favorables:

Figure 6. Argilites noires typique de la formation de Beauceville A noter, la fissilité de la roche.

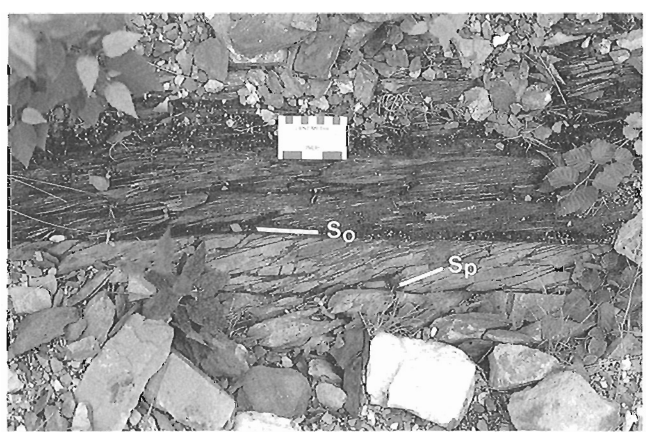

Figure 7. Relation entre le plan de stratification (So) et les plans de la schistosité pénétrative (Sp).

1) Le passage à un milieu réducteur (par exemple, faciès rouge à vert et à noir).

2) Le passage à un milieu plus instable indiqué par une répétition beaucoup plus rapide des faciès sédimentaires.

3) Le passage à une dilution terrigène plus faible (Jébrak et Gauthier, 1991).

4) La présence d'une brèche polygénique grossière dans les environs du gîte de Champagne. Une telle brèche est caractéristique d'un environnement de gîtes SEDEX (Large, 1983). Elle correspond généralement à un faciès de talus formé au pied de failles synsédimentaires, ces failles servant de canaux aux solutions hydrothermales. Dans le cas présent, la nature volcanique de la brèche et son passage latéral à des tufs fins suggèrent une autre origine. Ce dernier point sera clarifié dans la suite de nos travaux.

5) La présence d'activité hydrothermale. L'interval stratigraphique favorable se situera entre les premiers signes d'activité hydrothermale (le niveau sulfuré de l'Etchemin) et ceux de son tarissement (le niveau de mudstone noir à nodules de carbonates riche en barium).

CONCLUSION

Bien que le contexte géodynamique de l'amas sulfuré de Champagne soit inusité pour un SEDEX, c'est à dire qu'il repose dans un bassin d'avant-arc plutôt que dans un bassin épicratonique ou intracratonique, il possède des caractéristiques géologiques propres aux gîtes de type SEDEX, à savoir: 1) un encaissant sédimentaire de nature clastique, 2) une morphologie tabulaire et stratiforme, 3) de fortes teneurs en Pb et en Zn, 4) un horizon à nodules de carbonates riche en barium pour toît et 5) un biseau de brèche au voisinage.

Nos études font ressortir différentes caractéristiques de l'empilement volcano-sédimentaire qui peuvent annoncer la proximité d'un amas sulfuré tel que Champagne: 1) le passage de faciès rouges à des faciès noirs, 2) le passage de sédiments terrigènes grossiers à des sédiments fins et 3) la présence d'hydrothermalisme et de volcanisme.

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Geological Survey of Canada Project 890024

Imbricate thrusting, reverse-oblique shear, and ductile extensional shear in the Acadian Orogen, central Cape Breton Highlands, Nova Scotia[1]

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Lynch, J.V.G. and Tremblay, C., 1992: Imbricate thrusting, reverse-oblique shear, and ductile extensional shear in the Acadian Orogen, central Cape Breton Highlands, Nova Scotia; in *Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 91-100.*

Abstract

Acadian deformation in the central Cape Breton Highlands features: a) mid-crustal west-vergent imbricate thrusting within the Jumping Brook metamorphic suite; b) opposite facing oblique-reverse shear zones, which displace high grade gneissic complexes against medium to low grade volcanic arc assemblages; and c) late ductile extension.

Thrusting is marked by structural repetitions of clastic sequences and by the transition from greenschist to amphibolite metamorphism. Steep reverse faults consist of thick shear zones forming the boundary between supracrustal assemblages and deep-seated gneiss. Extensional faults, developed in the western domain, are typified by moderate- to shallow-dipping mylonite zones and associated schistosity. The mylonites merge with the earlier formed thrust and reverse fault systems that appear to have been re-activated during this later phase of deformation; both north and south transport directions are recorded within the extensional faults.

Résumé

La déformation acadienne survenue sur les hautes terres centrales de l'île du Cap-Breton comprend: des structures en écailles, de vergence ouest, dans la croûte moyenne, à l'intérieur de la série métamorphique de Jumping Brook; b) des zones de cisaillement obliques-inverses de direction opposée, qui déplacent des complexes gneissiques fortement métamorphisés et les amènent au contact d'assemblages d'arcs volcaniques moyennement à faiblement métamorphisés et c) une extension ductile tardive

Le chevauchememnt est marqué par des répétitions structurales des séquences clastiques, et par la transition dans la nature du métamorphisme, soit du faciès des schistes verts au faciès des amphibolites. Des failles inverses de fort pendage se composent d'épaisses zones de cisaillement qui forment la limite entre les assemblages supracrustaux et le gneiss de grande profondeur. Les failles de distension, apparues dans le domaine occidental, sont typiquement représentées par des zones mylonitiques de pendage modéré à faible et par la schistosité associée. Les mylonites fusionnent avec les réseaux plus anciens de failles chevauchantes et de failles inverses qui semblent avoir été réactivés au cours de cette phase ultérieure de déformation; à l'intérieur des failles de distension sont conservés les indices de directions nord et sud de transport.

[1] Contribution to Canada-Nova Scotia Cooperation Agreement on Mineral Development 1990-1992. Project carried by Quebec Geoscience Centre.

INTRODUCTION

In the central and western Cape Breton Highlands, highly contrasting metamorphic domains have been documented (Wiebe, 1972; Currie, 1987), and have been related to Acadian deformation (Plint and Jamieson, 1989). The distribution and age of zones imply the presence of faults or shear zones which accomodated significant vertical translation of units. Telescoping of an approximately 25-30 km thick section of the crust is indicated from barometric determinations in the western Highlands (Currie, 1987; Plint and Jamieson, 1989). Despite these significant advances, the nature and location of faults responsible for the juxtaposition of these highly contrasting crustal levels are not fully established. Thrust faults have been postulated in recent terrain models, in order to explain the distribution of terrane-diagnostic lithologies which occur across terrane boundaries (Raeside and Barr, 1990). This study is a preliminary report which presents mapping results focusing on some of the principal fault systems of the central Cape Breton Highlands. More specifically 1:50 000 scale mapping of the Cheticamp River sheet (NTS 11 K/10) was undertaken (Fig. 1). The work is an extension of the metallogenic investigation of Mengel et al. (1991), and it is hoped that advances in regional geology will improve the metallogenic context, helping to direct explorationists towards new possibilities.

The central Cape Breton Highlands consist of a complex assemblage of gneissic rocks, Silurian volcanic arc sequences, and diverse plutonic suites of Cambrian, Early Ordovician, Silurian, and Devonian ages (Raeside and Barr, 1990; Barr, 1990; Dunning et al., 1990). In the western Highlands, these are unconformably overlain by volcanic and clastic rocks of the Upper Devonian Fisset Brook Formation, as well as Upper Devonian to Carboniferous clastic rocks of the Horton Group. Age determinations, obtained mostly from plutonic and gneissic units, served in large part as a basis for terrane subdivision of the Cape Breton Highlands (Barr and Raeside, 1989; Raeside and Barr, 1990; Dunning et al., 1990). The older Bras d'Or terrane occurs in the eastern and southern Highlands, whereas the younger Aspy terrane dominates to the west and north. The Eastern Highlands Shear Zone (EHSZ) separates the two terranes (Raeside and Barr, 1990).

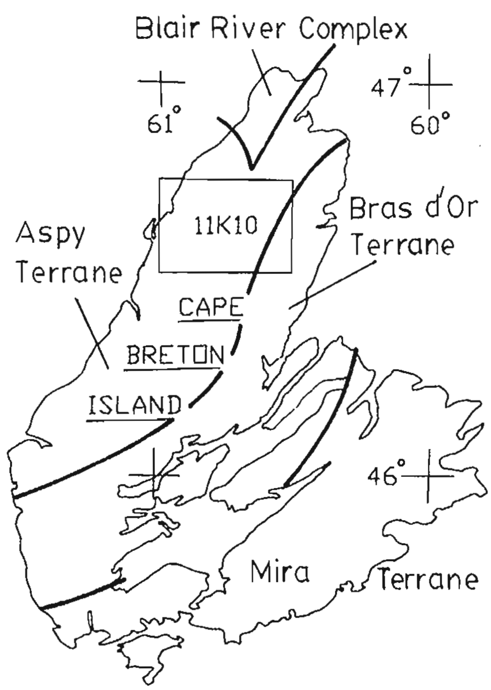

Figure 1. Location of study area, NTS map 11 K/10 Cheticamp River.

Studies with an emphasis on fault structures and shear zones include that of Craw (1984), who documented a system of imbricate west-verging thrusts along the Cheticamp River. To the east, an investigation by Lin (1990) has focused on structural aspects of the EHSZ.

TECTONIC ASSEMBLAGES OF THE CENTRAL CAPE BRETON HIGHLANDS

Several reports and maps have recently been published on the geology which underlies the Cheticamp River area in the central Highlands (Currie, 1987; Barr et al., 1987a; Jamieson et al., 1987; Raeside and Barr, 1990; Jamieson et al., 1990). A brief review of the principal lithotectonic groupings is presented here, as well as some new observations and correlations.

Bras d'Or terrane

The Bras d'Or terrane dominates the eastern portion of the study area (Fig. 2). Characteristic of the terrane are Late Proterozoic metasedimentary rocks, Cambrian calc-alkaline plutons, and Early Ordovician granitic plutons (Raeside and Barr, 1990). The Cambrian plutons include foliated diorite and tonalite which are reported to contain magmatic epidote, indicating deep levels of emplacement (6-10 kbar, Farrow and Barr, 1989). Cambrian granite such as the large Cheticamp pluton in the west of the map, has been assigned to the Bras d'Or terrane but occurs entirely within the defined limits of the Aspy terrane (Fig. 2).

The diorite and tonalite are interpreted to have intruded (Dunning et al., 1990) the McMillan Flowage Formation (Raeside and Barr, 1990) which underlies the southeastern portion of the study area (Fig. 2). The McMillan Flowage Formation consists of an assemblage of high grade metamorphic and supracrustal rocks. Widespread recrystallization, metamorphism, migmatization, and abundant syntectonic pegmatite dykes have obliterated most primary features. Lithologies include abundant quartzite, local marble and calc-silicate, and black amphibolite layers interpreted to be metamorphosed basaltic flows. This assemblage likely reflects a shallow depositional environment near continental crust, possibly a Late Precambrian passive margin or shallow rift. The basement of the McMillan Flowage Formation has not been identified.

Aspy terrane

Jumping Brook Metamorphic suite

An overview of the various rock units included in the Ordovician-Silurian metavolcanic and metasedimentary Jumping Brook metamorphic suite (JBMS) is presented in Jamieson et al. (1987), and Jamieson et al., (1990). These rocks are well exposed in the Cheticamp River area, and include the Jumping Brook complex of Currie (1987). Rocks of the JBMS vary in metamorphic grade from greenschist to amphibolite. Two broad lithological subdivisions or groupings are made: a lower unit which is predominantly volcanic in nature, and an upper clastic unit, although the stratigraphic position of these two units is not rigorously established. The diverse nomenclature for these, and regional correlations are summarized in Table 1 of Jamieson et al. (1990), who attribute a Late Ordovician-Early Silurian age to the JBMS.

The volcanic unit of the JBMS is dominated by metabasite which is pervasively sheared, isoclinally folded, and metamorphosed to greenschist and lower amphibolite grades. Interbedded siltstone and lapilli tuff of intermediate composition, and rhyolite occur locally. Metadiorite sills and intrusions are dispersed throughout the sequence and form the dominant lithology in places. Geochemistry of the volcanic rocks suggest a volcanic arc setting (Jamieson et al., 1990).

Clastic rocks of the JBMS typically include well bedded siltstone, phyllite, or schist, as well as coarse grained sandstone and conglomerate. Sequences are moderately to thinly bedded, and graded bedding is a common primary structure. Quartz-pebble conglomerate is a distinct and widespread lithology; it is poorly sorted but contains well-rounded quartz clasts. Sandstones are typically wackes; arkose is rare; conglomerate containing granite clasts also occurs. The moderately to poorly sorted nature of the clastic units, rapid fluctuations in clast size from bed to bed, as well as the graded bedding are interpreted to indicate deposition as gravity slide deposits, or turbidity flows. The deposits are suggestive of an environment with steep slopes. The style of sedimentation and clast composition are typical of clastic deposits formed in association with tectonically active continental margins, or with volcanic arcs constructed on continental margins (Dickinson and Suczek, 1979). The occurrence of arkose and conglomerate with granite clasts indicate that felsic plutonic rocks such as the nearby Cambrian Cheticamp pluton may have provided some of the detritus; the relative scarcity of feldspar in most sandstones, and the limited winnowing as indicated by poor sorting, indicate however that a quartz-rich source is needed as well. Within the regional context, the likeliest source rocks which fulfill this requirement are the extensive quartzites of the McMillan Flowage Formation.

During the course of the present mapping, the JBMS was extended eastward to a well exposed section along upper Ingonish River, and to rocks along Clyburn Brook (Fig. 2). These were previously assigned to the Fourchu Group by Wiebe (1972), and referred to as the Clyburn Brook volcanic-sedimentary unit by Barr et al. (1987b). The rocks are well preserved, metamorphosed to low or intermediate greenschist grades, and contrast greatly with the surrounding high grade metamorphic rocks. Quartz-pebble conglomerate is a common lithology along the Ingonish River section, and the upper reaches of South Clyburn Brook, and is a useful marker for correlation. The volcanic group along Clyburn Brook is more varied than analogous rocks to the west. It shows a full range of arc volcanics, including basalt-andesite-dacite-rhyolite, and depositional textures indicative of subaqueous volcanism featuring volcanic breccia, tuffs, epiclastic sandstone, and interbedded shale. Correlation of the JBMS to these rocks is presently based on similar lithologies and metamorphic grades; proof through age determinations is still required.

High grade gneisses

High grade gneisses of the Aspy terrane dominate the central highlands, where exposure is poor. The grouping consists of the Pleasant Bay complex (Currie, 1987), the Belle Cote Road orthogneiss (Jamieson et al., 1986), the Cheticamp Lake gneiss, and the Money Point Group (Barr et al., 1987). Age determinations (U-Pb on zircon) indicate that the intrusion of the Belle Cote Road orthogneiss occurred at approximately 433 Ma (Jamieson et al., 1986), and that it is roughly cogenetic with volcanics of the JBMS. The age determination is not pristine however, a significant component of Proterozoic lead was detected within the sample (Jamieson et al., 1987). Available geochronological and petrological data indicate that peak metamorphism in the Pleasant Bay complex approached 8 kbar and 750°C during late Silurian to mid-Devonian time (Plint and Jamieson, 1989).

Fisset Brook Formation

The Fisset Brook Formation is restricted to the western part of the study area. The formation is Late Devonian to Mississippian in age and consists of bimodal basaltic-rhyolitic volcanics and interbedded redbeds (Blanchard et al., 1984). The Fisset Brook Formation lies below coarse clastic and alluvial fan deposits of the Horton Group, and is considered to be cogenetic with granitic subvolcanic intrusive bodies such as the Salmon Pool pluton (Jamieson et al., 1986) which occur in western Cape Breton. Equivalent units to the Fisset Brook Formation are widespread through much of the Maritimes (Blanchard et al., 1984). In the Cheticamp River map area, the Fisset Brook Formation is bounded by steep brittle faults; however they are locally observed to lie unconformably upon the Cheticamp pluton, along a basal granite-boulder conglomerate. Basaltic dykes are abundant in the underlying pluton. The base of the Fisset Brook Formation appears to have been sheared in places along shallow dipping shear zones, separating it from schists of the JBMS below. The depositional environment established by Blanchard et al. (1984) is one of early silicic explosive volcanism and alluvial fan type deposits, followed by basaltic eruptions with associated fluvial and lacustrin sedimentation.

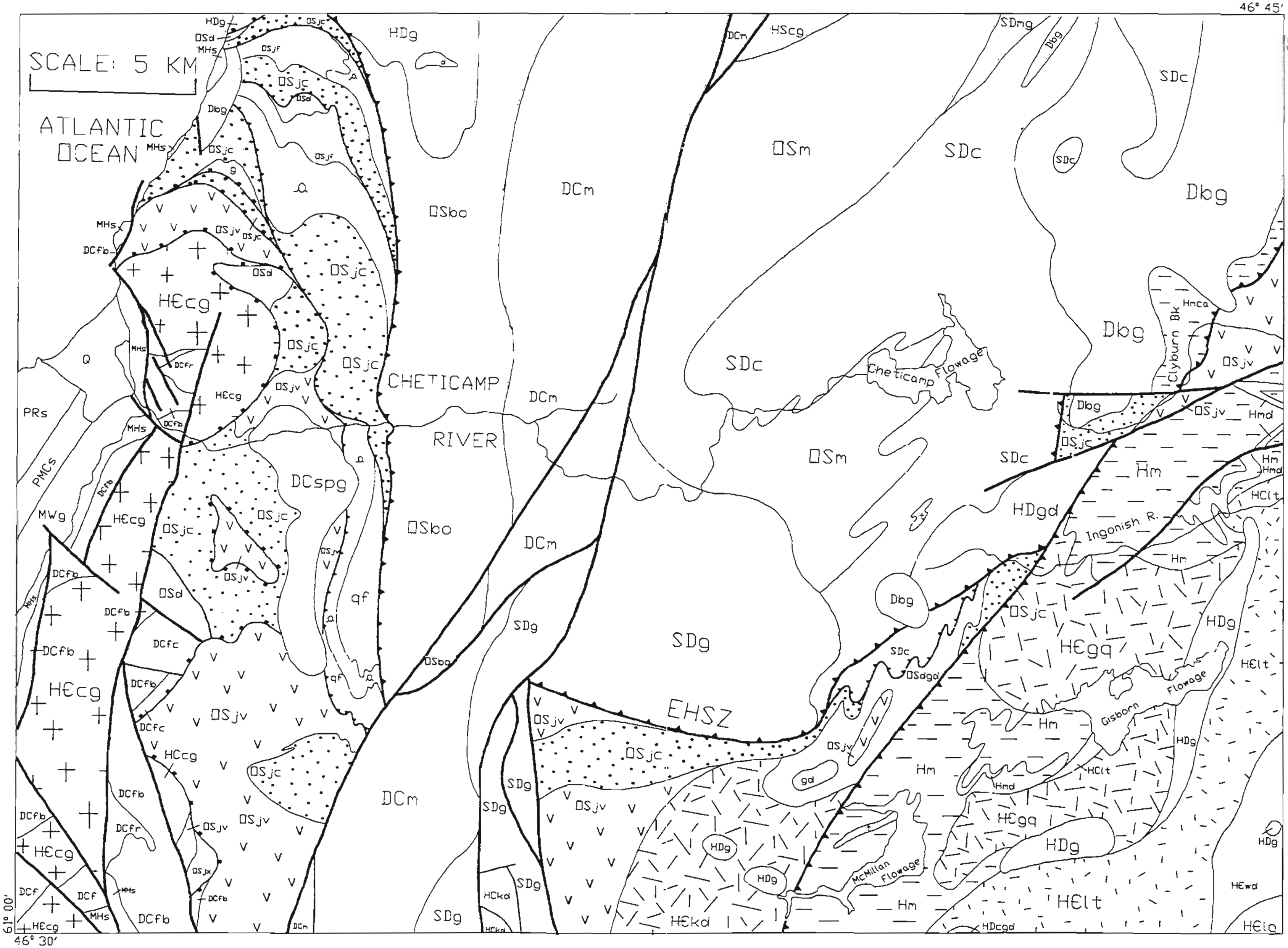

Figure 2. Geology of the Cheticamp River map, produced from 1991 field work and compiled from Barr et al. (1987b), Currie (1987); age dates quoted in legend are from Dunning et al. (1990).

LEGEND

QUATERNARY

Q	Drift, till, sand, gravel.

CARBONIFEROUS

PRs	RIVERSDALE GROUP: grey muscovitic sandstone, rare coaly beds.
PMcs	CANSO GROUP: grey to buff sandstone, minor red shale.
MWg	WINDSOR GROUP: gypsum, anhydrite, limestone, dolostone, minor red shale.
MHs	HORTON GROUP: lithic wacke, arkose, conglomerate, debris flows.

DEVONIAN AND MISSISSIPPIAN

	FISSET BROOK FORMATION
DCfb	Vessicular basalt flows, minor andesite, red siltstone, conglomerate.
DCfr	Rhyolite, quartz-feldspar porphyry.
DCfc	Conglomerate, red sandstone and siltstone.
DCf	Undifferentiated Fisset Brook Formation.
DCm	Margaree granite: coarse grained biotite-granite, rapakivi granite.
DCspg	Salmon Pool granite: pink medium grained equigranular biotite granite.
Dbg	Granite, and biotite muscovite granite.

SILURIAN AND DEVONIAN

SDg	Park Spur granite: foliated and locally schistose or mylonitic granite.
SDmg	Middle Aspy River granitic orthogneiss: foliated gneissic granite
SDc	Cheticamp Lake gneiss (396 Ma): orthogneiss, gneissic granodiorite, minor semi-pelitic paragneiss and biotite schist
SDgd	Garnetiferous gneissic granodiorite.

ORDOVICIAN AND SILURIAN

Aspy terrane

OSm	Money Point Group: hornblende-biotite schists and gneisses, calcsilicate rocks, minor amphibolite.
OSbo	Pleasant Bay complex and Belle Cote Road orthogneiss (433 Ma): quartzo-feldspathic orthogneiss of predominantly granodiorite and granite composition, minor diorite, local amphibolite, dispersed foliated pegmatite.
	JUMPING BROOK METAMORPHIC SUITE
OSjc	Weakly metamorphosed to schistose siltstone, wacke, quartz-pebble conglomerate, some arkose, and arkosic conglomerate, minor rhyolite, feldspar crystal tuff, and chloritic tuff.
OSjv	Chloritic schist, and schistose units of basalt, lapilli tuff, diorite, and metasediments.
OSjx	Chloritic breccia with clasts of chlorite schist.
OSjf	Medium to fine grained quartzo-feldspathic schist, and psammitic schist.
OSd	Diorite, and foliated chloritic diorite.
a	Medium grained black foliated garnetiferous amphibolite.
qf	Medium to fine grained schistose quartzo-feldspathic rock of uncertain origin, and foliated granite.
OSdgd	Interlayered foliated diorite and granodiorite.
HDg	Foliated granite of uncertain age, d meta-diorite, g granite.

CAMBRIAN OR OLDER

Bras d'Or terrane

HCcg	Cheticamp pluton (550 Ma): granite.
HClt	Ingonish River tonalite (555 Ma).
HCkd	Kathy Road diorite (560 Ma): medium to coarse grained gneissic diorite, biotite-amphibole gneiss, dispersed amphibolite and pegmatite.
HCwd	Wreck Cove dioritic suite (560 Ma):foliated medium to coarse grained diorite, tonalite.
HClg	Indian Brook granodiorite (564 Ma).
HCgq	Gisborne Flowage quartz-diorite (564 Ma): foliated medium to coarse grained quartz-diorite, granodiorite, local granite dykes and migmatite.
	MCMILLAN FLOWAGE FORMATION
Hm	High grade interlayered gneissic rocks containing amphibole and garnet: quartzite, black amphibolite, medium to fine grained quartzofeldspathic gneiss, folded pegmatite, and foliated granite.
Hmd	Marble and calc-silicate rock, minor quartzite and amphibolite or metabasalt.
Hmca	Roper Brook Amphibolite: garnetiferous amphibolite, biotite-rich quartzite, foliated granite.

foliation

gneissosity

mylonite

bedding

mineral or stretching lineation.

X outcrop visited during present study.

steep fault, defined, approximate, assumed.

thrust fault, defined, approximate, assumed.

reverse fault (teeth in uplifted block), defined, approximate, assumed.

brittle-ductile moderate to shallow-dipping extensional fault, defined, approximate, assumed.

geological contact, defined,approximate, assumed.

Deposition is thought to have occurred in a horst and graben setting. Total thickness for the Fisset Brook Formation in the study area is approximately 300 m.

ACADIAN DEFORMATION

Three different groups of faults are distinguished which were active at some point in Silurian to Devonian time; an early (a) imbricate thrust system, followed by (b) steep oblique-reverse shear zones, and late (c) low-angle extensional faults. The effects of deformation on mineralization are briefly discussed.

Imbricate thrusting in the western Highlands

Along a section of the Cheticamp River, Craw (1984) mapped imbricated, west-directed thrusting within the Jumping Brook metamorphic suite based on the presence of shear zones, repetition of lithological units, and metamorphic zones. The northern extension of this fault system is now mapped (Fig. 2). The continuation of the thrusts towards the south is not well constrained due to poor exposure, but have been inferred from regional mapping.

The structural imbrication is indicated by repetitions of the Jumping Brook metamorphic suite across west to northwest-dipping thrust faults (Fig. 2). Coarse wacke and quartz-pebble conglomerate served as a marker unit because they maintain their distinctive features at different metamorphic grades. The thrust exposed furthest to the west is highlighted by an abrupt transition from greenschist grade rocks in the footwall to black amphibolite in the hanging wall. The surface trace of this fault also corresponds with the almandine isograd north of the Cheticamp River. Thrust planes are schistose or mylonitic. Down-dip stretching lineations, C-S surfaces, and asymmetric pressure shadows surrounding coarse grained garnet, as well as sigmoidal inclusion trails within garnet, indicate a general westward direction of thrusting. Outcrop-scale folds of bedding in the hanging wall of thrusts are tight to isoclinal, upright to recumbent, and generally plunge moderately north to northeast.

Oblique-reverse faults

Three prominent oblique-reverse faults or shear zones have been mapped. Varying degrees of oblique or dip-slip motion are apparent along the faults, as estimated from the orientations of stretching lineations. However, the distribution of lithologies, changes in metamorphic grade, and textural features suggest a strong reverse component to the shear. Two of the faults occur in the eastern portion of the map area (Fig. 2), and include the Eastern Highlands Shear Zone (EHSZ) (Barr et al., 1987b; Lin, 1990), the other is found in the west. In general the reverse faults juxtapose high grade orthogneiss of the Aspy terrane, or basement rock of the Bras d'Or terrane, against lower grade or younger rocks of the Jumping Brook metamorphic suite and equivalent units. It is not known however if these faults are related. Gneissosity in the region strikes northeast, and is steeply dipping (Fig. 3).

The western reverse fault bounds the JBMS and appears to crosscut the older imbricated thrust system. The fault is well exposed at the headwaters of Corney Brook. Biotite-garnet orthogneiss, locally with kyanite and oligoclase (Plint and Jamieson, 1989), of dioritic to granodioritic composition, occurs east of the fault, whereas amphibolite and foliated quartz-pebble conglomerate occur to the west. The fault zone is approximately 50 m wide, strikes north, dips steeply to the east, and is dominated by coarse grained muscovite-biotite-garnet-staurolite schist interlayered with thin mylonite zones. Stretching lineations plunge steeply east to northeast. Conspicuous large garnet crystals with diameters of 2-3 cm are common. Penetrative C-S structures, asymmetric pressure shadows, rotated mineral trails, sigmoidal foliation traces within porphyroblasts, and down-dip stretching lineations indicate dip-slip motion with east-side up reverse and dextral components. Tight, upright, small-scale folds within mylonite layers plunge moderately to the south, and display an asymmetry suggesting east-side up movement during shear.

In the east, the Eastern Highlands Shear Zone (EHSZ) (Barr et al., 1987b; Lin, 1990) is a thick mylonitic zone (50-500 m wide) which strikes northeast to east, and dips steeply to the north and northwest. The shear zone juxtaposes high grade gneiss and orthogneiss on its northwestern side, against low grade volcanic and clastic rocks to the east which are correlated here with the Jumping Brook metamorphic suite. Detailed structural analysis of the EHSZ by Lin (1990) reports oblique movement during mylonitization with a dextral horizontal component and a northwest side-up vertical component. Field observations of C-S planes, moderately-dipping kink bands, and a generally steeply-plunging stretching lineation are consistent with Lin's (1990) interpretation. The mylonite is overprinted by late folding producing noncylindrical tight upright folds, plunging moderately to the northeast or southwest. The EHSZ forms the southern boundary to the Park Spur granite where thick mylonite strikes east, and is bounded to the west by north striking brittle faults. Large fault-bounded blocks of mylonitized Park Spur granite are distributed to the south, and may represent transported fragments of the EHSZ sinistrally offset to the south along brittle faults.

Another major reverse fault forms the western boundary of the McMillan Flowage Formation and Cambrian Gisborne Flowage quartz diorite (Fig. 2). The fault strikes northeast and is vertical to slightly overturned. Ductile fabrics along the fault crosscut the gneissic and plutonic rocks displaying penetrative C-S surfaces and asymmetric extentional shear bands (Hanmer and Passchier, 1991), which are accompanied by the growth of coarse grained muscovite-biotite-garnet. Asymmetric fabrics and the formation of a steeply-plunging penetrative down-dip stretching lineation indicate a southwest side-up sense of motion. Greenschist grade supracrustal rocks correlated to the JBMS, as well as gneissic rocks of the Kathy Road diorite, occur on the western side of

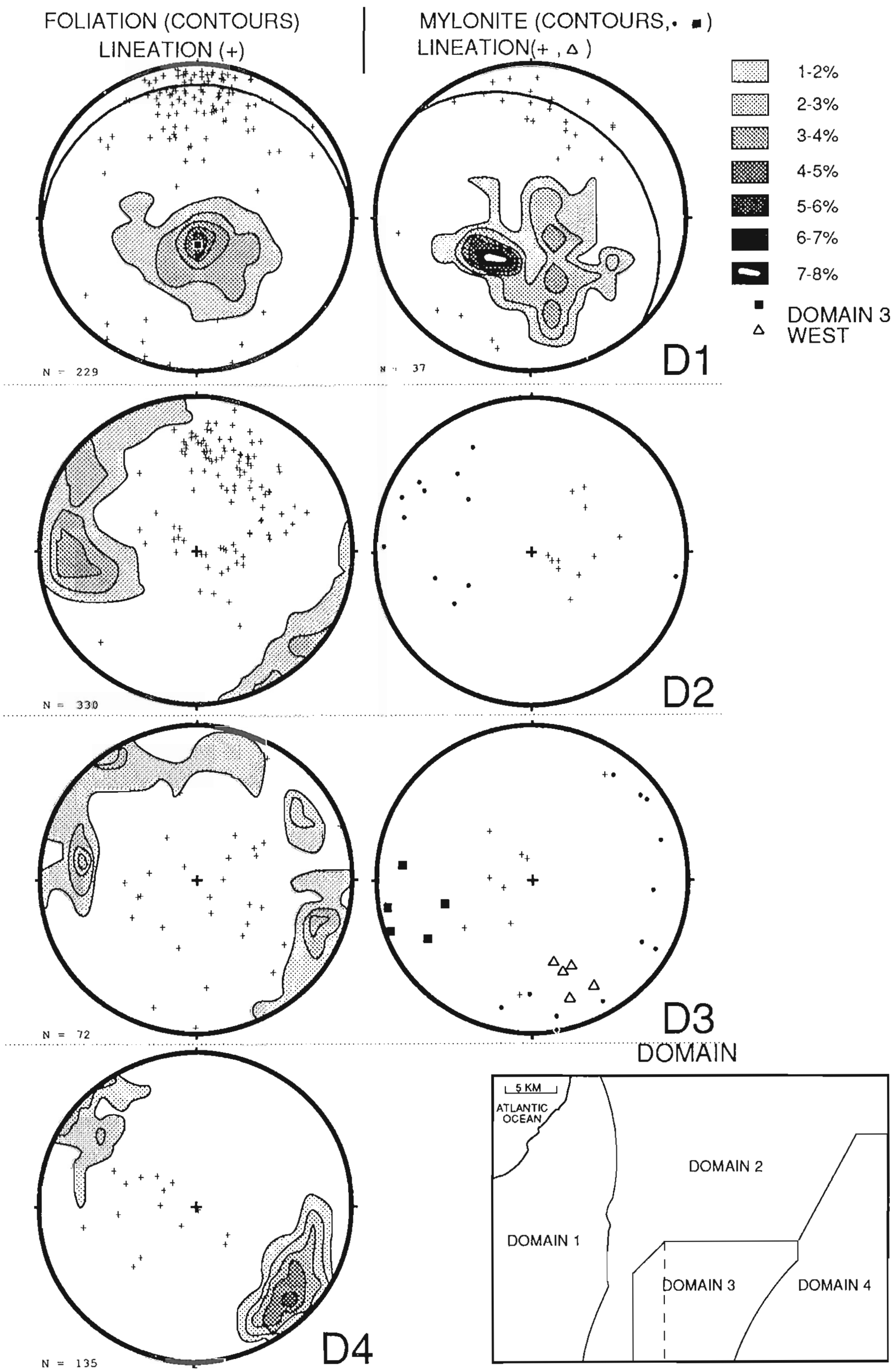

Figure 3. Stereonet plots and contour diagrams of planar and linear structures from different domains.

the fault. Mesothermal quartz-carbonate-pyrite gold veins of the INCO occurrence (Mengel et al., 1991) are hosted by the Kathy Road diorite. The veins consist of shallow-dipping shear veins and flat-lying extension gash veins, and have a geometry which suggests formation during horizontal compression, likely in relation to the nearby reverse fault, in a manner similar to the model proposed by Sibson et al.(1988) for other mesothermal deposits.

Ductile extensional shear

In the west, a series of shallow to moderately-dipping ductile shear zones occur within the Jumping Brook metamorphic suite and Cambrian Cheticamp pluton (Fig. 2, and domain 1 of Fig. 3). Mylonite up to 100 m thick forms the upper and lower boundaries of the Cheticamp pluton. The shear zones dip to the northeast, north, and northwest, and have been affected by open folding. Features indicating formation within an extensional deformation regime include associated, district-wide flat to shallow dipping schistosity (Fig. 3); vertical, quartz-filled tension gashes perpendicular to the north-plunging stretching lineation; and recumbant isoclinal folding. Preliminary analysis of shear-sense indicators within the mylonites, such as asymmetric extensional shear bands (Hanmer and Passchier, 1991), C-S surfaces, rotated feldspar augen, north-south stretching lineations and asymmetric kink bands indicate both movement of upper sheets down towards the north or up towards the south. More work is required for a full assessment of the fault kinematics. The extensional faults merge with and are apparently rooted in, the earlier formed imbricate thrust and reverse fault systems, which were possibly reactivated during the extensional tectonics. The earlier faults are locally overprinted by mylonite displaying shear sense compatible with the extensional phase, or contain a late generation of discreet slip surfaces with quartz-fibre growth that plunges moderately to the north, parallel to the stretching lineation of the extensional faults. Detailed chronological studies are required to work out these relations.

Low angle extensional shear is spatially related to the occurrence of the Fisset Brook Formation. Both are restricted to the western portion of the map area (Fig. 2). In two locations, the base of the Fisset Brook Formation is apparently sheared along shallow-dipping faults where in contact with the JBMS. Also, basaltic dykes of the Fisset Brook Formation which crosscut the Cheticamp pluton are incorporated into, and sheared by, thick mylonite zones which bound the pluton. Further constraint on the timing of extension may be obtained from a single K-Ar date which records an age of 365 Ma for metamorphism of the Jumping Brook complex in the west (Currie, 1987); the dominant stage of mica growth for these rocks associated with the flat to shallow-dipping schistosity, developed during extension.

The distribution of units reflects a complex deformational history during extension. The occurrence of the Cheticamp pluton along extensional mylonite in a position above and below younger rocks of the JBMS requires a history of combined incisement and excisement during extension (Lister and Davis, 1987), or alternatively necessitates a previous phase of deformation to uplift the pluton, possibly as a horst during Silurian arc formation, or during later thrusting. The full extent to which extensional faults overlap thrust faults is not known, but the two phases of deformation are clearly distinguished by their strongly contrasting fabric domains (Fig. 3).

In the JBMS, concordant polymetallic Cu-Zn-Pb lenses (Jamieson et al., 1990; Mengel et al., 1991) have been affected by the extensional shear, resulting in extensive recrystallization and metamorphism. A primary exhalative origin may be speculated for the deposits, but partial or complete remobilization has occurred.

High level felsic plutons such as the Salmon Pool pluton, as well as rhyolite of the Fisset Brook Formation, were intruded into the extending crust. These locally display features of hydrothermal brecciation, and intense stockwork veining. Zoned systems display potassic and phyllic alteration, overprinted by pyritiferous argillic alteration. Although significant discoveries are not reported, alteration characteristics and the tectonic regime are favourable for epithermal precious metal mineralization.

DISCUSSION

A terrane model has been applied to Cape Breton Island which distinguishes the Aspy terrane from the Bras d'Or terrane, (Barr and Raeside, 1989). However, the Aspy terrane is younger than the Bras d'Or terrane, in which case identification of a basement to the Aspy terrane becomes critical to the terrane model. Jamieson et al. (1990) suggested that the Aspy terrane lies on oceanic crust, occupying a position which marks ocean closure, even though rocks with clear oceanic crustal affinities are not known in this region. Alternatively, considering the respective ages of the Bras d'Or and Aspy terranes, and the fact that the Bras d'Or terrane occurs east and west of the Aspy terrane, the possibility that rocks of the Aspy terrane rest unconformably upon those of the Bras d'Or terrane still exists. In the absence of key outcrops, compositional features of clastic units are useful in identifying underlying basement rocks. In particular, quartz-rich turbidites of the Aspy terrane likely reflect a basement of continental crust. Hence, the abundant quartzite of the Bras d'Or terrane is an accountable source. Furthermore, geochemical evidence for a continental basement is provided by the inherited Proterozoic lead identified in the Belle Cote Road orthogneiss of the Aspy terrane (Jamieson et al., 1986), disallowing a primitive oceanic origin. The position of the Aspy terrane may mark a magmatic front above an older basement. More research, particularly in sedimentology and stratigraphy, is required to resolve this problem. Within the existing data however, the authors feel that the unconformity model (Fig. 4) must be given further consideration.

Compilations and maps of regional geology (Barr et al., 1987a, b) show that rocks of Ordovician age are with the exception of some Early Ordovician granite bodies in the Bras d'Or terrane (Dunning et al., 1990) almost entirely absent from the Cape Breton Highlands. This gap in the geological record would correspond to the above proposed unconformity, which would be regional in extent and span a considerable amount of time. Such an unconformity may reflect regional uplift, and within the context of Appalachian tectonics, may be a manifestation in Cape Breton of the Taconic event. Establishing this unconformity is critical if broader correlations of the Bras d'Or terrane are to be attempted.

Imbricate thrusting and reverse-oblique faulting typify compressional deformation of the Acadian orogen in the central Highlands. Opposite facing reverse faults which dip towards each other and which have transported high grade gneiss, is a geometry which may be interpreted to indicate the presence of a large scale thrust ramp in the subsurface; foreland-directed reverse faults developing at the upper hinge of the ramp, back-thrusting or reverse faults developing at the lower hinge. Significant contrasts in metamorphic regimes across faults, from approximately 8 kbar to 5 kbar or less (Plint and Jamieson, 1989) indicate that the ramp may have a considerable relief, on the order of 10 km, to have been able to accomodate the uplift. Thin-skinned imbricated thrusting, such as in the JBMS, may have developed along the upper ramp flat within the foreland portion of the system.

Extension of the crust and the development of shallow-dipping ductile shear zones occurred after the compressional phase, contributing to the removal of overthrust sheets and uplift of the footwall. The late Devonian Fisset Brook Formation was deposited upon the exhumed footwall of the thrust system, which appears to have been exposed during extension. Deep burial and peak metamorphism before extension is dated at approximately 380-420 Ma (Plint and Jamieson, 1989). Extensional collapse is interpreted to have ocurred as a consequence of an unstable overthickened crust developed during compression. The direction of extension along the orogen in a north-south direction, oblique to generally east-west compression, may indicate that lateral confinement was limited, resulting in orogen-parallel tectonic escape. It is noteworthy for comparison that analoguous late Devonian basin-forming processes, where shallow-dipping extensional shear zones have re-activated older thrusts, are reported from western

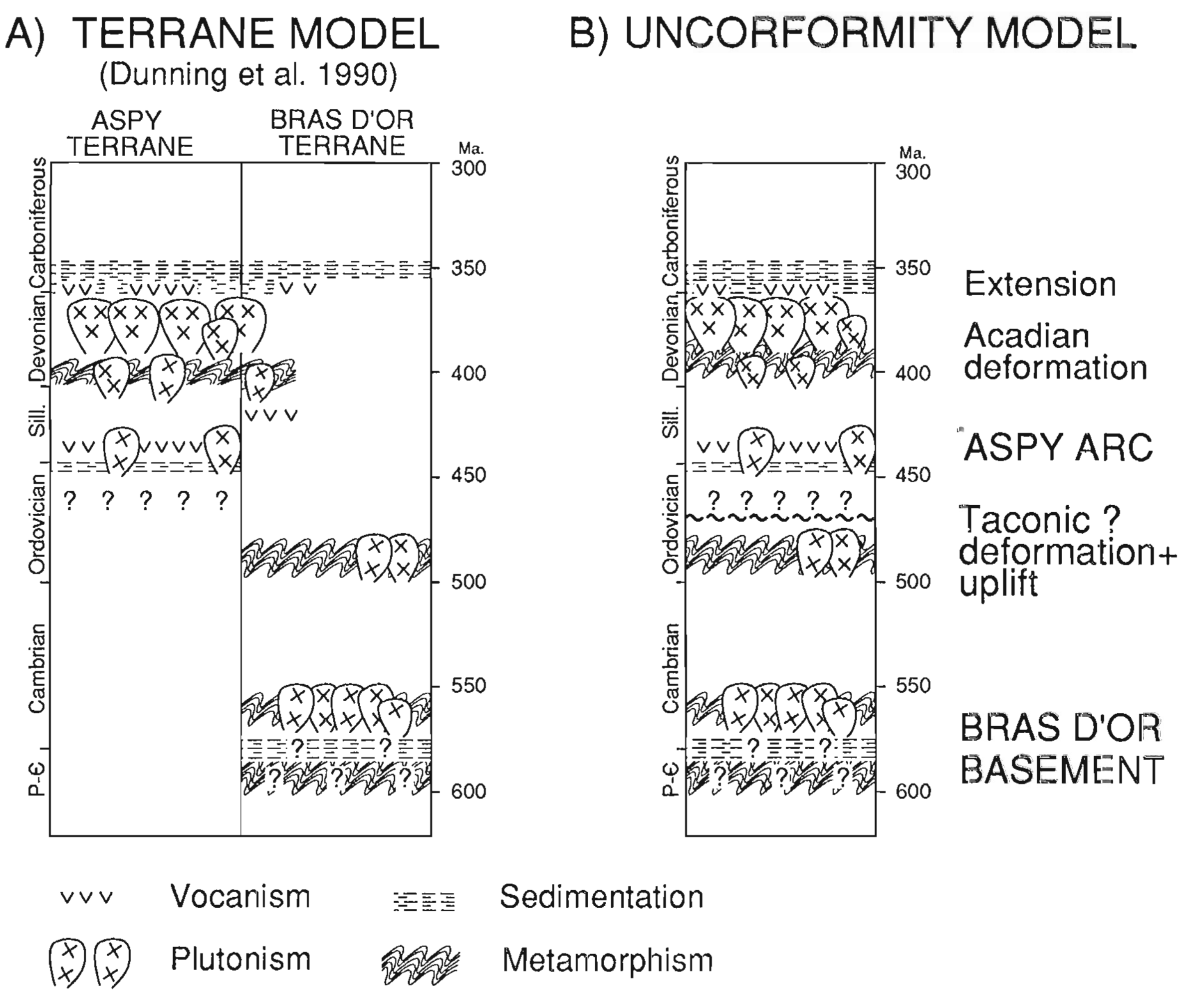

Figure 4. Diagram compares terrane model of Dunning et al. (1990) in (**A**) with alternative interpretation (**B**) which proposes that volcanic arc rocks of the Aspy terrane were deposited unconformably upon a basement consisting of the Bras d'Or terrane.

Europe and the north Atlantic region (McClay et al., 1986; Seranne and Seguret, 1987), and have been compared with the well documented Tertiary extensional tectonics of the western Cordillera (Lister and Davis, 1987).

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Structural study of relationships between gold occurrences and the Rocky Brook-Millstream Fault zone in the Upsalquitch Forks area, northern New Brunswick[1]

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Abstract

Gold occurrences of the Upsalquitch Forks area in northern New Brunswick are located along subsidiary faults in the Rocky Brook - Millstream Fault zone (RBMF). Detailed field studies of the Simpsons Field and McCormack-Jonpol gold occurrences reveal a complex pattern of isoclinal folds and anastomosing faults with structures indicating brittle-ductile deformation. The analysis of striated fault planes indicates dextral shear during deformation in the RBMF. Gold appears to be associated with brittle-ductile structures related to the dextral faulting, and is found in sedimentary rocks and/or in mafic to felsic intrusions of Late Silurian to Early Devonian age.

Résumé

Dans le nord du Nouveau-Brunswick, les indices aurifères de la région d'Upsalquitch Forks sont localisés le long de structures tributaires de la zone de faille de Rocky Brook-Millstream (RBMF). L'étude détaillée des indices de Simpsons Field et de McCormack-Jonpol a révélé la présence d'un patron structural complexe caractérisé par des plis isoclinaux et des failles fragiles-ductiles anastomosées. L'analyse de plans de failles striés suggère que le régime de contraintes associées à la déformation régionale a subi une rotation dextre. L'or est concentré au sein de structures fragiles-ductiles associées à un décrochement dextre le long de la RBMF. La minéralisation aurifère se trouve au sein de roches sédimentaires et/ou de roches intrusives de composition felsique à mafique d'âge Silurien tardif à Dévonien précoce.

[1] Contribution to the Canada-New Brunswick Cooperation Agreement on Mineral Development, 1990-95. Project carried by the Quebec Geoscience Centre.
[2] INRS-Géoressources

INTRODUCTION

Since 1985, several gold occurrences have been discovered along the Rocky Brook-Millstream Fault (RBMF) and related faults in the Appalachians of northern New Brunswick. A recent study has shown that most of these gold occurrences are either related to volcanogenic base-metal deposits or to different types of epigenetic mineralization (Ruitenberg et al., 1990). In the latter group, several gold occurrences of the Bathurst area were shown to be structurally controlled and spatially associated with the Elmtree and Rocky Brook-Millstream faults (Tremblay and Dubé, 1991).

This paper presents the result of field work during the summer of 1991 on gold showings of the "epigenetic" type in the Upsalquitch Forks area (map 21O/10) in northern New Brunswick (Fig. 1). This study represents the second year of an MDA-II project initiated in 1990 (Tremblay and Dubé, 1991) and is part of a large-scale study of gold deposits in the Canadian Appalachians (Dubé, 1990). Field investigations along the RBMF in the Upsalquitch Forks area were conducted to determine the structural setting and the genetic character of the gold occurrences. Seven gold showings were visited (Fig. 1). The best exposed sites are the Simpsons Field showing and the McCormack-Jonpol prospect. They are used herein as case studies for understanding the lithological and structural context of gold mineralization in the area. Other gold showings (Dalhousie Road, Mulligan Gulch, and Upsalquitch Forks) were documented by mapping existing trenches and logging drill cores. Angelier's reconstruction method (Angelier, 1984) was applied in order to determine the orientation of paleostress axes in the area and to define their relationships to gold-hosting structures.

REGIONAL GEOLOGY

In north-central New Brunswick (Fig. 1), four tectonostratigraphic units are recognized: the Elmtree and Miramichi terranes, and the Matapedia B and C cover sequences (Ruitenberg et al., 1990). The Elmtree and Miramichi terranes are made up of Middle to Late Ordovician volcanic and sedimentary rocks occurring as fault-bounded allochthonous packages. Late Ordovician to Early Devonian cover rocks of the Matapedia B and C series obscure the Ordovician terrane boundaries. The RBMF crosscuts the area along an east-to-west trend. It is interpreted as a dextral transcurrent fault zone related to a transpressional deformation regime (van Staal et al., 1990; van Staal and Fyffe, 1991).

According to Fyffe and Fricker (1987), the Miramichi Terrane is found south of the RBMF and consists of polydeformed epiclastic and volcanic rocks of the Tetagouche Group (van Staal, 1987). Uranium-lead analysis of zircon from the Tetagouche Group yielded a Late/Middle Ordovician Age (465 Ma, van Staal and Fyffe, 1991). The Elmtree Terrane occurs north and west of the RBMF. It is

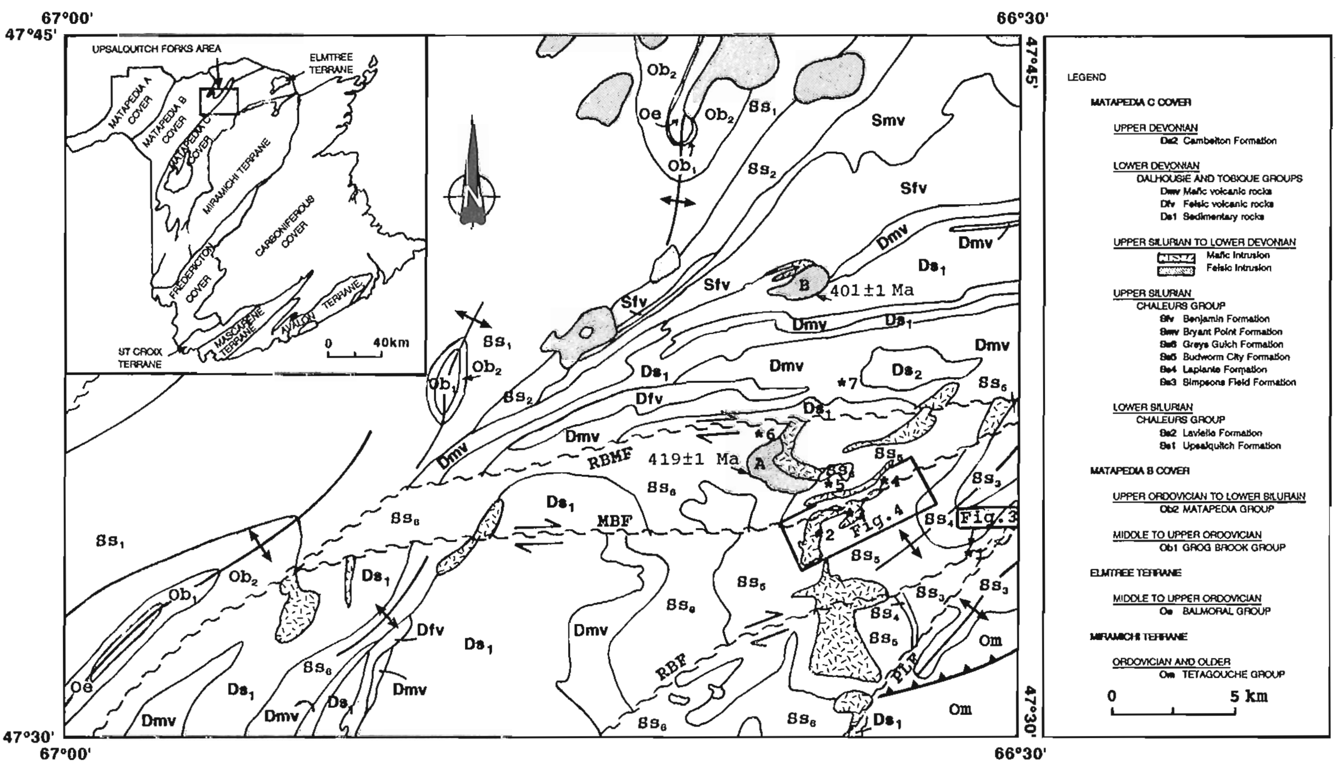

Figure 1. Regional geology of the Upsalquitch Forks area. After Walker and McCutcheon (1990). A: Mulligan Gulch Intrusion; B: Jerry Ferguson Porphyry. Gold occurrences are located by numbered asterix (*); 1- Simpsons Field; 2- Jonpol; 3- McCormack; 4- McCormack-East; 5- Dalhousie Road; 6- Mulligan Gulch; 7- Upsalquitch Forks. **PLF**-Portage Lake Fault, **RBF**-Ramsay Brook Fault, **MBF**-McCormack Brook Fault, **RBMF**-Rocky Brook Millstream Fault.

mainly made up of the Balmoral Group that consists of picritic basalt and andesite overlain by graptolitic slates and cherts of Caradocian age (McCutcheon and Bevier, 1990).

The Matapedia B Cover overlies the Elmtree Group and is made up of turbidites of the Grog Brook Group, and limestones and calcareous shales of the Matapedia Group. The Grog Brook is overlain by the Matapedia Group; their age ranges from Late Ordovician to Early Silurian (McCutcheon and Bevier, 1990).

The Matapedia C Cover occupies a large part of the Upsalquitch Forks area and hosts the gold occurrences. North of the RBMF, it consists of the Silurian Chaleur Group and the Lower Devonian Dalhousie Group. The Chaleur Group is made up of clastic rocks and limestones of the Upsalquitch and La Vieille formations that are overlain by mafic and felsic volcanic rocks of the Bryant Point and Benjamin River formations. The Dalhousie Group is composed of siltstones and sandstones with interbedded mafic volcanic rocks. South of the RBMF, the Matapedia C Cover comprises Upper Silurian Chaleur Group and Lower Devonian Tobique Group. The sedimentary rocks of the Chaleur Group consist of the Simpsons Field, Laplante, Budworm City, and Greys Gulch formations in ascending stratigraphic order (McCutcheon and Bevier, 1990). The Lower Devonian Tobique Group, which is lithologically similar to the Dalhousie Group, conformably overlies the Greys Gulch Formation.

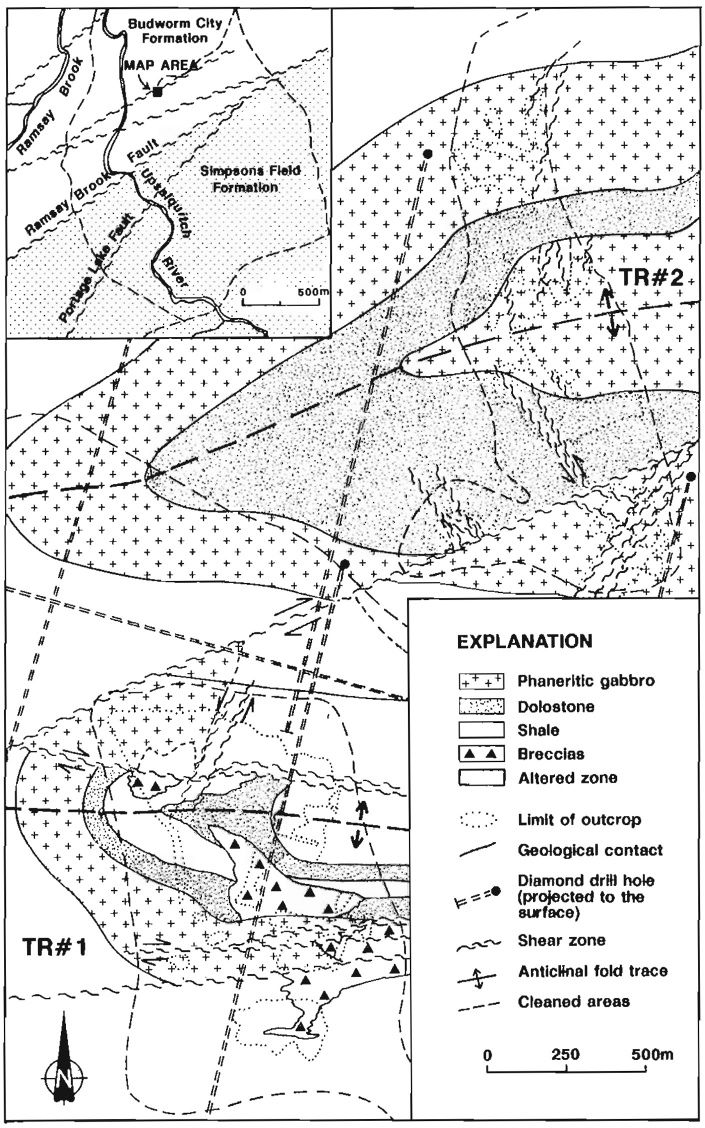

Figure 2. Location and geological map of the Noranda trenches #1 and #2 at Simpsons Field showing. Modified from Fitzpatrick (1986).

Mafic to felsic dykes and plutons of Upper Silurian to Lower Devonian age occur on both sides of the RBMF. South of the fault, mafic intrusions are plagioclase-bearing porphyritic gabbros and diorites. They were interpreted to be comagmatic with lavas of the Dalhousie and Tobique groups (Murphy, 1989). The Mulligan Gulch intrusion (Fig. 1) was dated 419 ± 1 Ma (U-Pb zircon age, McCutcheon and Bevier, 1990). Felsic intrusive rocks are found mainly north of the RBMF and consist of rhyodacite to high-silica rhyolite (Brewer et al., 1991). The Jerry Ferguson porphyry was intruded at the contact between the Chaleurs and Dalhousie groups and yielded a U-Pb zircon age of 401 ± 1 Ma (McCutcheon and Bevier, 1990). Most gold occurrences are hosted by the mafic intrusions or are situated in sedimentary host rocks of the Budworm City and Greys Gulch formations near the contacts of the intrusions.

GOLD IN THE UPSALQUITCH FORKS AREA

The Simpsons Field showing is considered to be a typical gold occurrence of the area. The Jonpol-McCormack prospect, comprising the Jonpol, McCormack Brook, and McCormack Brook East showings, is used to show large macroscopic-scale deformation features that occur in the area. Other gold occurrences (Dalhousie Road, Mulligan Gulch, and Upsalquitch Forks) will be discussed briefly.

Simpsons Field

Since 1985, Noranda Exploration carried out a program of geochemistry, trenching, and diamond drilling on this property (Fitzpatrick, 1986a). The best assays graded up to 0.234 oz/ton (8g/t) Au over one metre.

The Simpsons Field showing is hosted by the Budworm City Formation and is located near the junction of the Ramsay Brook (RBF) and the Portage Lake (PLF) faults. Good exposure is found in the Noranda trenches #1 and #2 (Fitzpatrick, 1986a; Fig. 2). A detailed map of trench #1 is shown in Figure 3. The least deformed unit is a green, plagioclase phyric (<1 mm) gabbro dyke with phaneritic groundmass. Along the contact with host rocks, the gabbro is locally aphanitic. The Budworm City Formation consists of dolostones, black shales, and breccias. Although the dolostones are strongly silicified and carbonatized along contacts with the gabbro, alternating dark- and light-colored beds remain visible and confirm the sedimentary origin of these rocks. Siltstones and shales are present to the south, and in the center, of trench #1 (Fig. 3). Millimetre beds of sandstone are boudinaged along the foliation plane. Contacts between shales and dolostones are gradational. Another rock type occurs locally and consists of breccias made up of hydrothermally-altered sedimentary clasts supported by a dark-colored, fine grained matrix of sedimentary origin.

Structure and ore morphology

Local structures seem to be related to competency contrasts among the various rock types and units. The contact between the gabbro and host rocks is faulted. In trench #1 (Fig. 3), a penetrative east-west schistosity is developed within the host rock and is associated with a tectonic stretching of brecciated

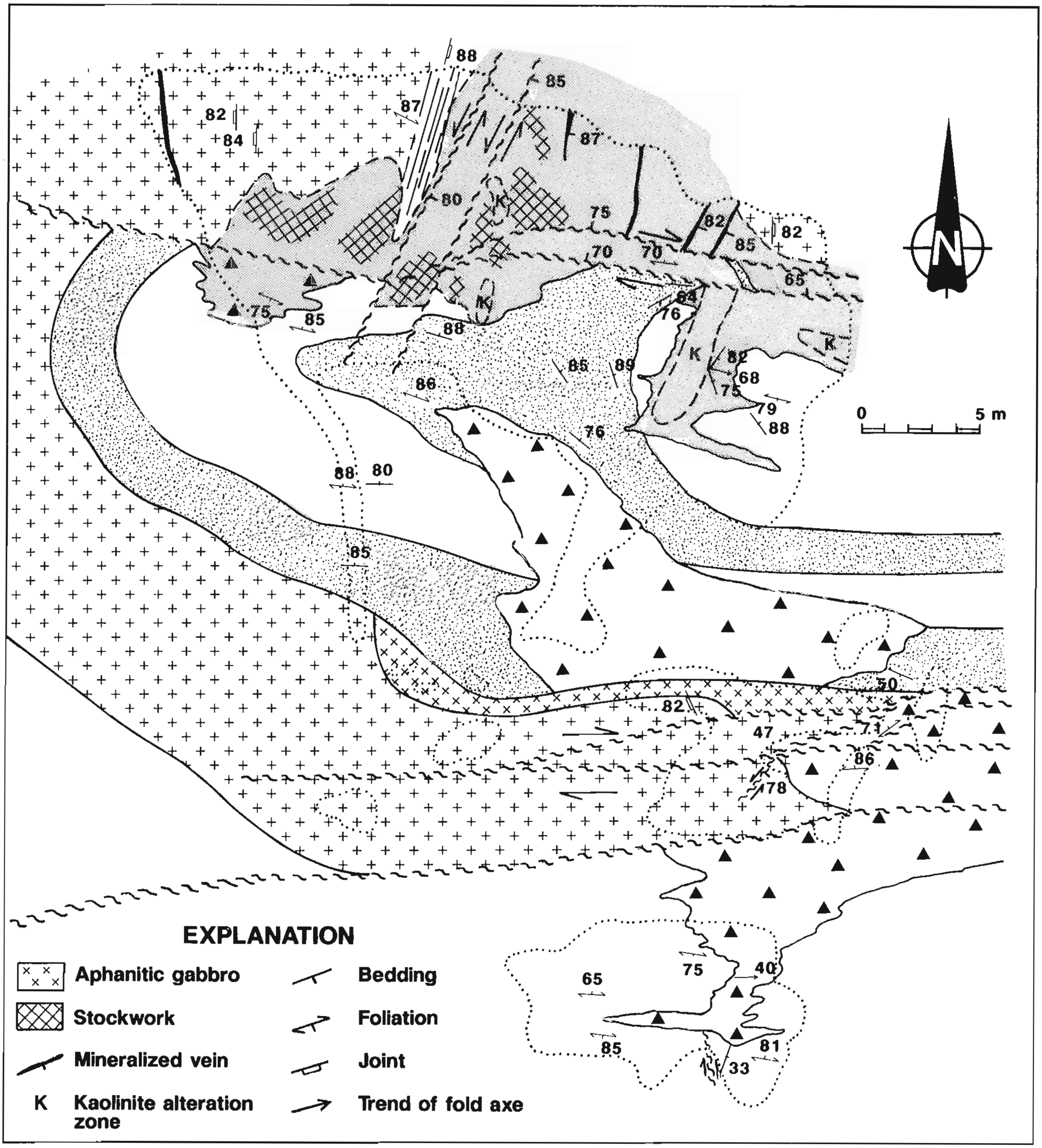

Figure 3. Detailed geological map of trench #1 of the Simpsons Field showing. Same symbols as Figure 2.

rocks. Isoclinal, east-plunging folds, and brittle-ductile faults are found in tr#1 and tr#2 (Fig. 2, 3). In the gabbro, foliation-parallel, anastomosing shear zones are associated with boudinage. Along the northern limb of the fold in trench #1, dolostones are truncated by a metre-wide fault zone. The regional foliation is best developed along the contact between sheared sedimentary rocks and altered gabbro. Antithetic sinistral faults locally displace this zone of intense deformation.

Brittle deformation fabrics are mainly found in the altered part of the gabbro. Subvertical, extensional veins of siderite-quartz-pyrite-hematite trend approximately 360° to 035°, perpendicular to the main shear zone. A stockwork of veinlets made up of pyrite, siderite, quartz, and hematite is found in both the gabbro and sedimentary rocks.

There are two types of alteration found at Simpsons Field. The first type is attributed to the intrusion because it appears as a silicification associated with disseminated pyrite and carbonate minerals in rocks adjacent to the gabbro. The second alteration type is attributed to the regional deformation because it is mostly found near fault zones and associated veinlet stockworks. It consists of patchy zones of carbonatization and silicification with minor hematitic and argillic material.

Gold is found within the veinlet stockworks and the extensional veins and is therefore associated with the second type of alteration. Gold-bearing veins are made up of quartz-siderite and minor pyrite, arsenopyrite, and hematite (Ruitenberg et al. 1990). Gold occurs as isolated grains within pyrite or along grain boundaries (Clemson, 1986).

McCormack-Jonpol prospect

The McCormack-Jonpol prospect is located on both sides of the McCormack Brook Fault (MBF), an ENE-WSW trending splay of the RBMF. The property was investigated for gold by Noranda Exploration from 1985 to 1988.

The area is underlain by the Budworm City Formation, a unit that is intruded by small plutons and sills of gabbro (Fig. 4). Monolithic to polymictic breccias surround the intrusions. The monolithic breccias are made up of subangular to subrounded clasts of finely laminated siltstone. The clasts are supported by a shaly matrix. The polymictic breccias consist of subrounded rhyolitic, carbonaceous, and epiclastic rock fragments in a carbonate-albite-quartz matrix (Murphy, 1989). The presence of intraformational deformation and the lack of foliation in the clasts suggest that most of these rocks were partly consolidated during the brecciation process.

An interpretative geological map of the area is presented in Figure 4. It shows a structural style characterized by isoclinal folds and brittle-ductile fault zones. Dragged, and refolded folds are wedged along anastomosing faults. The map pattern shows that both the gabbro and breccias were folded concordantly with their host rocks.

The structural pattern shown in Figure 4 is also found at the outcrop scale. Figure 5 shows a fold-interference pattern that is defined by alternating beds of dark shale and grey sandstone. In this outcrop, only one axial-planar foliation is present. These superimposed folds are believed to be tectonic rather than slump folds because the fold hinges are parallel to regional fold trends.

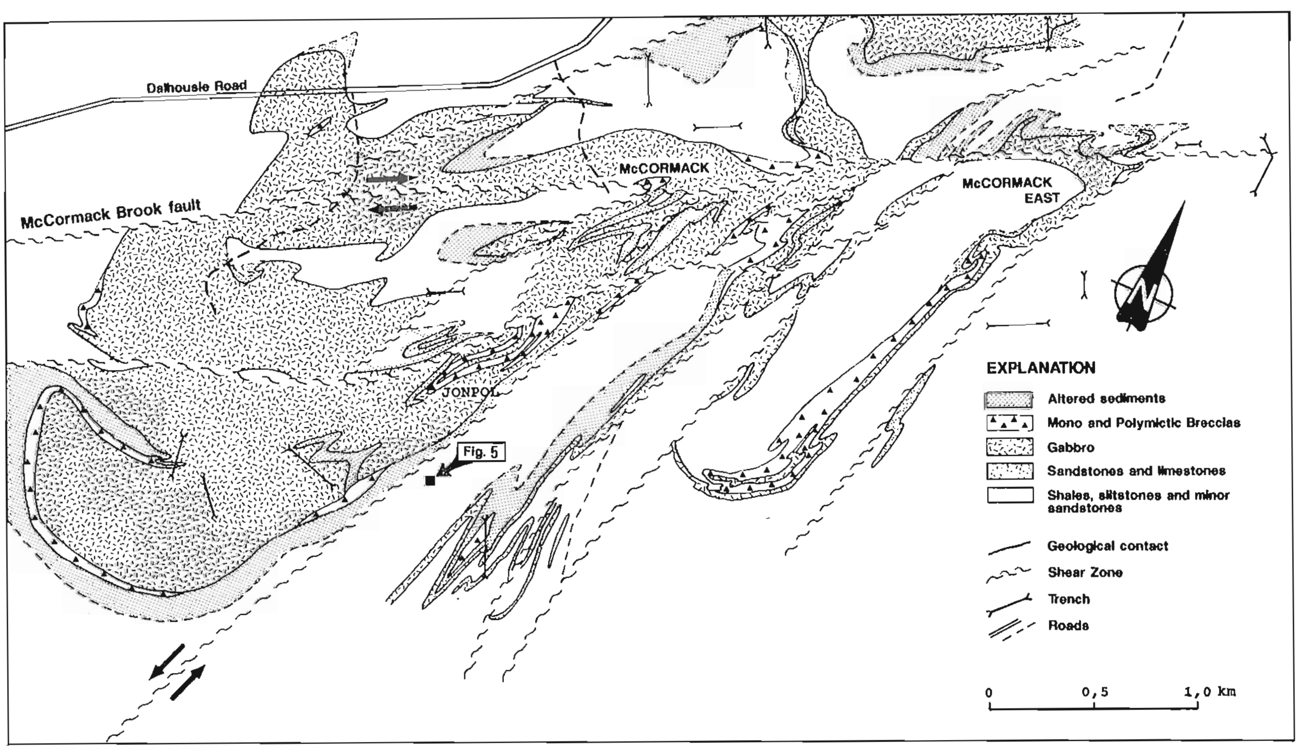

Figure 4. Interpretative geological map of the Jonpol-McCormack prospect. Modified from Adair (1988) and Mitton (1988).

At McCormack-Jonpol, two sets of subvertical faults are associated with folds. Set one is trending 025°-205° and is parallel to the regional foliation. The MBF represents the second fault set, trending approximately 060°-240°. These two sets form an anastomosing pattern along the trace of the MBF although in detail, dextral faults of set two commonly offset sinistral shears of set one. The regional foliation is reoriented into subparallelism with the shear zone walls near the MBF. Near major faults, sedimentary host rocks are most deformed and bear a penetrative schistosity, whereas intrusive rocks show a brittle-ductile rheological behaviour. In drill cores, ductile fabrics seem to be dominant along shear zones. Contacts between intrusive rocks and sedimentary rocks are tectonized, strongly altered, and show well-developed foliation-parallel shear veins. In magmatic rocks, the deformation evolves gradually into a more brittle behaviour where extensional veins predominate.

Mineralization and alteration

Mineralized zones are found within carbonatized and silicified gabbros and sedimentary rocks along shear zones. In gabbroic rocks, mafic minerals are altered to chlorite and epidote. Hematite is found as stringers and patches in bleached zones (Adair, 1988). Orange to brownish limonitic staining was also observed along contacts between the gabbro and sedimentary rocks. Murphy (1989) recognized two stages of alteration. The earliest stage is attributed to quartz and albite replacement of the original gabbroic texture. The later alteration stage is characterized by quartz, siderite, ankerite, and fuchsite (Adair, 1988; Murphy, 1989). It is related to the regional deformation and metamorphism, and is associated with gold mineralization.

Gold is associated with disseminated sulphides in intrusive rocks; it is spatially associated with brittle-ductile shears and related alteration zones. Sulphides include pyrite, arsenopyrite, and chalcopyrite. Gold occurs in crystal lattices of pyrite, and at boundaries between pyrite grains (Fitzpatrick, 1986b; Murphy, 1989). In drill cores, best assays (0.041 oz/ton (1.37 g/t) over 1.34 m) were found in quartz-carbonate shear veins and stockworks associated with fault zones. Breccias locally contain minor pyrite and arsenopyrite (5-15%), which contain low-grade gold (30 to 500 ppb; Mitton, 1988).

Other gold occurrences

North of the Simpsons Field and McCormack-Jonpol prospects, there are three gold showings that are hosted by felsic to intermediate intrusive rocks. These are the Dalhousie Road, Mulligan Gulch, and Upsalquitch Forks gold showings (Fig. 1).

The Dalhousie Road showing was the first gold occurrence to be found in the Upsalquitch area during the 1980s (Ruitenberg et al. 1989). It is located 1 km north of the McCormack-Jonpol prospect (Fig. 1). It consists of an east-trending, subvertical dioritic intrusion hosted by calcareous siltstones. Gold is associated with quartz-carbonate-pyrite-arsenopyrite veins that occur along brittle-ductile shear zones within the diorite.

The Mulligan Gulch showing is 500 m south of the RBMF (Fig. 1). The mineralization occurs in a porphyrytic monzonite pluton intruding Late Silurian carbonaceous siltstones (Murphy, 1989). The alteration is characterized by quartz, ankerite, sericite, and epidote. In drill core, gold occurs in altered, fine- to medium-grained porphyry close to a sheared contact with host rocks, or in pyrite-sphalerite-galena-chalcopyrite veins.

The (Au-Ag-Cu)-bearing Upsalquitch Forks showing (Fig. 1) is the only gold occurrence located north of the RBMF (1.7 km). It is hosted by volcanic rocks of the Dalhousie Group. Gold occurs in an altered dioritic intrusion in fault contact with unaltered basalts. Quartz, calcite, sericite, and hematite are the dominant alteration products. Veins are also present and consist of quartz and calcite with disseminated pyrite, chalcopyrite, galena, pyrrhotite, and sphalerite. Gold-rich zones are concentrated along shears within the diorite whereas silver-rich zones are associated with quartz-sulphide veins close to the unaltered diorite (Hoy, 1987).

Brittle-to-ductile deformational style is a common characteristic of these showings. At Dalhousie Road, anastomosing shear zones crosscut both the intrusion and the calcareous host rocks. The pattern formed by shear zones suggests that the intrusion has been boudinaged and rotated in a dextral sense. At Mulligan Gulch, two sets of quartz-carbonate veins transect the altered zone. One set consists of 50-150 cm wide, foliation-parallel shear veins trending N230° and dipping 70° to 85°NW. The other set consists of extensional veins (<10 cm) trending N110° to N140 and dipping 55° to 75°SW. At Upsalquitch Forks, contacts between the diorite and basalts are sheared and fractured. Alteration is predominant in the diorite and is associated with a stockwork of 5 to 50 cm wide veinlets. Most veinlets trend east, parallel to the shear zone. A secondary set of extensional, subvertical veins trends N150°. The geometry of sheared and extensional veins observed in these two sites suggests a maximum stress vector ($\sigma 1$) oriented to the southeast, which is compatible with a dextral movement along the RBMF and related faults (see below).

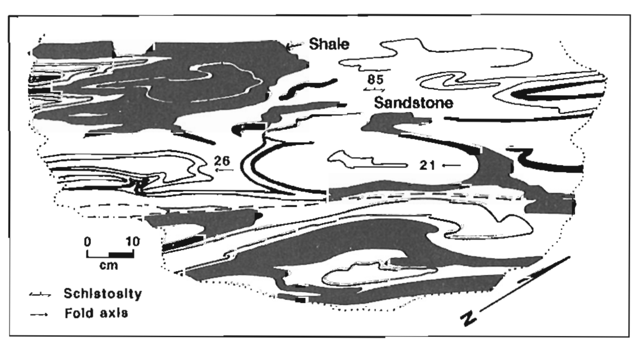

Figure 5. Hand sketch of an outcrop showing refolded folds. Located south of the Jonpol showing (see Figure 4).

PALEOSTRESS ANALYSIS OF THE UPSALQUITCH FORKS AREA

In northern New Brunswick, Acadian regional deformation closely followed the sedimentation and magmatic activity in Late Silurian to Early Devonian time (McCutcheon and Bevier, 1990). Faulting appears to have been almost synchronous with folding in a continuous process of crustal shortening. East-trending, dextral strike-slip faulting accommodated late tectonic movements and reoriented pre-existing structures. This interpretation is supported by a paleostress reconstruction analysis (cf. Angelier, 1984) of the area. For this analysis, striated fault planes were measured in competent rocks around the Simpsons Field and McCormack-Jonpol auriferous sites in order to determine the orientation of stress axes ($\sigma 1$, $\sigma 2$, $\sigma 3$). At McCormack-Jonpol, three maximum stress directions ($\sigma 1$) were calculated (Fig. 6A); namely 112/09°, 324/08° and 350/29°. Field evidence (see above) suggests a dextral sense of shear along the RBMF and the MBF. The $\sigma 1$ vector oriented at 112° is attributed to an early brittle deformation. During continuous rotational deformation, $\sigma 1$ migrated through 324/08° to 350/29°. The fold-related deformation probably occurred during the last two compressive components (324° and 350°). This interpretation is supported by similar results obtained for $\sigma 1$ calculated from striated planes and $\sigma 1$ extrapolated from the maximum concentration of the regional schistosity (Fig. 6B). The rotation of $\sigma 1$ is compatible with dextral movement along northeast-southwest and east-west faults and shear zones.

The analysis of striated fault planes from the Simpsons Field area gave two compressive tensors: 1) $\sigma 1$=353/13°; $\sigma 3$=128/71°, and 2) $\sigma 1$=209/10°; $\sigma 3$=344/76° (Fig. 6C). Brittle

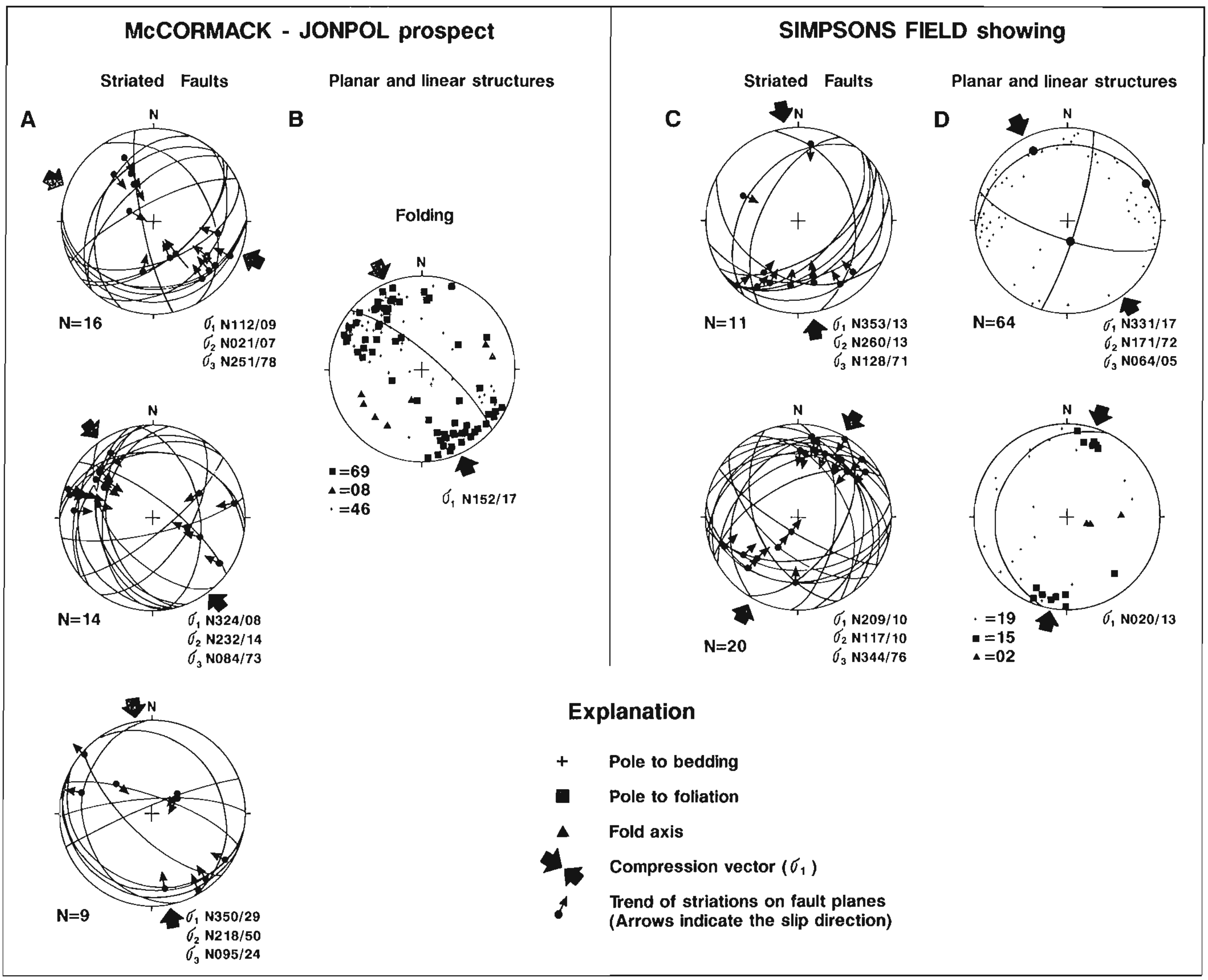

Figure 6. Equal area, lower hemisphere stereographic projections for structures described in this paper. **A**) and **C**) refer to paleostress axes calculated by the Angelier's method from computation of fault planes and associated striations; **B**) and **D**) refer to stress axes deduced from planar and linear structures compiled from each site. See text for discussion.

Table 1. Characteristics of gold occurrences of the Upsalquitch Forks area

Occurrences	Commodities	Metallic minerals	Alteration minerals	Host rocks	Structure	References
Simpsons Field	Au, As	Au, Asp, Py	Qtz, Sid, Kao, Chl, Ser, Hem	Gabbro and sediments	Stockwork along dextral, fold-related-faults, east plunging axes, along the RBF	3, 4, 10, 11
McCormack Brook	Au, As, Cu	Asp, Py, Cp, Au	Qtz, Ank, Sid, Chl, Ep	Gabbro and sediments	Brittle-ductile shear associated to folding along the MBF	5, 7, 10, 11
McCormack Brook East	Au, As, Cu	Asp, Py, Cp, Au	Qtz, Ank, Sid, Hem	Gabbro and breccia	Brittle-ductile deformation, associated to the MBF	5, 7, 8, 10, 11
Jonpol	Au, As, Cu	Py, Asp, Cp, Au	Qtz, Ank, Sid, Chl, Hem, Fus	Gabbro and breccia	Shearing, veining, and refolded folds associated to the MBF	1, 5, 7, 8, 10, 11
Dalhousie Road	Au, As	Py, Ars, Mag	Qtz, Ank, Ser	Diorite and Limestone	Dextral shearing and related veining along subsidiary faults of the MBF	5, 8, 9, 11
Mulligan Gulch	Au, Zn, Cu, Pb	Py, Sp, Cp, Ga	Qtz, Ank, Arg, Ser, Ep	Monzonite	Extensional, and sheared veins associated to the RBMF	5, 7, 8, 9, 10, 11
Upsalquitch Forks	Au, Ag, Cu, Zn, Pb, Hg, Bi	Py, Cp, Ga, Po, Sp	Qtz, Cal, Ser, Hem	Diorite and basalts	Ductile fault zones and related veining north of the RBMF	2, 6, 8, 9, 10, 11

Abbreviations. ***Commodities:*** **Ag-** Silver, **As-** Arsenic, **Au-** Gold, **Bi-** Bismuth, **Cu-** Copper, **Hg-** Mercury, **Pb-** Lead, **Zn-** Zinc. ***Metallic minerals:*** **Asp-** Arsenopyrite, **Au-** Gold, **Cp-** Chalcopyrite, **Ga-** Galena, **Mg-** Magnetite, **Py-** Pyrite, **Po-** Pyrrhotite, **Sp-** Sphalerite. ***Alteration minerals:*** **Ank-** Ankerite, **Cal-** Calcite, **Chl-** Chlorite, **Ep-** Epidote, **Fus-** Fuschite, **Hem-** Hematite, **Kao-** Kaolinite, **Qtz-** Quartz, **Ser-** Sericite, **Sid-** Siderose. ***References:*** **1-** Adair, 1988, **2-** Burton, 1987, **3-** Clemson, 1986, **4-** Fitzpatrick, 1986a, **5-** Fitzpatrick, 1986b, **6-** Høy, 1987, **7-** Mitton, 1988, **8-** Murphy, 1989, **9-** Ruitenberg et al., 1989, **10-** Ruitenberg et al., 1990, **11-** Walker and McCutcheon, 1991.

faults associated with the first compression system are attributed to an early event that was associated with the development of veinlet stockworks. To verify this hypothesis, veinlet systems of the area were analyzed and compared to computed $\sigma1$ vectors (Fig. 6D). The intersection between planes of maximum concentration of veinlet poles gives the $\sigma2$ (171/72°). $\sigma1$ (331/17°) and $\sigma3$ (064/05°) are given by the bisectrices of acute and obtuse angles respectively, and are perpendicular to $\sigma2$ on the great circle formed by the plane of poles (Fig. 6D). Similar results obtained for the computed $\sigma1$ vector (353/13°) given by striated planes, and for the $\sigma1$ vector (331/17°) calculated from stockworks, seems to relate these structures to the same compressive system. As mentioned above, stockworks are crosscut by the schistosity and seem to predate the folding. Therefore, the second compressional system at south-southwest orientation is correlated with the development of the regional schistosity and associated folds. The $\sigma1$ vector (209/10°) calculated from striated planes (Fig. 6C) is subparallel to the $\sigma1$ vector (020/13°) that is deduced from the schistosity trend (Fig. 6D). As for the McCormack-Jonpol results, the evolution of $\sigma1$ from SSE-NNW to SSW-NNE suggests that related faults evolved within a clockwise rotational deformation regime.

DISCUSSION AND CONCLUSION

Gold occurrences of the Upsalquitch Forks area are related to brittle-ductile deformation associated with fault zones. Gold-bearing structures are hosted by sedimentary rocks and/or by mafic to felsic intrusive rocks of Late Silurian to Early Devonian age. Surrounding sedimentary rocks are generally more deformed than the intrusions. Gold mineralization was concentrated in a plumbing system formed by shear-related, open space filling. Deformation-related, widespread alteration crosscuts an earlier and less extensive magmatic alteration. Main characteristics of gold showings of the area are summarized in Table 1. Differences among the various occurrences are attributed to the geometry of the RBMF and related faults, as well as to the nature of the host rocks. Along the MBF and south of it, gold occurs in gabbroic rocks and is found with arsenopyrite, pyrite, and chalcopyrite. Alteration products consist of quartz and Fe-carbonate minerals. Other gold showings of the area are hosted in diorite or in monzonite. Pyrite, sphalerite, chalcopyrite, and galena are found with quartz and ankerite/calcite alteration zones.

South of the RBMF, gold showings are hosted by the Pridolian (408-414 ± 6 Ma) Budworm City and Greys Gulch formations (McCutcheon and Bevier, 1990). The Mulligan Gulch porphyry (419 ± 1 Ma, McCutcheon and Bevier, 1990) and others mafic plutons were thus emplaced closely after the sedimentary host rocks (Murphy, 1989). North of the RBMF, the Jerry Ferguson porphyry (401 ± 1 Ma) cuts rocks of Gedinnian age (401 ± 9 to 408 ± 6 Ma, Fyffe and Fricker, 1987). Once again, the time interval between the deposition of sediments and intrusion emplacement is very short. Dostal et al. (1989) interpreted these magmatic rocks to have been formed in an extension environment during the development of pull-apart basins.

In the Upsalquitch Forks area, a progressive rotation of the stress field is suggested by paleostress analysis of fault data. This is in agreement with the transpressive tectonic regime attributed to pre-Carboniferous fault zones of northern New Brunswick (van Staal et al., 1990; Tremblay and Dubé, 1991). Transpressive tectonics appear to be characteristic of this segment of the Northern Appalachians as suggested by various studies along other strike-slip fault zones of the area such as the Elmtree fault (Tremblay and Dubé, 1991), the Fredericton and Saint-George faults (Ruitenberg et al., 1990) in New-Brunswick, and the Grand Pabos fault (Malo and Béland, 1989) in Gaspé.

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Effective porosity of tight shales from the Venture Gas Field, offshore Nova Scotia

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Abstract

Quantitative models are being developed to describe the hydrocarbon generation, migration and accumulation history of the Sable and Jeanne d'Arc basins, offshore eastern Canada. These models require input of petrophysical information of shales, specifically, on the effect of pressure on permeability/porosity as a function of pressure. As part of a study being carried out to provide such information, effective porosity of tight shales has been measured on a suite of 10 shale samples from depths of 4600-5600 m in the Venture Gas Field offshore Nova Scotia. Various measuring techniques and procedures have been applied with the participation of different laboratories. The purpose is to determine measurement accuracy of shale porosities. The results indicate that porosities are 1.5-12.0 %, values at the low end of shale porosities, and are generally consistent with porosities obtained by different methods and laboratories. However, the results may be less consistent for samples associated with the presence of certain clay minerals.

Résumé

On met actuellement au point des modèles quantitatifs pour décrire la formation des hydrocarbures, leur migration et leur accumulation dans les bassins de l'île de Sable et de Jeanne d'Arc, au large de la côte est du Canada. Ces modèles nécessitent l'entrée de données pétrophysiques sur les shales, en particulier sur l'effet de la pression sur la perméabilité et la porosité. Dans la cadre d'une étude entreprise pour recueillir ces données, on a mesuré la porosité efficace de shales compacts sur une série de dix échantillons de shale prélevés à des profondeurs variant entre 4600 et 5600 m dans le champ gazéifère Venture au large de la Nouvelle-Écosse. Avec la participation de différents laboratoires, on a appliqué diverses techniques et méthodes de mesure. L'objectif visé est de déterminer la précision des mesures. Les résultats indiquent que les porosités varient entre 1,5 et 12,0 %, valeurs situées à l'extrémité inférieure de la porosité des shales, et elles sont en général comparables aux porosités obtenues par différentes méthodes et différents laboratoires. Cependant, les résultats pourraient différer quelque peu lorsque les échantillons contiennent certains minéraux argileux.

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INTRODUCTION

Quantitative models are being developed that describe the hydrocarbon generation, migration and accumulation history of the Sable and Jeanne d'Arc basins, offshore eastern Canada. These models require petrophysical information of shales as input. Although there is an abundance of such information on sandstones and carbonate rocks, there is a lack of similar data for shales (Mudford and Best, 1989). Specifically, information on the effect of pressure on permeability and porosity and on relative permeability of shales.

Various studies are being carried out to provide such petrophysical data. As part of these studies, the effective porosity of tight shales has been measured on a suite of 10 shale samples from depths of 4600-5600 m in the Venture Gas Field offshore Nova Scotia. Various techniques and procedures have been applied. Different laboratories have also participated. The purpose of this study is to determine the accuracy and error range of shale porosity data being input into these models. Although the American Petroleum Institute's recommended practices (API, 1960) are generally followed, recent studies (Soeder, 1986) have indicated the possibility of reduced measurement accuracy when certain clay minerals are present.

The distribution of oil and gas in sedimentary basins is the result of a complex, dynamic interaction of chemical, physical and geological processes. The accuracy and predictive ability of hydrocarbon charge models for these sedimentary basins depend largely on the quantity and quality of the data available from the basins, and on the validity of the assumptions made during model construction. A long term study is underway to develop such models for the Jeanne d'Arc Basin offshore Newfoundland and Sable basin offshore Nova Scotia. The purpose is to determine the hydrocarbon charge risk and to provide quantitative information on the remaining resource endowment of the basins. The study is coupling the good quality data base, currently available, with previous interpretations (Keen and Williams 1990; Geological Survey of Canada East Coast Basin Atlas Series; Bell, 1989; Cant, 1990) and modern modelling techniques (Welte and Yalcin, 1987; Ungerer et al, 1990).

This study is seen as a series of linked models that account for:

1. The initial deposition preservation and distribution of organic matter, its quality, type and volume (Source Potential).
2. The subsidence, compaction, pore pressure, thermal maturity and generation history of the source rock system (Primary Migration).
3. The subsidence, compaction, thermal, pressure and diagenetic history of the aquifer system (Secondary Migration).
4. The geometrical and mechanical evolution of the major traps in the basin (Entrapment).

Modelling source rock and reservoir systems in this way requires extensive assumption and information regarding specific rock properties, particularly their progressive evolution with basin subsidence and compaction. For example, information on porosity reduction in muds as they are compressed, heated, and in some cases pressurized, is essential in attempts to derive accurate compaction-corrected subsidence histories. Furthermore, an understanding of porosity/permeability relationships with depth, including permeability characteristics of tight shales, is required to model accurately the sealing capacity of shales at depth and to simulate numerically formation pore pressure evolution through time. Comprehensive data regarding these properties are lacking in many areas, particularly the physical and chemical properties of shales. Therefore, studies are also underway to extend our basic knowledge of the physical properties of shales both in their present day state (ie. compacted, tight) and in a temporal framework.

Various methods of measuring bulk volume, V, and pore volume, V_p, for effective porosity, ϕ_E, determination:

$$\phi E = V_p/V \tag{1}$$

are listed in the American Petroleum Institute's Recommended Practice for Core-Analysis Procedures (API, 1960). The calliper method (API, 1960) has generally been used for our bulk volume determination of low porosity rocks (Katsube, 1981; Katsube and Hume, 1987; Katsube et al., 1990; Katsube and Scromeda, 1991). The advantages of the method are: the specimens are not contaminated and therefore can be used for other tests, the procedure is rapid and simple, and accurate results can be obtained if the specimen is of a true regular shape. The limitations are that irregular specimen shapes can not be measured. However, this problem has often been overcome by determining the bulk density, δ, of a regular shaped specimen of the same sample by use of the following equation

$$V = W_D / \delta, \tag{2}$$

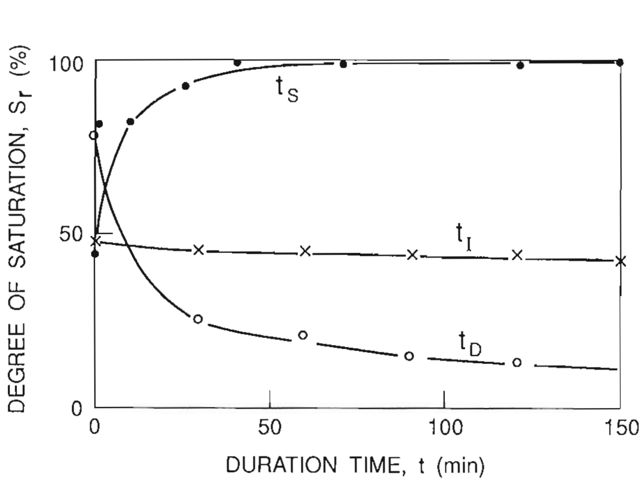

Figure 1. Typical examples of vacuum drying (t_I), saturating (t_s, under atmospheric pressure), and oven-drying curves (t_D) from Katsube and Scromeda (1991).

and then using that value to determine the volume of the irregular shaped one. The method usually used for pore volume determination of low porosity (<1.0%) rocks has been one similar to the fluid saturation method (API, 1960) using water (Katsube, 1981; Melnyk and Skeet, 1986; Katsube and Hume, 1987; Katsube and Scromeda, 1991). It takes the difference between the weight of the fluid saturated rock, W_W, and that of the dry rock, W_D:

$$V_p = (W_w - W_D) / \delta_w \quad (3)$$

where δ_w is the bulk density of water. Effective porosity (ϕ_E) is derived by inserting these results into equation (1). The advantages of this method are that many specimens can be handled at one time and the procedure is very accurate (API, 1960). The limitations are that the procedure is slow and special precaution is necessary to insure complete saturation.

The procedure that has commonly been used to determine W_W and W_D for low porosity rocks follows the next three steps (Katsube and Scromeda, 1991):

(1) dry and de-gas a rock specimen under vacuum at room temperature,

(2) saturate the specimen by introducing water into the vacuum chamber, and leaving the sample immersed under atmospheric pressure until measuring W_W,

(3) driving out the water by heating (105°C) the saturated specimen, before measuring W_D.

The results using this procedure have generally been consistent with those obtained by other methods for low porosity rocks (Katsube, 1981; Katsube and Walsh, 1987; Katsube et al., 1990, 1991). However, a recent study (Soeder, 1986) has shown that a drying temperature above 50°C can result in textural changes in certain clays which can reduce measurement accuracy in some cases. In addition, some of our shale samples have shown slight inconsistencies under repetitive measurements. For this reason, a comprehensive study of the effective porosity measuring methods and procedures for shales is underway. This paper compiles some of the experimental results obtained to date for the 10 samples from the Venture Gas field. The measurements are made under varied procedures including various drying temperatures using the fluid saturation method (API, 1960) for pore volume determination. The results are compared with those obtained by other methods and by other laboratories.

METHOD OF INVESTIGATION

Samples

The 10 shale samples have been obtained from depths between 4700m and 5600m in three wells located in the Venture Gas Field, offshore Nova Scotia (Katsube et al., 1991), as indicated in Table 1. The results of lithological and petrofacies classification (Katsube et al., in preparation) are listed in Table 2. Cylindrical plugs with a diameter of 2.54 cm were cored in the vertical direction from 10.16 cm split core samples. Such plugs were obtained from all 10 samples. The plugs were then cut to a thickness of 0.5-1.0cm for the porosity measurements.

Experimental projects

Effective porosity measurements of these 10 shale samples were first made in 1988. Triplicate measurements were made on identical specimens, the results of only the first set being published (Katsube et al., 1990, 1991). These measurements will be referred to as Experimental Project 1988, EP-1988. Several of the same specimens of these samples (samples number 5, 6 and 10) were used in an experimental project in

Table 1. Sampling information on locations and depths from Katsube et al. (1991)

Sample Number	Venture well (I.D.)	Depth (m)	Length (cm)	Direction
1	B-13	4693.40	10	H & V
2	"	4694.95	5	V
3	B-43	4916.36	4	V
4	"	4962.48	8	V
5	B-52	5122.35	6	V
6	"	5131.54	6	V
7	B-52	5271.59	5	V
8	"	5273.53	5	V
9	B-52	5553.65	6	V
10	"	5556.95	10	H & V

H = sample taken in horizontal direction
V = sample taken in vertical direction

Table 2. Lithological and petrofacies classification from Katsube et al. (in preparation)

Sample Number	Lithologic Description	Petrofacies
1	Litharenitic, Shaley and Sandy Siltstone	Siltstone
2	Calcareous, Fossiliferous and Litharenitic Sandstone	Sandstone
3	Litharenitic, Shaley and Sandy Siltstone	Siltstone
4	Organic Matter-Rich, Silty Shale	Shale
5	Organic Matter-Rich, Silty and Sandy Shale	Shale
6	Sandy, Skeletal Lime Wackestone	Lime Wackestone
7	Litharenitic, Shaley and Sandy Siltstone	Siltstone
8	Shaley, Litharenitic Sandstone	Sandstone
9	Interbedded Sublithorenitic Sandstone and Silty Shale	Shale
10	Parallel Laminated Organic Matter-Rich Silty Shale	Shale

Table 3a. An example of the results for saturating sample 5 measured under Experimental Program, ED-1991(1): extent of saturation with time (t_s) under atmospheric pressure

time (t_s) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
0	5.5136	74.4	67.8
1	5.5368	97.6	88.9
5	5.5381	98.9	90.1
10	5.5393	100.1	91.2
15	5.5405	101.3	92.3
20	5.5405	101.3	92.3
30	5.5410	101.8	92.7
45	5.5421	102.9	93.7
60	5.5432	104.0	94.7
90	5.5459	106.7	97.2
120	5.5459	106.7	97.2
195	5.5468	107.6	98.0
225	5.5471	107.9	98.3
285	5.5474	108.2	98.5
315	5.5474	108.2	98.5
375	5.5474	108.2	98.5
1435	5.5490	109.8	100.0

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

Table 3b. An example of the results for vacuum drying sample 5 measured under ED-1991(1): extent of saturation with vacuum drying time (t_l)

time (t_l) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
0	5.5490	109.8	100.0
5	5.5359	96.7	88.1
10	5.5290	89.8	81.8
25	5.5170	77.8	70.9
40	5.5102	71.0	64.7
60	5.5036	64.4	58.7
80	5.4990	59.8	54.5
110	5.4938	54.6	49.7
185	5.4846	45.4	41.4
220	5.4824	43.2	39.3
280	5.4782	39.0	35.5
310	5.4770	37.8	34.4
370	5.4740	34.8	31.7

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

1990, EP-1990, the result of number 6 being published in Katsube and Scromeda (1991). The same specimens from the 10 samples were remeasured twice in 1991, EP-1991(1) and EP-1991(2), followed by another experimental project using new specimens from 5 samples (numbers 3, 4, 5, 7 and 9), EP-1991(3). The fluid saturation method (API, 1960) was used for all of these experimental projects, but with slight variations for each case, as will be described later.

In addition, effective porosity measurements using different methods have been carried out by four different laboratories, which will be referred to in this paper as L-1 to L-4. Separate specimens were used in all of these measurements.

The liquid saturation method was used by laboratories L-1 to L-3 on selected samples. Mercury porosimetry was used at L-4 on all 10 samples and the results have been reported in Katsube et al. (1990, 1991). Further details of the measurement procedures will be described later.

Experimental procedures

The experimental procedure followed in EP-1990 forms the basis of the fluid saturation method used in all of the EP projects, with the procedures in the rest of the projects being slight variations. The experimental procedures used in EP-1990 follow the following steps (Katsube and Scromeda, 1991):

Table 3c. An example of the results for oven-drying sample 5 measured under ED-1991(1): extent of saturation with oven-drying time (t_D)

time (t_D) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
0	5.4750	35.8	32.6
30	5.4631	23.9	21.8
60	5.4587	19.5	17.8
90	5.4566	17.4	15.9
135	5.4537	14.5	13.2
210	5.4502	11.0	10.0
240	5.4490	9.8	8.9
300	5.4471	7.9	7.2
360	5.4460	6.8	6.2
1327	5.4410	1.8	1.6
(5 days later)			
1447	5.4468	7.6	6.9
1627	5.4441	4.9	4.5
1747	5.4428	3.6	3.3
2717	5.4392	0.0	0.0
2777	5.4392	0.0	0.0
2837	5.4392	0.0	0.0

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

(1) Vacuum drying

An air dried (room temperature) specimen is placed in a glass beaker using tweezers. The beaker is then placed in a vacuum chamber with vacuum applied at 760 mm Hg for periods varying from 15 to about 4000 minutes. After each of these specified periods, the sample is weighed. This procedure is repeated until a constant weight is obtained.

(2) Vacuum saturation

Vacuum is applied for 15 minutes to de-gas the specimen before it is saturated by introducing deionized, distilled water (DDW) into the vacuum chamber. After saturation, vacuum is applied for another 15 minutes for degassing before the immersed specimen is left under atmospheric pressure for periods varying from 1 to 60 minutes. The specimen is weighed after each period. Before weighing, the specimen is removed from the beaker and dried using a kimwipe until the surface does not shine due to surface moisture films. Considerable care is taken to keep the surface-drying process consistent. The specimen is returned to the beaker and the procedure is repeated until a constant weight is obtained.

(3) Oven drying

Once saturation is completed, the specimen is placed in a beaker and heated at 100-105°C for periods varying from 30 to about 1000 minutes. After each period, the sample is cooled in a desiccator for 7 to 20 minutes, weighed, and then reheated. This procedure is repeated until a constant weight is obtained.

Shorter vacuum drying, saturating and oven drying times were applied in EP-1988. They were set at 15, 40 and 240 minutes, respectively (Katsube, 1981). Steps (1) and (2) were reversed for EP-1991(1). EP-1991(2) followed the same procedure as EP-1991(1) after being water saturated for one month under atmospheric pressure. EP-1991(3) followed the same procedure as EP-1991(1), but the oven drying was carried out at 7 seven different temperature levels ranging from 50°C to 116°C.

The fluid saturation method was used in laboratory L-1, with vacuum drying at 105°C being the final step. In L-2, subsequent to the specimens being dried in a vacuum oven at 25°C for 24 hours, the helium porosity was measured. In L-3, the fluid saturation method was applied, using 30,000 ppm NaCl fluid, and oven drying under vacuum at 50°C, subsequent to the salts being removed by a methanol filled

Table 4a. An example of the results for saturating sample 5 measured under Experimental Program, ED-1991(3): extent of saturation with time (t_s) under atmospheric pressure

time (t_s) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
0	2.8634	39.3	64.9
1	2.8749	50.8	83.8
5	2.8757	51.6	85.2
10	2.8762	52.1	86.0
15	2.8771	53.0	87.5
20	2.8772	53.1	87.6
30	2.8787	54.6	90.1
45	2.8794	55.3	91.3
60	2.8807	56.6	93.4
90	2.8816	57.5	94.9
120	2.8823	58.2	96.0
160	2.8827	58.6	96.7
180	2.8831	59.0	97.4
210	2.8833	59.2	97.7
290	2.8839	59.8	98.7
325	2.8841	60.0	99.0
390	2.8847	60.6	100.0
420	2.8847	60.6	100.0
1430	2.8847	60.6	100.0

W_r :Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

Table 4b. An example of the results for vacuum drying sample 5 measured under ED-1991(3): extent of saturation with vacuum drying time (t_l)

time (t_l) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
0	2.8832	59.1	97.5
5	2.8735	49.4	81.5
10	2.8691	45.0	74.3
25	2.8624	38.3	63.2
40	2.8585	34.4	56.8
60	2.8553	31.2	51.5
80	2.8530	28.9	47.7
100	2.8513	27.2	44.9
130	2.8486	24.5	40.4
160	2.8472	23.1	38.1
190	2.8458	21.7	35.8
265	2.8431	19.0	31.4
325	2.8414	17.3	28.6
355	2.8412	17.1	28.2
400	2.8397	15.6	25.7
(continued 17 hours 31 min. later)			
420	2.8495	25.4	41.9
440	2.8477	23.6	38.9
535	2.8433	19.2	31.7
615	2.8417	17.6	29.0
675	2.8391	15.0	24.8
735	2.8386	14.5	23.9

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

soxholet extractor. The methods and procedures for the mercury porosimetry measurements used in L-4 for these samples are described elsewhere (Katsube, 1981; Katsube et al., 1991).

Parameters

The duration times for drying and degassing under vacuum, for saturating under atmospheric pressure, and for heat treatment are represented by t_I, t_S and t_D, respectively, and are expressed in minutes. Duration of saturation, t_S, is measured from the point that the 15 minutes of vacuum drying and 15 minutes of vacuum degassing the immersed sample is completed. The weight of a specimen at any given time is represented by W_r in grams. When W_r reaches a constant value during the saturation process, the sample is considered to be fully saturated and its weight is represented by W_W. When W_r reaches a constant value during oven drying, the sample is considered to be completely dried and its weight is represented by W_D. The weight difference between W_r and W_D is ΔW_r:

$$\Delta W_r = W_r - W_D. \tag{4}$$

The degree of saturation, S_r, is the weight of the water content (ΔW_r) at any given time over the weight of the maximum water content (W_W - W_D), and is expressed as follows:

Table 4c. An example of the results for oven-drying sample 5 measured under ED-1991(3): extent of saturation with oven-drying time (t_D) at 50°C and 60°C

time (t_D) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
50 °C			
0	2.8530	28.9	47.7
60	2.8450	20.9	34.5
120	2.8430	18.9	31.2
180	2.8419	17.8	29.4
300	2.8402	16.1	26.6
360	2.8397	15.6	25.7
435	2.8393	15.2	25.1
(continued 16 hours, 31 min. later)			
465	2.8461	22.0	36.3
495	2.8442	20.1	33.2
525	2.8425	18.4	30.4
555	2.8413	17.2	28.4
585	2.8402	16.1	26.6
630	2.8392	15.1	24.9
720	2.8377	13.6	22.4
780	2.8377	13.6	22.4
855	2.8374	13.3	22.0
60 °C			
0	2.8496	25.5	42.1
30	2.8436	19.5	32.2
60	2.8415	17.4	28.7
90	2.8396	15.5	25.6
150	2.8379	13.8	22.8
235	2.8364	12.3	20.3
270	2.8361	12.0	19.8
300	2.8361	12.0	19.8
360	2.8360	11.9	19.6
(continued 16 hours, 28 min. later)			
390	2.8421	18.0	29.7
430	2.8392	15.1	24.9
460	2.8384	14.3	23.6
490	2.8377	13.6	22.4
550	2.8363	12.2	20.1
650	2.8348	10.7	17.7
710	2.8343	10.2	16.8
750	2.8342	10.1	16.7

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

Table 4d. An example of the results for oven-drying sample 5 measured under ED-1991(3): extent of saturation with oven-drying time (t_D) at 70°C and 80°C

time (t_D) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
70 °C			
0	2.8493	25.2	41.6
30	2.8417	17.6	29.0
60	2.8392	15.1	24.9
120	2.8365	12.4	20.5
150	2.8355	11.4	18.8
240	2.8340	9.9	16.3
300	2.8336	9.5	15.7
345	2.8332	9.1	15.0
(continued 16 hours, 22 min. later)			
375	2.8423	18.2	30.0
410	2.8394	15.3	25.3
470	2.8369	12.8	21.1
530	2.8353	11.2	18.5
620	2.8335	9.4	15.5
680	2.8331	9.0	14.9
725	2.8328	8.7	14.4
80 °C			
0	2.8483	24.2	39.9
30	2.8395	15.4	25.4
60	2.8367	12.6	20.8
90	2.8352	11.1	18.3
130	2.8337	9.6	15.8
220	2.8315	7.4	12.2
280	2.8306	6.5	10.7
340	2.8294	5.3	8.8
(continued 16 hours, 5 min. later)			
370	2.8380	13.9	22.9
400	2.8357	11.6	19.1
440	2.8335	9.4	15.5
470	2.8324	8.3	13.7
530	2.8316	7.5	12.4
630	2.8303	6.2	10.2
690	2.8292	5.1	8.4
730	2.8292	5.1	8.4

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

$$S_r = \Delta W_r / (W_W - W_D). \quad (5)$$

EXPERIMENTAL RESULTS

Results of three repeated measurements (i, ii, iii) on the specimens of set "a" under Experimental Project EP-1988 are listed in Table 6. Similar procedures have been applied in the three sets of measurements, as previously stated. The results of set i) have previously been published (Katsube et al., 1990, 1991). The porosity values show small variations but can be generally considered constant, except for sample number 4. It shows an increase from 5.8 % to 8.5 % as a result of the successive measurements. Samples 8 and 10 also show continuous increases, but they are small.

The results of measurements on three samples from specimen set "a" performed under EP-1990 are also listed in Table 6. A slightly different procedure from EP-1988 has been applied in these measurements, as previously stated. The vacuum drying, saturating, and oven-drying curves have been determined for these samples, with a typical example from Katsube and Scromeda (1991) shown in Figure 1. Details of the experimental results for sample number 6 have been previously published (Katsube and Scromeda, 1991).

Results of repeated measurements on the 10 samples of specimen set "a" performed under EP-1991(1) and EP-1991(2) are listed in Table 6. Identical procedures to those of EP-1990 have been applied as previously stated. An example of the results of vacuum drying, saturating, and oven-drying for sample number 5 measured under EP-1991(1) are listed in Table 3.

The results of measurements on 5 samples from specimen set "f" performed under EP-1991(3) are also listed in Table 6. Identical procedures to those of EP-1991(1) and EP-1991(2) have been applied, except for the oven drying procedure which dries the specimens at temperatures varying from 50°C to 116°C, as previously stated. An example of the

Table 4e. An example of the results for oven-drying sample 5 measured under ED-1991(3): extent of saturation with oven-drying time (t_D) at 90°C and 100°C

	time (t_D) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
90 °C				
	0	2.8437	19.6	32.3
	30	2.8358	11.7	19.3
	60	2.8336	9.5	15.7
	90	2.8319	7.8	12.9
	120	2.8317	7.6	12.5
	180	2.8306	6.5	10.7
	240	2.8291	5.0	8.3
	270	2.8286	4.5	7.4
	380	2.8278	3.7	6.1
	(continued 16 hours, 14 min. later)			
	410	2.8368	12.7	21.0
	440	2.8341	10.0	16.5
	500	2.8327	8.6	14.2
	560	2.8304	6.3	10.4
	650	2.8293	5.2	8.6
	710	2.8287	4.6	7.6
	755	2.8278	3.7	6.1
100 °C				
	0	2.8475	23.4	38.6
	30	2.8359	11.8	19.5
	60	2.8336	9.5	15.7
	90	2.8314	7.3	12.1
	120	2.8308	6.7	11.1
	180	2.8290	4.9	8.1
	270	2.8277	3.6	5.9
	330	2.8275	3.4	5.6
	370	2.8269	2.8	4.6
	(continued 16 hours, 36 min. later)			
	400	2.8346	10.5	17.3
	430	2.8324	8.3	13.7
	460	2.8311	7.0	11.6
	520	2.8290	4.9	8.1
	610	2.8275	3.4	5.6
	670	2.8271	3.0	5.0
	720	2.8265	2.4	4.0

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

Table 4f. An example of the results for oven-drying sample 5 measured under ED-1991(3): extent of saturation with oven-drying time (t_D) at 116°C

	time (t_D) (min)	W_r (g)	ΔW_r (mg)	S_r (%)
116 °C				
	0	2.8455	21.4	35.3
	30	2.8336	9.5	15.7
	60	2.8307	6.6	10.9
	90	2.8289	4.8	7.9
	120	2.8277	3.6	5.9
	150	2.8270	2.9	4.8
	210	2.8256	1.5	2.5
	270	2.8248	0.7	1.2
	330	2.8245	0.4	0.7
	(continued 16 hours, 11 min. later)			
	360	2.8282	4.1	6.8
	390	2.8268	2.7	4.5
	420	2.8258	1.7	2.8
	450	2.8256	1.5	2.5
	510	2.8250	0.9	1.5
	590	2.8244	0.3	0.5
	650	2.8242	0.1	0.2
	690	2.8241	0.0	0.0

W_r : Weight of specimen at time t.
ΔW_r: Weight difference between W_r and W_D of the oven-dried specimen.
S_r : Degree of saturation.

results of vacuum drying, saturating, and oven-drying for sample number 5 measured under EP-1991(3) is listed in Table 4.

The results for W_W and W_D for EP-1991 to EP-1991(3) are listed in Table 5. Only the final results for measurements at 116°C have been used to determine W_D for EP-1991(3). The effective porosities (ϕ_E) derived using equations (1), (2) and (3) are listed in Tables 5 and 6. The degree of saturation (S_r) has been calculated (equation (5)) for every measurement of W_r and is listed in Tables 3a to 4f. A typical example of the three curves (from Katsube and Scromeda, 1991): vacuum drying of the air dried sample, saturation under atmospheric pressure, and oven-drying, is shown in Figure 1. The vacuum drying curve expressed by t_I in Figure 1 decreases slightly with time. The saturation curve expressed by t_S increases rapidly with time and reaches a constant value. The oven-drying curve expressed by t_D decreases with time and eventually also reaches a constant value.

Table 5. Final results of the effective porosity measurements for Experimental Projects EP-1991(1) to EP-1991(3)

Sample No.	δ (g/cc)	W_w (g)	W_D (g)	ΔW (mg)	ϕ_E (%)
EP-1991(1)					
1	2.77	7.0366	6.8871	149.5	6.0
2	2.58	9.6622	9.4528	209.4	5.7
3	2.62	8.6975	8.4216	275.9	8.6
4	2.80	7.5363	7.2234	312.9	12.1
5	2.64	5.5490	5.4392	109.8	5.3
6	2.73	7.6619	7.6238	38.1	1.4
7	2.65	9.8545	9.5463	308.2	8.6
8	2.57	7.2852	6.9583	326.9	12.1
9	2.69	8.7772	8.7289	48.3	1.5
10	2.74	4.5374	4.4043	133.1	8.3
EP-1991(2)					
1	2.77	7.0011	6.8283	172.8	7.0
2	2.58	9.6414	9.4184	223.0	6.1
3	2.62	8.7125	8.3500	362.5	11.4
4	2.80	7.5775	7.1773	400.2	15.6
5	2.64	5.5580	5.4388	119.2	5.8
6	2.73	7.6678	7.6082	59.6	2.1
7	2.65	9.8443	9.5023	342.0	9.5
8	2.57	7.2474	6.9236	323.8	12.0
9	2.69	8.7665	8.7040	62.5	1.9
10	2.74	4.5457	4.3938	151.9	9.5
EP-1991(3)					
3	2.62	6.9212	6.7086	212.6	8.3
4	2.80	7.0193	6.7720	247.3	10.2
5	2.64	2.8847	2.8241	60.6	5.7
7	2.65	7.3155	7.1077	207.8	7.8
9	2.69	6.3262	6.2494	76.8	3.3

$\phi_E = \delta(W_w - W_D) / W_D$; from equations (1), (2) and (3).
W_w = wet weight   δ = bulk density
W_D = dry weight ϕ_E = effective porosity
$\Delta W = W_w - W_D$

Table 6. Comparison of effective porosity (ϕ_E) measurements made by different methods and procedures for the shale samples from the Venture Gas Field. The units are expressed in percent (%)

Smpl No.	EP-1988 i)	ii)	iii)	EP-1990	EP-1991 (1)	EP-1991 (2)	EP-1991 (3)	ϕ_{Hg}	Lab
1	3.8	3.1	3.5		6.0	7.0		4.9	4.8[c] 3.5[d]
2	5.1	4.5	4.7		5.7	6.1		7.7	
3	4.6	4.0	4.1		8.6	11.4	8.3	6.3	
4	5.8	6.7	8.5		12.1	15.6	10.2	7.6	
5	3.7	3.5	4.0	6.8	5.3	5.8	5.7	3.7	
6	0.8	0.7	0.8	0.93	1.4	2.1		1.6	
7	5.4	4.9	5.3		8.6	9.5	7.8	5.9	
8	9.2	9.8	10.2		12.1	12.0		9.1	7.7[d]
9	1.0	0.8	0.8		1.5	1.9	3.3	2.3	
10	5.8	6.2	6.6	7.4	8.3	9.5		5.8	5.7[c] 5.5[e]
Spcm	a	a	a	a	a	a	f	b	c,d,e

Smpl : Sample
No. : Number
Spcm : Specimen sets
ϕ_{Hg} : Effective porosity determined by mercury porosimetry (L-4, Katsube et al, 1990,1991).
[c] : Measurements performed by laboratory L-1.
[d] : Measurements performed by laboratory L-2.
[e] : Measurements performed by laboratory L-3.

Results of the measurements performed by the four different laboratories, L-1 to L-4, on specimen sets b, c, d and e are also listed in Table 6. Specimen set b was used for mercury porosimetry (Katsube et al., 1990, 1991) by L-4. Specimens c, d and e were used for the fluid saturation (including helium porosity) methods by laboratories L-1, L-2 and L-3, respectively.

DISCUSSION AND CONCLUSIONS

The results obtained for sample #5 (tables 3 and 4) indicate the following optimum saturation, vacuum pumping, and oven-drying times:

1. A saturation time of 285 min. is sufficient to obtain a degree of saturation (S_r) of approximately 98.5% to 98.7% for sample 5. Similar results were obtained for the remaining shale samples with degrees of saturation ranging from 96.9% to 100% after 285 min. of saturation.
2. After 735 min. of vacuum drying, a constant S_r value had not been obtained for sample 5. The vacuum pumping results for all the saturated shale samples, however, showed a significant decrease in degree of saturation with time, the final values ranging from 6.4% to 41.7%. These results differ from previous experiments (Katsube and Scromeda, 1991) where the final S_r value was 70.3%.
3. Tables 4c to 4f indicate that there was a significant decrease in the final values obtained for W_r as the oven-drying temperature was increased. An oven-drying time of 690 min. at 116°C was sufficient to obtain a

constant weight for most of the shale samples. Further experiments are required to verify that 116°C is the ideal oven-drying temperature without the risk of damaging the shale samples.

The results indicate that the effective porosities of these samples are in the range of 1.5-12.0 %, values at the low end of published shale porosities of 4-50% (Daly et al., 1966; Parkhomenko, 1967; Magara, 1971).

The three effective porosity values for each of the 10 samples obtained under EP-1988 generally show no systematic change, and can be considered consistent, except for sample number 4. It shows a increase from 5.8% to 8.5% as a result of the successive measurements. Samples number 8 and 10 also show continuous increases, but they are small.

The effective porosity values obtained under Experimental Projects EP-1990 to EP-1991(3) are generally larger than those obtained under EP-1988, a result that could be due to the shorter oven-drying time of the latter. However, when comparing the results of EP-1990 to EP-1991(2) which uses specimens "a" with those of EP-1991(3) which uses specimens "f", there are indications that the increased porosity values may also be a result of damage caused to the specimens as a result of repeated measurements. This idea is supported by the fact that many of the effective porosity values obtained by the other laboratories are often relatively close to those of EP-1988.

In conclusion, these results are interpreted to generally show relatively good consistency between porosities obtained by different methods, different procedures and different laboratories, except when damage is considered to have occurred as a result of repeated measurements. The considerable porosity change with measurement repetition shown by one of the samples (sample number 4) is associated with its containing significant amounts of illite, smectite or organic material (Katsube et al., 1990). Further study and analysis of these data are scheduled to take place in the future to clarify some of the uncertainties seen in these measurements.

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Geological Survey of Canada Project 870057

Mélanges and coticule occurrences in the northeast Exploits Subzone, Newfoundland[1]

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Abstract

The 1991 field season was spent examining the Dog Bay Melange and coticules (spessartine garnet-quartz layers) that are common associates of melanges in the northeast Exploits Subzone.

The Dog Bay Melange is defined by large altered gabbro, mafic volcanic, and smaller sandstone blocks in an otherwise bedded section of shale and siltstone at Dog Bay Point.

Dark grey shales with thin coticule layers and nodules are associated with the Dog Bay, Carmanville, and Dunnage melanges, and they occur in an area of regionally metamorphosed rocks that extends from Aspen Cove to Ragged Harbour in the Carmanville area. Regional relationships indicate a Middle Ordovician or earlier age for the coticules and associated melanges. The intermediate position of the Dog Bay Melange between the Dunnage and Carmanville melanges, and coticule occurrences in all examples imply continuity of a coastal lithic belt for 50 km across the northeast Exploits Subzone.

Résumé

Les travaux de terrain de la saison 1991 ont été consacrés à l'étude du mélange et des coticules de Dog Bay (couches de spessartite et quartz) qui sont couramment associés aux mélanges contenus dans le nord-est de la sous-zone d'Exploits.

Le mélange de Dog Bay est composé de grands blocs altérés de gabbro, de roches volcaniques mafiques et de blocs de grès de plus petite taille dans une coupe de shale et de siltstone stratifiée à la pointe Dog Bay.

Les shales gris foncé à minces couches et nodules de coticules sont associés aux mélanges de Dog Bay, de Carmanville et de Dunnage, et ils reposent dans une zone de roches métamorphisées régionalement qui s'étend d'Aspen Cove à Ragged Harbour dans la région de Carmanville. Les liens régionaux indiquent que les coticules et les mélanges associés remontent à l'Ordovicien moyen ou même à une époque antérieure. La position intermédiaire du mélange de Dog Bay entre les mélanges de Dunnage et de Carmanville, et la présence de coticules dans tous les exemples laissent supposer une continuité dans une zone lithique littorale sur une distance de 50 km à travers le nord-est de la sous-zone d'Exploits.

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INTRODUCTION

The northeast Exploits Subzone contains more examples of melanges than any other area of the Newfoundland Dunnage Zone. It contains the well-known Dunnage and Carmanville melanges and the lesser known Dog Bay Melange, all either dated or interpreted now as Ordovician. Silurian melanges, such as the Joeys Cove Melange of New World Island, are also present (Fig. 1).

The Dunnage Melange, extending across the islands of Exploits Bay from Lewisporte to Dunnage Island, was discovered by Marshall Kay (Kay and Eldridge, 1968; Kay, 1970, 1972) and described by Horne (1969), Williams and Hibbard (1976), Hibbard and Williams (1979), and Lorenz (1984). The Carmanville Melange in the eastern part of the Exploits Subzone was discovered by Kennedy and McGonigal (1972) and studied more extensively by Pajari et al. (1979) and Williams et al. (1991). Melange at Dog Bay Point, between these two occurrences, was first described by Wu (1979) and indicated on the Carmanville map of Currie et al. (1980). A somewhat larger area of melange at Dog Bay Peninsula was shown on a figure by Karlstrom et al. (1982), but without descriptions of rocks and relations. Silurian melanges of New World Island occur within a greywacke-conglomerate sequence that locally forms an upward-shoaling cover to the Dunnage Melange. They were recognized by Jacobi and Schweichert (1976) and McKerrow and Cocks (1978) and later described by Reusch (1983; 1987) and Arnott et al. (1985).

All of these melanges are being reinvestigated and compared as part of a regional study of melanges in the Newfoundland Dunnage Zone. The study is also being done in conjunction with the Lithoprobe East Program and as part of a mapping project of the Geological Survey of Canada to update the geology of the Carmanville and Comfort Cove areas (Currie, 1992). This brief report gives the results of 1991 field work on the Dog Bay Melange, and the extent and possible significance of coticule rocks that occur in or are associated with the Dunnage, Dog Bay and Carmanville melanges.

DOG BAY MELANGE

The name Dog Bay Melange is proposed for the chaotic rocks at Dog Bay Point that contain outsize blocks of altered gabbro, mafic volcanics, and quartz sandstone (Fig. 2). The chaotic rocks alternate with bedded sections that consist of

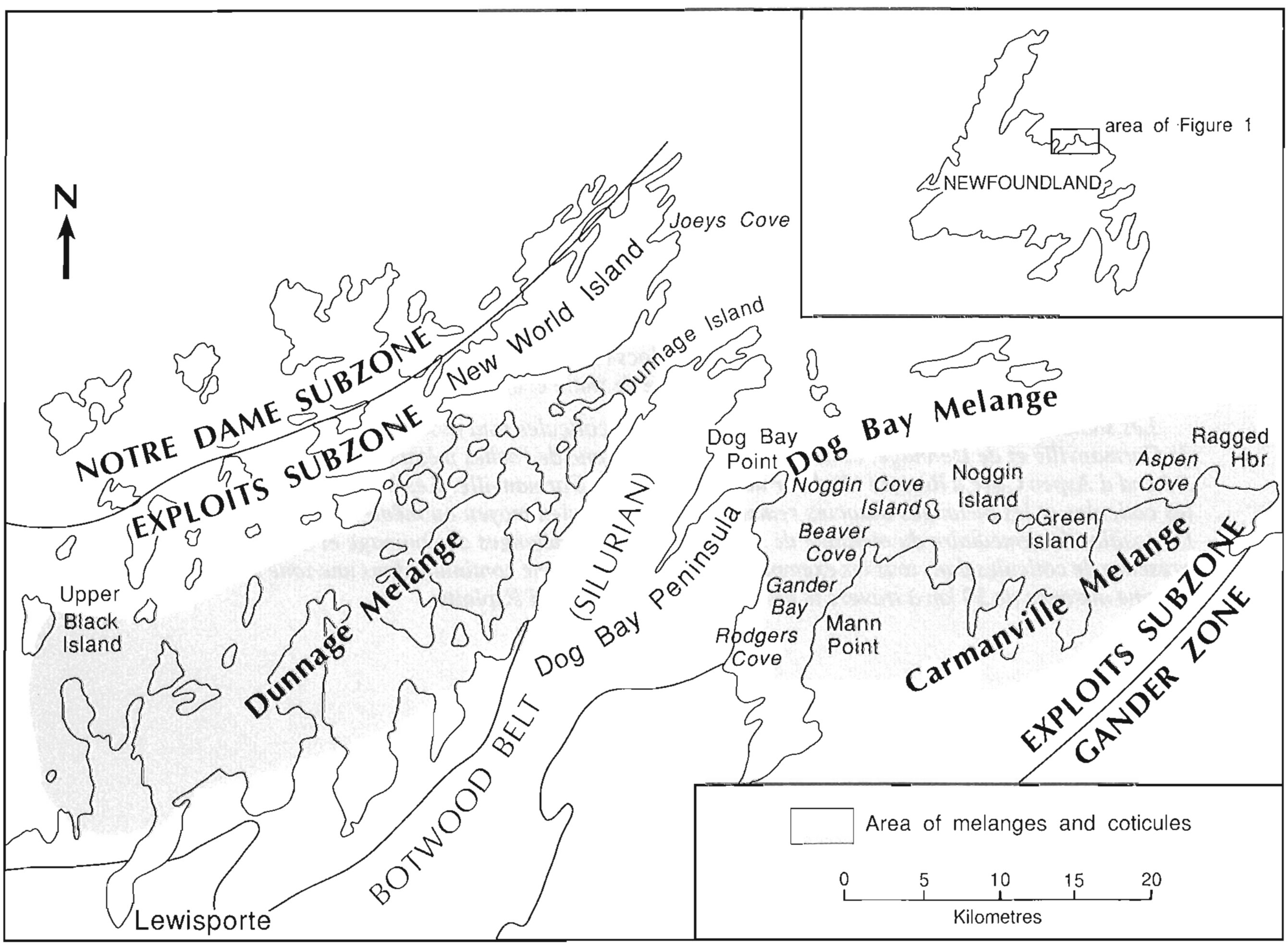

Figure 1. Distribution of melanges and coticule rocks in the northeast Exploits Subzone.

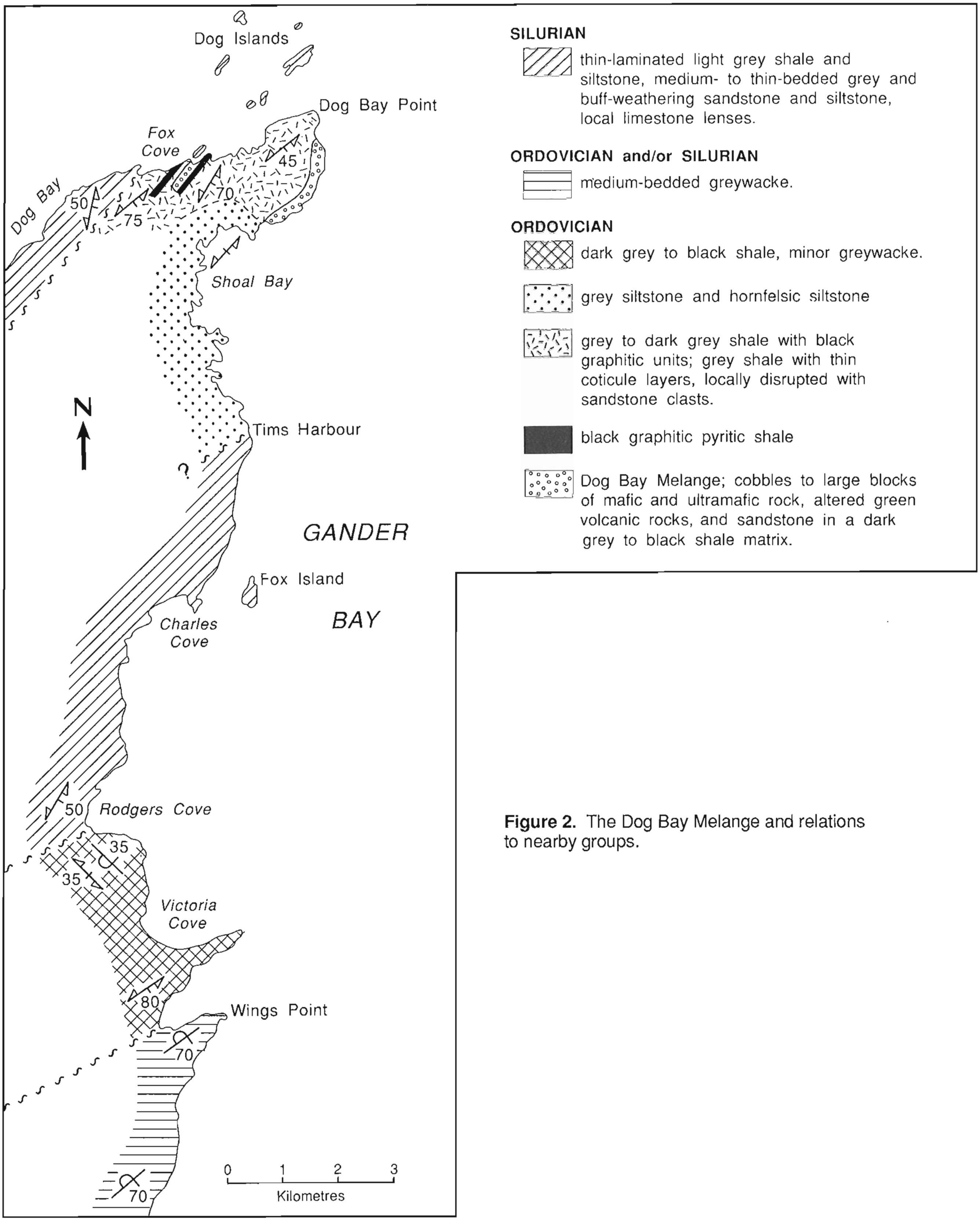

Figure 2. The Dog Bay Melange and relations to nearby groups.

highly cleaved grey shales and siltstones with conspicuous units of black pyritic shale, and dark grey shale with thin beds of fine grained garnet and quartz (coticule).

All or parts of the Dog Bay Peninsula were mapped by Patrick (1956), Williams (1964), McCann (1973), Wu (1979), Currie et al. (1980), Karlstrom et al. (1982); and compiled by Dean (1978), although there is little agreement on lithic subdivisions. The rocks were assigned to the Indian Islands Group of Late Ordovician and/or Silurian age based on an occurrence of Favosites near Tims Harbour on the east side of the peninsula (Williams, 1964; 1972) and Silurian corals and brachiopods from the west side of the peninsula (Berry and Boucot, 1970). The Dog Bay Melange is excluded from the Indian Islands Group because its lithologies are typical of Middle or Early Ordovician rocks in the Exploits Subzone.

Dark grey shales and siltstones associated with the Dog Bay Melange are faulted against light grey shales and siltstones of the Indian Islands Group to the west of Dog Bay Point at Fox Cove. On the east side of Dog Bay Point, rusty-weathering pyritic black shales with gabbro blocks are continuous with medium-bedded black shales and grey siltstones, in turn continuous with monotonous grey siltstones and hornfelsic siltstones farther south at Shoal Bay. The rocks at Shoal Bay and southward to Rodgers Cove of Gander Bay were assigned to the Indian Island Group (Williams, 1964; Dean, 1978; Wu, 1979; Currie et al., 1980). If the Dog Bay Melange is Ordovician and a Dunnage or Carmanville correlative, an unconformity or tectonic boundary is required somewhere in the vicinity of Shoal Bay or southward (Fig. 2).

Relationships between the Indian Islands Group and Ordovician rocks at Rodgers Cove to the south are also debatable. Patrick (1956) and Wu (1979) placed a stratigraphic contact on the west side of the cove between light grey thin-laminated shales and siltstones, assigned to the Indian Islands Group, and dark grey shales, assigned to the Ordovician Davidsville Group (Kennedy and McGonigal, 1972; Wu, 1979). Williams (1964) placed a fault in an area of no exposure at Rodgers Cove between the Indian Islands Group to the west and Ordovician shales and greywackes to the east. The present study also favours a major fault at Rodgers Cove, separating thin-bedded light grey shales and siltstones to the west (Indian Islands Group) and an overturned section of Ordovician black shales and coarse greywackes to the east. A structural boundary is demanded farther southeast between steeply-dipping, northwest-facing greywackes at Wings Point that are followed northwestward and up-section by older rocks.

Rocks interpreted as Ordovician at Dog Bay Point consist of about 80% layered sedimentary rocks, about 15% disrupted black and grey shales, and about 5% blocks in a black and grey shaly matrix. Larger blocks that define melange are mainly altered heterogeneous gabbro and or ultramafic rocks, internally brecciated altered green volcanic rocks, and tough quartz sandstone. Smaller blocks are sandstone, gabbro, silicic volcanic rock or porphyry, and rusty-weathering quartz-sulphidic rock. Some discrete blocks are up to 5 m in diameter (Fig. 3). Disrupted gabbro and mafic volcanic blocks define an almost continous zone that extends for 100 m just south of Dog Bay Point. The most prominent blocks occur along the shoreline from the northern mouth of Shoal Bay to Dog Bay Point. Another zone of altered calc-silicate (gabbro?) blocks from 1 m to cobble and pebble sizes occurs opposite an island 2 km west of Dog Bay Point where sheared dark grey to black shale surrounds lenses of extremely tough altered rock. Other zones consist of disrupted black and grey shale, some with pebble to cobble size clasts of sandstone. All of the blocks, with the exception of quartz sandstones, are extremely sheared and altered, and the rock types are not easily identified in the field. Some mafic plutonic and volcanic blocks may have been deformed and metamorphosed to coarse actinolite assemblages before entrainment.

The bedded sedimentary rocks associated with melanges are well-cleaved grey to dark grey and black shales and siltstones. Some are thin-bedded to finely-laminated grey shales and siltstones like those assigned to the Indian Islands Group at Rodgers Cove. Two narrow units of black pyritic graphitic shale about 10 m wide occur west of Dog Bay Point.

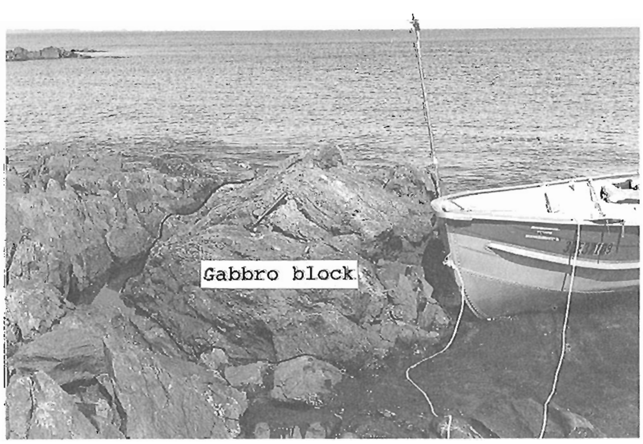

Figure 3. Gabbro block in the Dog Bay Melange near Dog Bay Point.

Figure 4. Bedded section of black shale and light grey siltstone south of Dog Bay Point.

Where sheared or broken, they contain dark tough sandstone lenses containing characteristic grains of vitreous smoky quartz. Medium-bedded black and grey shale and siltstone occur at the northern entrance of Shoal Bay (Fig. 4). These are locally disrupted and form the matrix for altered gabbro boulders and cobbles.

Another diagnostic rock type in the bedded section is dark grey shale, in 10-30 cm units, separated by 1-2 cm reddish- or bluish-weathering resistive silty beds. The silty beds consist of dense clusters of fine grained anhedral garnets and quartz. Coticules were identified here because of a familiarity with similar rocks in association with the Dunnage and Carmanville melanges, especially with the bedded sections associated with the Carmanville Melange at Green (Woody) Island, Noggin (Grassy) Island, and Noggin Cove Island (Fig. 1).

The Dog Bay Melange and bordering rocks are affected by a regional cleavage that trends northeast and dips steeply- to moderately-southeast. Both blocks and matrix, especially south of Dog Bay Point, are crossed by steep zones of intense brecciation that trend southeast. Karlstrom et al. (1982) interpreted the regional cleavage as a second phase structure that transects local folds and thrusts of earlier generation.

No fossils occur within the Dog Bay Melange but its black shales, coticules, and preponderance of mafic plutonic and volcanic blocks are all suggestive of a Middle or Early Ordovician age. The melange may therefore occupy a horst or anticline bounded on both sides by younger rocks of the Indian Islands Group (see also Karlstrom et al., 1982).

COTICULE OCCURRENCES IN THE NORTHEAST EXPLOITS SUBZONE

Coticules are thin layers or nodules of fine grained spessartine garnet and quartz developed at low metamorphic grade from manganese-rich siltstones and shales. In the northeast Exploits Subzone, they occur in bedded or disrupted sections associated with the Dunnage Melange, the Dog Bay Melange, the Carmanville Melange, and regional metamorphic rocks that extend from Aspen Cove to Ragged Harbour (Fig. 1).

The coticules occur in resistive beds about a centimetre or less in thickness that weather buff, reddish-brown, and greyish-blue. They are separated by thicker beds of grey to dark grey shales. The coticules comprise less than 10% of most shaly sequences. They display prominent folds. Some are of outcrop scale with axial planar cleavage (Fig. 5), but smaller scale, plastic styles of folding and disruption that predate the regional cleavage are also typical. Manganese nodules or concretions are associated with some occurrences in the Dunnage Melange (Hibbard and Williams, 1979).

In thin section, the rocks contain numerous small round garnets, commonly concentrated in bands, clusters or rinds bordering or surrounding quartz-rich areas. Some beds contain up to 80% garnet. The garnets form equidimensional anhedral grains with dusty dark interiors and lighter peripheries. In some examples, clear overgrowths produce euhedral outlines. Tourmaline is locally present as an accessory mineral.

In the Dunnage Melange, the coticule rocks form local bedded sections associated with black shales. In other places they are completely disrupted and form melange matrix. They are also present in sections that are interpreted to record sedimentary transitions from underlying rocks of the Exploits Group (New Bay Formation) to overlying chaotic rocks of the Dunnage Melange (Hibbard and Williams, 1979).In an example from the Dunnage Melange of Upper Black Island (Fig. 1), the coticule beds are plastically folded and disrupted. The proportions and grain sizes of garnets produce a metamorphic zonation parallel to bedding that is also present around the terminations of disrupted beds; indicating that garnet growth and metamorphic zonation followed disruption.

In bedded sediments associated with the Dog Bay Melange, the coticules are associated with black, graphitic, pyritic shales and lighter grey shales. These are host to discrete, large mafic-ultramafic and other blocks.

Dark grey to black shales contain numerous thin beds of coticule at Green (Woody) Island, Noggin (Grassy) Island, and Noggin Cove Island (Fig. 1). Olistostromes included with the Carmanville Melange (Pajari et al., 1979; Williams et al., 1991) have stratigraphic relationships with the coticule-bearing shale sections. Coticules also occur in association with black shale of the Carmanville Melange in a shoreline outcrop at Mann Point, and they are associated with volcanic rocks of the Noggin Cove Formation at Beaver Cove. These rocks in the Carmanville area were originally assigned to the Davidsville Group (Kennedy and McGonigal, 1972; Pajari et al., 1979) that occurs mainly east of Gander Bay and extends southwestward to Gander Lake and beyond.

Figure 5. Folded shales with thin coticule beds at Green (Woody) Island.

Throughout the extensive area of regional metamorphic rocks from Aspen Cove to Ragged Harbour, coticules occur in shaly sequences that locally include thick greywacke units. South of Aspen Cove, the coticule-bearing rocks are host to large discrete mafic and ultramafic blocks. The size and sparse distribution of the discrete blocks resemble occurrences of singular large blocks in the Dog Bay Melange. The coticule-bearing rocks in this area were assigned to the Gander Group of the Gander Zone (Kennan and Kennedy, 1983) but they are now assigned to the Exploits Subzone of the Dunnage Zone (Currie, 1992).

Coticules are scarce or absent in the Ordovician Exploits Group to the west of the Dunnage Melange and they are unknown in the Ordovician sedimentary rocks assigned to the Davidsville Group to the east of the Carmanville Melange. They are also unknown in the clastic sequences of Ordovician-Silurian age that occur above Caradocian black shales throughout the Exploits Bay area.

The coticule rocks are poorly dated but they are interpreted as Middle Ordovician or earlier. Those of the Dunnage Melange are pre-Caradocian as Caradocian shales form a cover to the melange at New World Island. Those of the Carmanville Melange at Green (Woody) Island may be Early Ordovician based on trace fossils (Pickerill et al., 1978; Pajari and Currie, 1978). Middle and Early Ordovician ages are typical for coticules that occur in interior parts of the Appalachian/Caledonian Orogen (Kennan and Kennedy, 1983).

The association of coticules and melanges throughout the northeast Exploits Subzone defines an Ordovician lithic belt that extends from the southwest portion of the Dunnage Melange at Lewisporte all the way to Ragged Harbour of the Carmanville area. This major belt trends easterly and none of its rocks is traceable inland. They are excluded from the Indian Islands Group at Dog Bay Point and from the Davidsville Group in the Carmanville area.

DISCUSSION

Since their discovery, the various melanges of northeast Newfoundland have been related to (1) subduction and filling of a marine trench (Bird and Dewey, 1970; Kay, 1976; McKerrow and Cocks, 1978), (2) transport of allochthons (Williams et al., 1991), (3) gravitational slumping (Horne, 1969; Kennedy and McGonigal, 1972; Hibbard and Williams, 1979; Pajari et al., 1979), (4) tectonic disruption of already lithified units (Karlstrom et al. 1982), and combinations of these processes.

A model for the Dunnage Melange as a major slump implied stratigraphic and sedimentologic links with the Exploits Group to the west (Hibbard and Williams, 1979). A similar model for the Carmanville Melange implied derivation from the east by collapse that followed obduction of the Gander River Complex (Pajari et al., 1979). However, other features are difficult to explain by simple surficial slumping; such as (1) structural and metamorphic variations in the Carmanville Melange that imply tectonic recycling of intensely deformed and metamorphosed melange (Williams et al., 1991), and (2) the Dunnage Melange is the locus for contemporary intrusions (Lorenz, 1984).

The much greater proportion of black shales and coticules associated with the melanges, compared to their proportions in nearby groups, suggests a depositional setting that contrasted with that of surrounding groups. The coticules probably represent distal chemical precipitates associated with submarine volcanism within an extensive deep marine basin. A tectonically active basin is indicated by slumping of the shaly rocks and periodic influxes of greywackes, conglomerates and bouldery olistostromes. Tectonic processes may have been operative in convergent zones within the basin or at its periphery.

The Dunnage Melange is overlain by Caradocian black shales that are also recognized in the adjacent Exploits Group, and Caradocian black shales are present in the Carmanville area. This implies cessation of tectonism across the marine basin and its opposing sides. The Caradocian shales of the Exploits Group (Lawrence Harbour Formation) and the Caradocian shales above the Dunnage Melange (Dark Hole Formation) are overlain by a thick greywacke unit (Point Leamington and Sansom formations, respectively). Probable equivalents in the Carmanville area are the greywackes at Wings Point and along the east side of southern Gander Bay. This implies infilling and shallowing of the marine basin. The greywacke unit is absent in the centrally located Dog Bay area, where possible equivalents in the basal Indian Islands Group are finer and shalier, and therefore possibly more distal. Silurian melanges at New World Island, which are equated with thrusting and telescoping of Ordovician-Silurian sections represents destruction of the marine basin.

Significantly, coticules are associated with stratabound manganese and sulphide deposits of volcanogenic origin. Renewed mineral exploration in the Exploits Subzone may therefore be warranted.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 730044

National Gravity Survey Program of the Geological Survey of Canada, 1991-92

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Hearty, D.B. and Gibb, R.A., 1992: National Gravity Survey Program of the Geological Survey of Canada, 1991-92; in *Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 129-132.*

Abstract

In 1991, six gravity surveys were completed under the national gravity survey program; two were reconnaissance surveys located in the Yukon Territory and the Arctic, two were local gravity surveys over targets in the Northwest Territories and B.C., and two were traverses along LITHOPROBE seismic profiles in Manitoba/Saskatchewan and Ontario. More than 3600 new gravity stations were added to the National Gravity Data Base as a result of these surveys. In addition, absolute gravity measurements were made at six sites in Quebec, Manitoba and the Northwest Territories in support of the GSC's crustal dynamics program. The GSC assisted two private companies with airborne gravity tests during the year.

Résumé

En 1991, six levés gravimétriques ont été réalisés dans le cadre du programme national de levés gravimétriques. Deux d'entre eux étaient des levés de reconnaissance dans le Yukon et l'Arctique, deux étaient des levés locaux au-dessus de cibles dans les Territoires du Nord-Ouest et la C.-B. et deux étaient des levés de traverses longeant les profils sismiques de LITHOPROBE au Manitoba et en Saskatchewan ainsi qu'en Ontario. Plus de 3600 nouvelles stations gravimétriques ont été ajoutées à la Base de données gravimétriques nationale grâce à ces levés. De plus, des mesures gravimétriques absolues ont été prises à six endroits au Québec, au Manitoba et dans les Territoires du Nord-Ouest pour appuyer le programme de la CGC sur la dynamique de la croûte terrestre. La CGC a collaboré avec deux sociétés privées à réaliser des essais gravimétriques aériens au cours de l'année.

INTRODUCTION

In 1991-92, the national gravity mapping program comprised two reconnaissance surveys and four local surveys over geological targets. Significant progress towards completion of regional gravity coverage of Canada's landmass and waters was made in the Yukon Territory and in the Arctic, two logistically difficult areas. Local surveys were completed in response to requests from the Continental Geoscience Division and the Lithoprobe project.

Details of these surveys and highlights from the gravity standards program, gravity map production and developments in airborne gravity are given below.

YUKON TERRITORY

The Geophysics Division, the Pacific Geoscience Centre, the Geodetic Survey of Canada, the Mapping and Charting Establishment of DND and the US Defense Mapping Agency (DMA), in a major cooperative effort, completed the regional gravity survey of part of the southern Yukon Territory in the period June to August, 1991. Approximately 1850 gravity observations, with a spacing of 10-12 km covering twenty-five 1:250,000 map sheets, were collected in year one of a three year survey program. During the first pass over the survey area gravity was read at each station and in a second pass the horizontal and vertical positions of each station were determined by the Geodetic Survey using a Litton Inertial Survey System (ISS/LASS II). GPS was used to establish control for the ISS survey and to obtain horizontal and vertical positions at a sample of 200 stations to validate the GPS technology for use next year. The costs of this survey were shared by GSC and DMA in the ratio of 1:3. The data will contribute to the National Gravity Mapping Program, to studies of resource potential and geological structure in the Yukon, and to a better determination of the geoid for the surveying industry.

BAFFIN ISLAND

Between March 25 and April 18, 1991, a gravity and bathymetric survey was conducted on the ice around Bylot Island and in Admiralty Inlet at the northwest corner of Baffin Island. The survey, part of the National Gravity Mapping Program, was carried out by the Geophysics Division in cooperation with the Canadian Hydrographic Service and PCSP. A total of 546 gravity observations and 4500 depth soundings was established on a 6 km and 2 km grid respectively. Horizontal positioning was established using portable GPS receivers (TANS Pathfinder) in differential mode.

DARNLEY BAY, NWT

In July, a local gravity survey was conducted by the Geophysics Division over the Darnley Bay gravity anomaly in response to a request from the Continental Geoscience Division. One hundred and five gravity stations were established at 1 to 3 km intervals along three intersecting profiles across the anomaly having a total length of approximately 160 km. Horizontal and vertical positions were derived from NTS maps and altimeters respectively. Topographic monuments were used as control for the altimetry.

NORTHERN MANITOBA AND SASKATCHEWAN

In response to a request from the Continental Geoscience Division, gravity profiles were observed by the Geophysics Division along Lithoprobe seismic lines between Thompson, Manitoba and Beauval, Saskatchewan. A total of 895 gravity stations was established at 1 to 2 km intervals along nine profile lines having a total length of approximately 1000 km. Horizontal and vertical positions were established under contract using traditional survey techniques and GPS for control. Gravity measurements were made at each gravity station using a LaCoste and Romberg gravity meter. Measurements were also made using the recently acquired Scintrex gravity meter to evaluate its performance.

SUDBURY, ONTARIO

Gravity profiles were observed by the Geophysics Division along Lithoprobe seismic lines near Sudbury in response to a request from the Continental Geoscience Division. One hundred and twenty-three gravity stations were established at 1 to 3 km intervals along approximately 150 km of profile line. Horizontal and vertical positions were supplied under contract using traditional survey techniques.

GRAVITY STANDARDS

Approximately 125 gravity control stations of the Canadian Gravity Standardization Net (CGSN) were inspected in southern Quebec and New Brunswick. Stations were categorized according to national standards and descriptions were updated.

Absolute gravity measurements were completed at previously observed sites in Charlevoix (2), Schefferville (2), Churchill and Yellowknife in support of postglacial rebound and secular change studies.

GRAVITY MAP PRODUCTION

Five free air and twenty-three Bouguer anomaly maps in the National Earth Science Series (NESS) (scale 1:1,000,000) were printed and are available through the publications office. Published maps and those currently in compilation are indicated in Figure 1.

One Open File Bouguer gravity map (GSC 2382) was produced for Northern Foxe Basin, District of Franklin (scale 1:1,000,000).

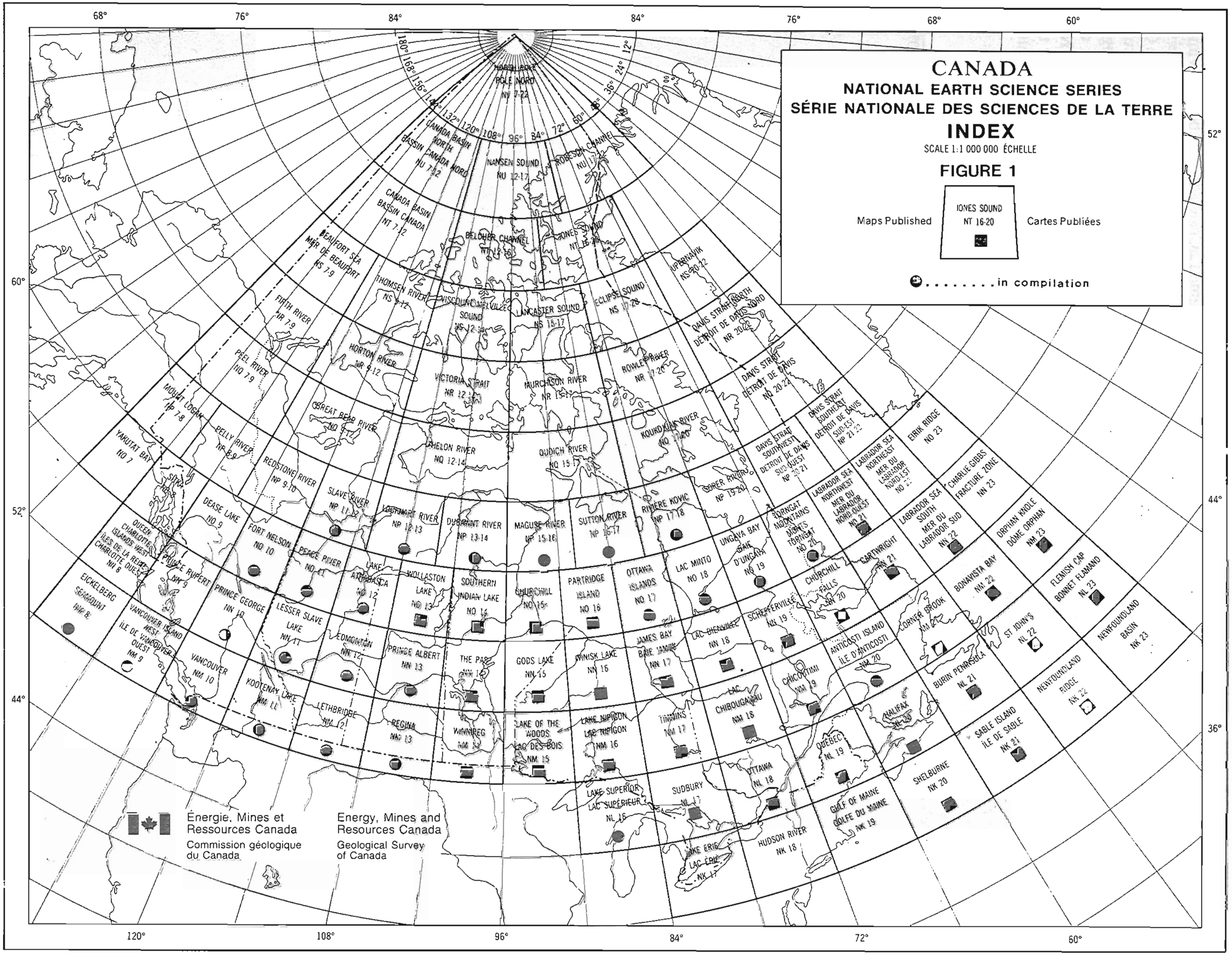

Figure 1. National Earth Science Series index for gravity maps.

AIRBORNE GRAVITY

The Geophysics Division assisted Intera Kenting with airborne gravity tests over Lake Ontario and over the Alexandria test range east of Ottawa in March. The Division's LaCoste and Romberg dynamic gravimeter (SL-1) installed in a twin engined Navajo aircraft was used to measure the gravity field. Navigation data were obtained using Trimble-4000 GPS receivers in differential mode and vertical control was provided by a radar altimeter and a barometric altimeter on board the aircraft. A negative gravity anomaly about 60 km wide and 40 mGal in amplitude, chosen as a target for the tests, was resolved to within ±5 mGal. Analysis is continuing to explain differences between radar and GPS vertical accelerations.

In collaboration with Sander Geophysics, road tests were conducted using the same LaCoste and Romberg gravimeter (SL-1), installed in a pickup truck, with two GPS receivers and associated data recording equipment. However, the GPS vertical measurements were deemed not sufficiently accurate for the test due to "unsettled" ionospheric activity during the test period. Additional testing is being pursued with dual frequency GPS receivers combined with a Litton LTN-90 inertial navigation system to resolve the problem. Air tests are planned for late November using a Cessna 402 to evaluate the system.

Geological Survey of Canada Projects 860065, 860066, 860068, 860071, 860072

A reconnaissance geotraverse through southwestern Newfoundland[1]

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Abstract

Polyphase deformed gneisses and schists of the Port aux Basques Complex (PaBC) occur between the Cape Ray Fault Zone and Rose Blanche. Three distinct divisions of the complex are proposed based on lithology and separated by shear zones. These are the Grand Bay, Port aux Basques and Otter Bay divisions. Rocks of the complex were affected by four generations of folds and a complex history of shearing, localized in narrow fault zones. D2 structures and syntectonic high temperature and pressure metamorphism were probably produced during orogenic thickening as a result of Silurian collision. The D3 and D4 structures are tentatively attributed to respectively sinistral and dextral transpression, which was manifested by sinistral and dextral strike-slip and (or) reverse movement along the major fault zones.

Résumé

Les gneiss et les schistes déformés au cours de plusieurs phases du complexe de Port aux Basques (CPaB) sont situés entre la zone de failles de Cape Ray et Rose-Blanche. Trois divisions distinctes du complexe sont proposées, basées sur la lithologie et séparées par des zones de cisaillement. Il s'agit des divisions de Grand Bay, de Port aux Basques et d'Otter Bay. Les roches du complexe ont été affectées par quatre générations de plis et des cisaillements complexes dans des zones faillées étroites. Les structures D_2 et le métamorphisme syntectonique à température et pression élevées ont probablement été produits durant l'épaississement orogénique causé par la collision silurienne. Les structures D_3 et D_4 ont été temporairement attribuées à une transpression respectivement senestre et dextre, qui s'est manifestée par un coulissage senestre et dextre ou un déplacement inverse le long des principales zones de failles, ou les deux.

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INTRODUCTION

The lower Paleozoic rocks in southwestern Newfoundland are among the least understood in the Canadian Appalachians. They comprise orthogneisses of the Cape Ray Igneous Complex, Windsor Point Group volcanic and sedimentary rocks, the enigmatic Port aux Basques gneisses (Brown, 1976, 1977) and metasedimentary and volcanic rocks of the Harbour Le Cou- and Bay du Nord groups (Brown, 1976; Chorlton, 1983). Intrusive granitoid bodies ranging in age from Ordovician to Devonian (Dunning et al. 1990) occur in the area, which is cut by several northeast trending shear zones.

The high grade metamorphism and complex deformation that affected most rocks in this region renders determination of the tectonic setting and tectonic-stratigraphic correlations with other parts of Newfoundland problematic. This paper reports on our early progress in understanding the geology and structure of this region.

CAPE RAY IGNEOUS COMPLEX

The Cape Ray Igneous Complex (CRIC) in southwestern Newfoundland (Fig. 1) comprises mainly orthogneisses of tonalitic-granodioritic composition, generally regarded as a deep remnant of the Taconic magmatic arc. These gneisses enclose large bodies of mafic-ultramafic plutonic rocks (Long Range mafic-ultramafic Complex; Brown, 1977; Dunning and Chorlton, 1985), which are supposed to represent either Early Ordovician ophiolitic remnants that survived the superimposed magmatism associated with the formation of the Middle Ordovician Taconic magmatic arc (Dunning et al., 1989) or the remnants of a large ophiolitic klippe (Brown, 1977) The regional deformation and metamorphism of these rocks was pre-Silurian as early Silurian mafic intrusions are post-metamorphic (Currie and van Berkel, 1989).

Our observations largely corroborate the findings of the earlier workers, although we failed to find evidence for the ophiolitic klippe model of Brown (1977). We noted that the mafic rocks frequently exhibit rhythmic compositional layering and that they locally appeared to intrude biotite-bearing para gneisses. We are not therefore convinced that the Long Range mafic-ultramafic Complex necessarily represents large ophiolitic remnants, rather than the relicts of arc-related layered mafic-ultramafic intrusions.

WINDSOR POINT GROUP

A long narrow belt of highly strained and variably mylonitized metasedimentary and volcanic rocks of the Windsor Point Group separate the Cape Ray Igneous Complex from the Port aux Basques gneisses (Fig. 1) and marks the Cape Ray Fault Zone (CRFZ) (Chorlton, 1983; Wilton, 1983; Dubé et al., 1991). Composed mainly of a volcanic-volcaniclastic assemblage, these rocks were assigned a Devonian age by a tenous correlation with plant-fossil-bearing strata in the Billiards Brook Formation in the La Poile River area (Chorlton, 1980). However, as demonstrated to us by B. Dubé, the Windsor Point Group also contains polymictic conglomerates with igneous clasts, limestones, rhyolitic tuffs in channel scours, and, on Jerret Point (Fig. 1), strongly deformed tholeiitic pillow basalts, indicating that at least part of the Windsor Point Group was deposited subaqueously. Prevailing biotite-grade greenschist facies metamorphism (Brown, 1977 and Chorlton, 1983), a polyphase structural history (Wilton, 1983; Dubé et al., 1991) and a total absence of fossil bearing-strata cast doubt on its assumed Devonian age. The very high strain in parts of the Windsor Point Group does not preclude the possibility that some or all of its assembled units were originally unrelated and juxtaposed tectonically.

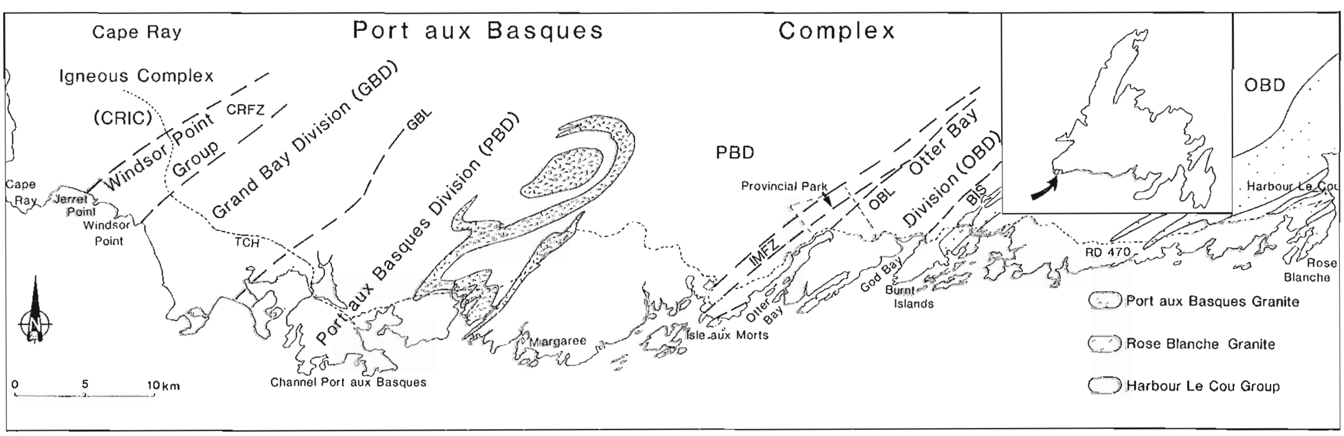

Figure 1. Map of southwestern Newfoundland with the distribution of the rock units and location of structures discussed in this paper. Geology modified after Brown (1976, 1977).

PORT AUX BASQUES COMPLEX

The Port aux Basques gneiss was originally defined by Brown (1976, 1977) as a unit of psammitic gneisses and pelitic schists with variable amounts of amphibolite that extended from the Cape Ray Fault Zone east to Harbour Le Cou (Fig. 1). Our preliminary work shows that the Port aux Basques gneiss (sensu Brown) can be divided into three major lithological units, separated by shear zones. From west to east we have named these units the Grand Bay Division, the Port aux Basques Division and the Otter Bay Division (Fig. 1). Thus, we rename the Port aux Basques Gneiss the Port aux Basques Complex (PaBC)

Grand Bay Division

The Grand Bay Division (GBD) mainly consists of thick bedded pink to grey feldspathic sandstones, converted by amphibolite-facies metamorphism into garnet bearing biotite gneiss. They are interlayered locally with amphibolite and gneissic granite, but neither of these rock-types are widely distributed. The psammite beds locally exhibit grading, with thin biotite- and muscovite-rich semipelitic schists at the top of some units. True pelites are rare or absent. In the western part of the division lensoid ultramafic bodies occur in mylonitized psammites, which are locally carbonatizised and mineralized with magnetite, pyrite and chalcopyrite. The contact with the Windsor Point Group to the west is marked by a buff-weathered mylonite zone.

Port aux Basques Division

The Port aux Basques Division (PBD) mainly consists of interstratified psammitic, semipelitic and pelitic paragneisses and schists, interlayered with gneissic granite (Port aux Basques Granite) and abundant amphibolite sheets. It includes locally also some ultramafic bodies. The Port aux

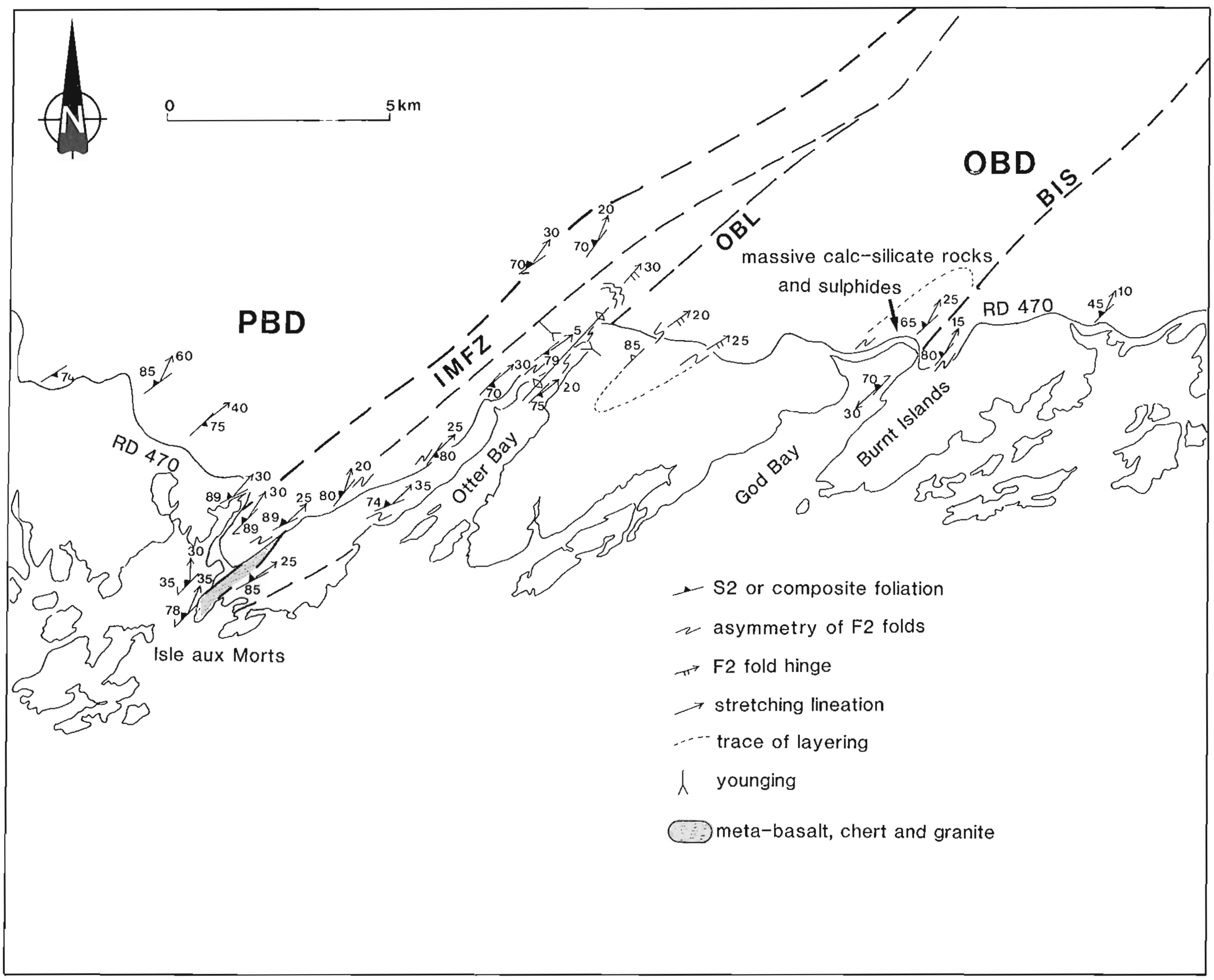

Figure 2. Structural map of the area near the Isle aux Morts Fault Zone (IMFZ) and Burnt Islands Shear (BIS). Faults and lineaments are indicated by heavy dashed lines. See also legend of Fig. 5 for meaning of shear symbols.

Basques Division is intruded by several generations of later pegmatites and granitoid dykes. This division is separated from the Grand Bay Division by the Grand Bay Lineament (GBL) (Fig. 1), which is thought to represent a major shear zone, as the strain in the rocks of the Port aux Basques Division increases markedly on approaching this lineament.

The psammites, which are generally light grey rocks, dominated by quartz, rather than feldspar, preserve original compositional grading in a few places. Biotite and red garnet are commonly present, while rare, thin, red layers of garnet-quartz concentrations (coticule) were also shown to us by M. Piasecki. Metamorphic grade increases towards the southeast where sillimanite is prevalent instead of kyanite and migmatization is abundant (Burgess et al., 1992).

The amphibolites intercalated with the psammites and pelites are generally parallel to the gneissic or compositional layering and have experienced the same degree of polyphase deformation as the metasedimentary rocks.They thus form an integral part of the rock package. Marked variation in plunge between F2 folds in amphibolite sheets and associated paragneisses at the same outcrop combined with occasional preserved crosscutting intrusive contacts show that some of the amphibolites represent mafic dykes. Preliminary geochemical results suggest a tholeiitic composition for the amphibolites. The usual parallelism of the dykes to the layering results from strain associated with the two phases of tight to isoclinal folding and subsequent strains associated with shear zone development (see below). While these intrusive amphibolites tend to be internally homogeneous (except for garnetiferous margins), other amphibolites comprise irregular bodies displaying internal compositional layering. The latter are generally characterized by numerous complexly folded epidosite knots and veins. While hornblende is present in all these amphibolites, the latter vary from pure hornblendites, to pale feldspathic rocks containing less than 15 % randomly oriented hornblende blades. These banded amphibolites are crosscut by intrusive amphibolites. We believe that the mineralogical variations reflect original compositional differences and tentatively interpret these banded amphibolites as metavolcanic tuffs and/or thin flows. Initial geochemical results suggest that the banded amphibolites have basaltic compositions.

Otter Bay Division

The Otter Bay division (OBD) was recognized as distinct from the Port aux Basques by Piasecki (unpublished contract report), who included it within the Harbour Le Cou Group (sensu Brown, 1976). We consider that these rocks are different from both the Port aux Basques Division and the Harbour Le Cou Group and so have renamed them as the Otter Bay Division. The Otter Bay Division is separated from the Port aux Basques Division by the Isle aux Morts Fault Zone (IMFZ) (Fig. 1,2), and consists mainly of thin- to thick-bedded, locally graded, light grey feldspathic psammites interstratified with thin beds of pelite (Fig. 3). The pelites locally contain sulphides, particularly in shear zones, where they weather with a rusty stain. Within the Otter Bay Division several stratigraphic units are recognizable by their distinctive lithologies. They include:

1. Clear grey psammites with subordinate sillimanite-bearing pelites and abundant garnet-clinopyroxene-anorthite-clinoamphibole-bearing calc-silicate pods and lenses (Fig. 3). Rare tholeiitic amphibolites exhibiting some HFSE depletion are present. Igneous textures preserved in low strain pods indicate that they represent metagabbro or diabase. This unit occurs immediately east of Isle aux Morts and around the head of Otter Bay.
2. Plagioclase-hornblende gneisses associated with well-developed calc-silicate pods, which are tentatively interpreted as para-amphibolites derived by the metamorphism of limy mudstone. These distinctive rocks alternate with psammitic gneisses and sillimanite-bearing pelites and occur around the nose of a large F2 fold at the head of God Bay (Fig. 2). They enclose a semi-massive pyrrhotite-pyrite body exposed in the roadcut near the entrance to the village of Burnt Islands.

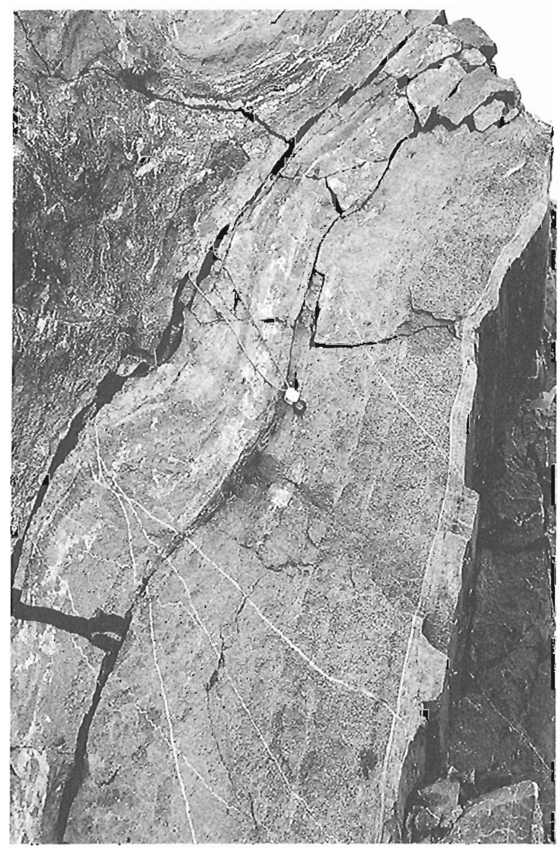

Figure 3. Graded psammite bed of the Otter Bay Division. Younging is towards the right. White pods in the psammite bed consist of calc-silicate. GSC 1991-567 O

3. Garnetiferous psammites with pink beds of coticule and purplish sillimanite-garnet-bearing pelites, generally devoid of calc-silicates. These rocks occur immediately east of the village of Burnt Islands.
4. Highly strained tholeiitic pillow basalts, with recrystallized interpillow chert occur in the village of Isle aux Morts. These rocks have not previously been described, but are locally up to 150 metres thick. Entirely surrounded by minor shear zones of the Isle aux Morts Fault Zone, their relationship with the Port aux Basques Division and rest of the Otter Bay Division is not clear at present.

The Otter Bay Division was defined by Brown (1976) as reworked and retrogressed Port aux Basques gneiss, but the scarcity of amphibolite, the abundance of calc-silicates and apparent difference of psammite composition suggest instead that the Otter Bay Division is a lithological division different from the Port aux Basques Division. The persistence of sillimanite, garnet and biotite and the presence of granitic mobilizates up to near the contact with the Harbour Le Cou Group indicates that there is no significant difference in metamorphic grade between the Otter Bay Division and Port aux Basques Division.

HARBOUR LE COU GROUP

The Harbour Le Cou Group of Brown (1976) occurs in the easternmost part of the study area around Rose Blanche and Harbour Le Cou (Fig. 1). It comprises low amphibolite-facies lustrous silvery fissile pelitic to semipelitic schists with subordinate feldspathic psammite locally containing black (volcanic?) quartz grains. It also contains fuchsite-rich layers, which have not been observed in the nearby Otter Bay Division. The contacts between the Harbour Le Cou Group and the Otter Bay Division are obscured because both rock suites have been invaded by extensive light grey garnet-bearing two-mica granites, referred to as the Rose Blanche Granite by Brown (1976). The Harbour Le Cou Group probably represents a high grade correlative of the Middle Ordovician (Dunning et al., 1990) Bay du Nord Group (Chorlton, 1983).

DEFORMATION AND METAMORPHISM

The deformation histories of the various rock units in southwestern Newfoundland are complex, and in part are probably unrelated. For instance peak metamorphism and the main deformation in the Cape Ray Igneous Complex took place in the Ordovician and predated peak metamorphism and deformation in the Port aux Basques Complex, which was in the Silurian (Dunning et al., 1990). Described here are the results of our work in the the Port aux Basques Complex along the coast, the Trans Canada Highway (TCH) and road 470 (RD 470).

Small-scale structures

The earliest deformation (D1) is represented by isoclinal folds (F1) in compositional layering (=bedding) and an axial plane schistosity (S1) defined by biotite and muscovite. Locally there is a bedding parallel foliation that is folded into highly non-cylindrical F1 folds (Fig. 4 A,B).

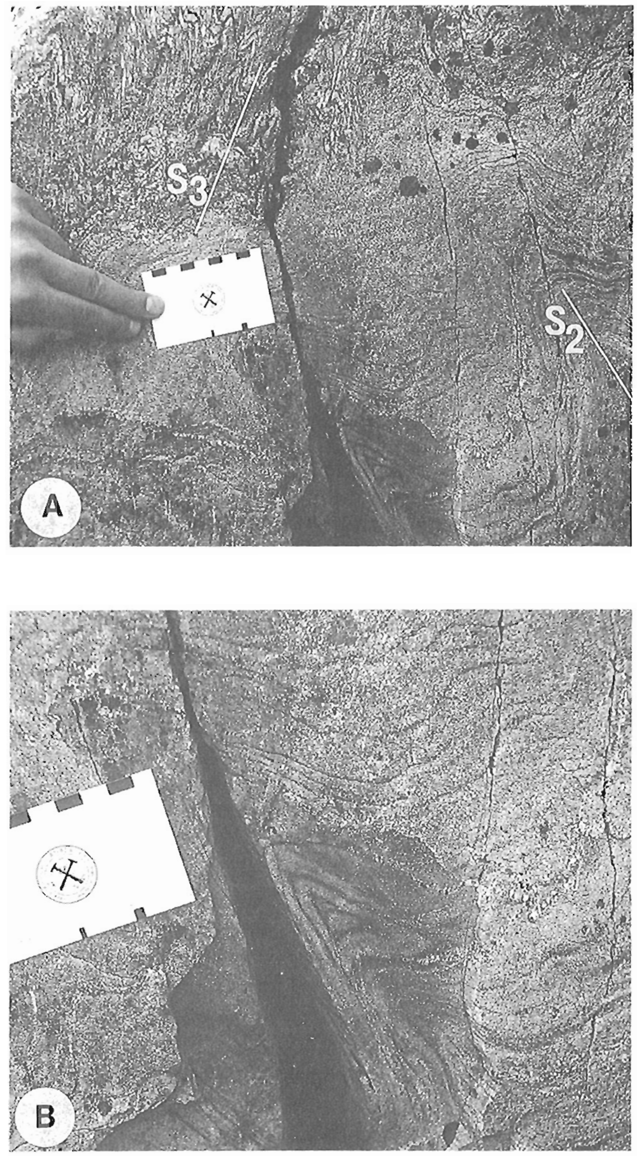

Figure 4. (A) Overprinting relationships between F1, F2 and F3 folds in psammites and pelites of the Port aux Basques Division exposed in a vertical section. Note that the F3 folds are not well developed in the psammites. GSC 1991-567 G (B) Close up of F1-F2 mushroom interference pattern of (A). The hinge line of the F2 folds is near horizontal suggesting that the F1 fold had a sheath-like structure prior to F2 refolding.

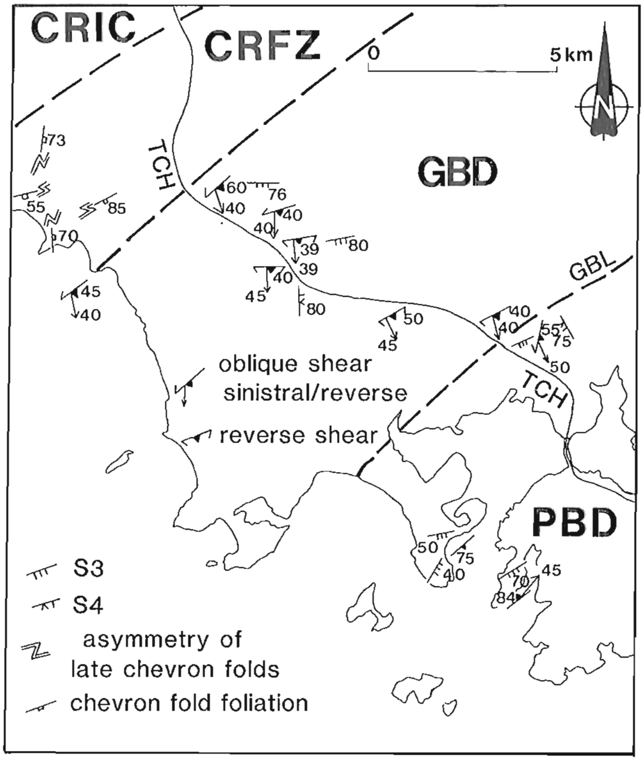

Figure 5. Structural map of the region between the Cape Ray Fault Zone (CRFZ) and the Grand Bay lineament (GBL). See also legend of Figure 2 for meaning of structural symbols.

The D1 structures are consistently overprinted by D2 structures (Fig. 5). F2 folds comprise upright to recumbent, tight to isoclinal folds (F2) and a differentiated crenulation cleavage (S2), which is generally transformed into a schistosity by advanced transposition (Williams et al., 1977). The dominant schistosity is therefore generally a composite foliation of S0,S1 and S2. F1-F2 fold interference patterns (Fig. 4) vary between coaxial refolding and mushroom types. The F2 hinge lines are commonly (sub)parallel to a strong stretching and/or mineral lineation in S2 in the high-strain zones in the Port aux Basques and Grand Bay divisions. The strength of the stretching lineation increases on approaching the major shear zones, which locally are also characterized by markedly non-cylindrical F2 folds. These non-cylindrical F2 folds generally fold the stretching lineation that elsewhere lies in S2. Such structures are typical of progressive deformation associated with shear zones.

D3 comprises upright to moderately inclined, shallowly-plunging chevron folds (F3), kinkbands or small crenulations. The accompanying axial plane cleavage (S3) generally has a northeast to east-northeast trend and a variable dip in a southerly or southeasterly direction (Fig. 5). The variable dip of S3 may in part be the result of folding by upright to tight, northerly trending F4 folds (Fig. 5), although the evidence for small scale F4 folding is scarce and in most places appears to be absent.

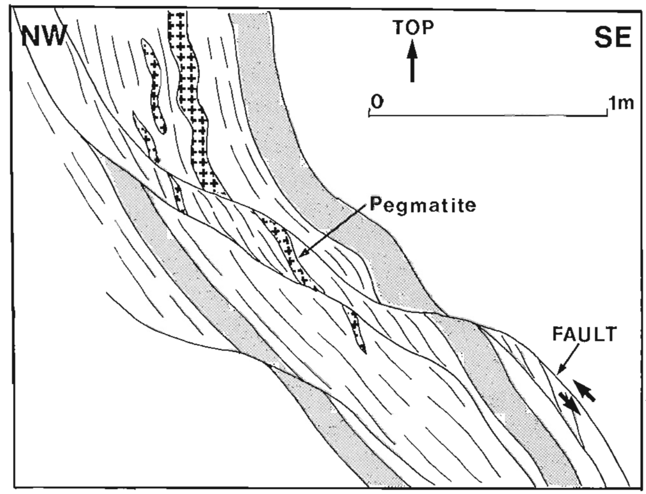

Figure 6. Field sketch of brittle-ductile faults in the Port aux Basques Division exposed in a vertical section.

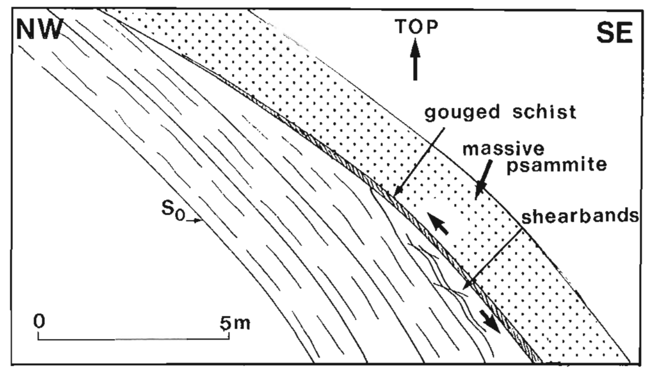

Figure 7. Field sketch of a thrust ramp in rocks of the Grand Bay Division close to the Grand Bay lineament.

Numerous small scale shear zones and ductile/brittle faults have been observed in the superb outcrops and quarries along the highway between Cape Ray and Harbour Le Cou. These shear zones generally dip steeply to shallowly to the south or southeast and are commonly localized in phyllonites or gouged schist layers, although they locally ramp through competent lithologies as discrete brittle faults (Fig. 6,7). Drag, cut-offs, shearbands, and back rotated pinch and swells (Fig. 7,8,9) generally indicate a sense of reverse movement or overthrusting to the north or northwest but some of these faults in the Otter Bay Division indicate tectonic transport towards the southeast. Subhorizontal transcurrent movements have been observed in some shear zones close to Isle aux Morts and in the Otter Bay Division. These shear zones affect F2 structures (Fig. 10) and are thus late syn- or post-D2.

Deformation and its relationship to metamorphism

D1 and D2 are accompanied by high grade metamorphism, of which at least the syn-D2 kyanite-sillimanite, clockwise pressure-temperature path is attributed to crustal thickening during the Silurian (Burgess et al., 1992). We have at present no adequate information concerning the kinematics of D2 to test this inference. However, we wish to emphasize that F3 and F4 folds generally have shallow plunges and steeply dipping axial surfaces suggesting that S2, which represents the main anisotropy in the rocks, had a shallow dip prior to F3/F4 folding and may have been the product of overthrusting typical of collision zones.

LARGE SCALE SHEAR ZONES - GENERAL CHARACTERISTICS

The Port aux Basque Complex contains numerous narrow high- strain zones, of which several mark the boundaries between the various lithological units. The high strain zones are characterized by straight gneiss (Hanmer, 1988), mylonites and/or phyllonites, which contain generally well developed stretching and mineral lineations and monoclinic fabric elements. They are therefore interpreted as ductile shear zones. The hinge lines of folds that formed before or during the formation of the shear zones are generally parallel to the stretching lineations, which combined with the presence of rare sheath folds suggest that these shear zones record relatively high finite strains and represent zones of non-coaxial deformation (Hobbs et al., 1976, p. 283-288). Hence, the transport direction is assumed to coincide with the orientation of the stretching and mineral lineations. The stretching lineations are generally defined by the long dimension of strongly segmented minerals (e.g garnet and sillimanite), boudins, pressure shadows, cigar-shaped sillimanite nodules and ribbons of streaked-out quartz and feldspar aggregates.

Isle aux Morts Fault Zone

This fault zone separates the Port aux Basques and Otter Bay divisions (Fig. 2) and appears to have the character of a major tectonic boundary as suggested earlier by M. Piasecki (unpublished contract report), since the Port aux Basques and Otter Bay divisions represent two distinctly different packages of rocks. The western boundary of the Isle aux Morts Fault Zone is marked by a steeply northwest-dipping, partly retrograded, rusty stained garnet, sillimanite, biotite and muscovite-bearing phyllonite .The rusty stain is mainly due to the oxidation of abundant pyrite.The eastern boundary is not well defined but is tentatively placed along the eastern extension of a lens of meta pillow basalt, chert and semi-pelitic schist, intruded by narrow granite sheets (Fig. 11,12), in the centre of the village of Isle aux Morts. Rocks in the Isle aux Morts Fault Zone generally have moderately to shallowly northeast-plunging stretching lineations, coaxial

Figure 8. Straight gneiss developed in rocks of the Port aux Basques Division near the Grand Bay lineament. Note layer-parallel fault and cut-off above person. GSC 1991-567 U

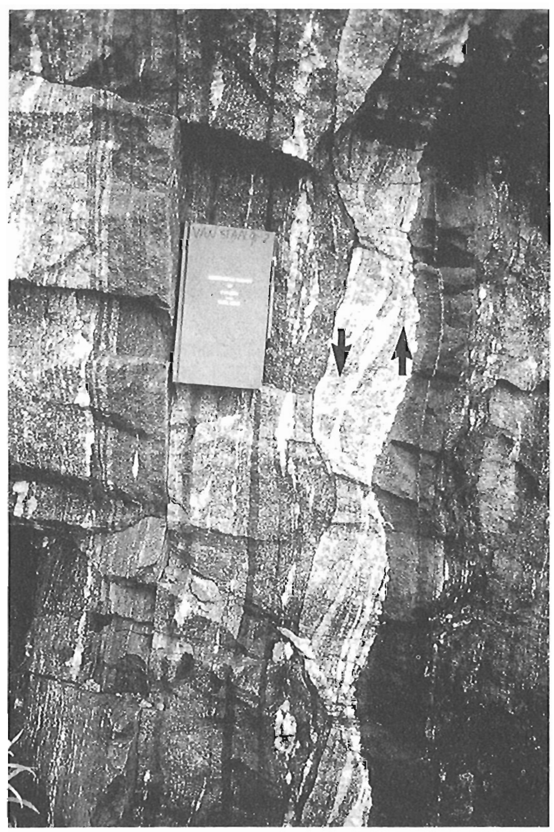

Figure 9. Back-rotated pinch and swell in foliated aplite in the Port aux Basques Division. Layering dips steeply towards the southeast (right) and arrows indicate sense of shear in country rocks.

with F2-hinge lines. The Isle aux Morts Fault Zone is clearly visible as a series of narrowly spaced topographic lineaments on the airphotos and has been mapped up to an unnamed river in the northwestern part of the Otter Bay Provincial Park. Farther to the northeast the Isle aux Morts Zone appears to converge with the Otter Bay lineament (OBL). It is therefore possible that the Otter Bay Lineament is the real eastern boundary of the Isle aux Morts Fault Zone. If correct, the Isle aux Morts Fault Zone encloses a large, shallowly northeast-plunging, upwards facing F2 antiform (Otter Bay Antiform) that is truncated on both limbs by shear zones (Fig. 2). Small scale examples of such F2 fold limb truncations are common in the Isle aux Morts Fault Zone (Fig.10) and other northeast-trending, steeply northwest-dipping fault zones in the Otter Bay Division, like the Burnt Islands Shear (BIS). They are due to the development of bedding-parallel ductile faults in the foldlimbs. The Otter Bay Antiform is partly exposed on the northwestern side of Otter Bay and indicated by a change in asymmetry of small scale F2 folds and facing direction.

Figure 10. F2 fold truncated by bedding-parallel fault in the Isle aux Morts Fault Zone. GSC 1991-567 K

The movement picture of the Isle aux Morts Fault Zone appears complex and needs further investigation. However, we have recorded evidence for both sinistral and dextral strike-slip. Drag, S-C fabrics and shearbands in partly mylonitized granite sheets, amphibolites and semi pelitic schists (Fig. 12) along the eastern side of the Isle aux Morts Fault Zone suggest mainly sinistral strike slip. These granite sheets intruded during the sinistral shearing (see below). The sinistral strike-slip thus took place late-syn or post D2, since the shears truncate F2 fold hinges . No significant retrogression has been observed in the zones affected by sinistral shearing. However, the narrow zone of rusty phyllonites along the western boundary of the Isle aux Morts Fault Zone shows retrogression of sillimanite and biotite to muscovite. These retrogressed phyllonites are accompanied by abundant shearbands indicating dextral strike-slip (Fig. 2) and tentatively have been interpreted to represent a younger phase of movement than the sinistral shearing.

Burnt Islands Shear

The Burnt Island Shear (BIS) (Fig. 2) forms a marked lineament and resembles in many respects the Isle aux Morts Fault Zone, although it is more narrow (50-75 m thick) and appears to have a less complicated movement history. The Burnt Island Shear follows the eastern margin of the God Bay and part of Burnt Islands Brook and has a steep northwesterly

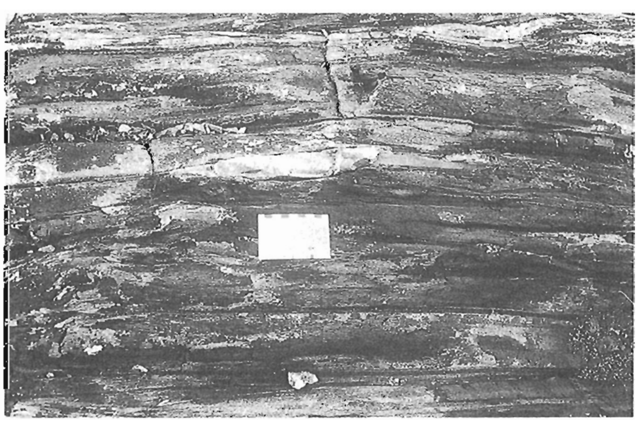

Figure 11. Strongly deformed pillow basalts with flattened inter-pillow chert, Isle aux Morts Fault Zone. GSC 1991-567 D

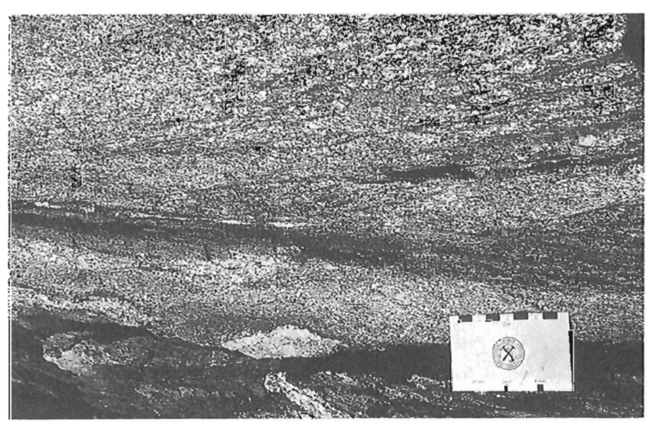

Figure 12. Sinistral S-C fabric or shearbands in syntectonic granite sheet that intruded amphibolite in the Isle aux Morts Fault Zone. GSC 1991-567 N

dip. Stretching lineations are moderately to shallowly plunging towards the northeast and parallel to F2-hinge lines. Rare shearbands suggest dextral strike-slip. The Burnt Island Shear appears to truncate a large F2 fold, the nose of which is partly exposed around the head of God Bay and contains a large pod of calc-silicate and para-amphibolite. The rocks on the eastern side of the Burnt Island Shear are represented by pelites and coticules but not calc-silicates, suggesting a significant dextral strike-slip offset.

Grand Bay Lineament

The Grand Bay lineament (GBL) coincides with the inferred position of the contact between the Port aux Basques and Grand Bay divisions (Fig. 5). Exposures of this contact have not been discovered (yet) but the strain markedly increases in the Port aux Basques Division on approaching this lineament. The Grand Bay lineament is therefore also interpreted as a major shear zone. The closest exposures to the Grand Bay lineament are found in the Port aux Basques Division along the Trans Canada Highway opposite the RCMP station. Rocks of the PBD have been transformed here into a straight gneiss, L-S tectonite with well developed down dip to slightly oblique stretching and mineral lineations (Fig. 5, 8). Several narrow gouged schist bands represent minor layer-parallel faults. Cut-offs and shearbands in the schists (Fig. 7,8) indicate a sense of overthrusting towards the north-northwest, parallel to the stretching lineation. These small faults cut through F2 folds, which if representative of the movement along the Grand Bay lineament, suggest thrusting of the the Port aux Basques Division on top of the Grand Bay Division late-syn or post D2.

The strength of the stretching lineations decreases towards the east, away from the Grand Bay lineament and stretching lineations are locally even absent. Most rocks in the Port aux Basques Division are therefore best described as S > L tectonites. Nevertheless, post-F2, small scale brittle-ductile reverse or thrust faults (Fig. 6) with a tectonic transport towards the northwest occur commonly in the western and central part of the Port aux Basques Division and are thought to be kinematically related to the shearing in the Grand Bay lineament. Penetrative stretching lineations reappear in the eastern part of the Port aux Basques Division, but are here parallel to those in the Isle aux Morts Fault Zone and probably kinematically related to it.

Figure 13. Sinistral chevron folds in the Windsor Point Group cut and overprinted by crenulation cleavage associated with dextral chevron folds. GSC 1991-567 F

Cape Ray Fault Zone

The strain in the Grand Bay Division increases progressively on approaching the Cape Ray Fault Zone (CRFZ). The contact between the Grand Bay Division and the Cape Ray Fault Zone is marked by a complicated zone of mylonites and ultramylonites, the protolith of which is not always apparent. Stretching lineations and kinematic indicators in the Grand Bay Division suggest a dominantly north-northwest-directed sense of overthrusting or reverse shear (Fig. 5), which is probably related to the reverse-sinistral movement documented by M. Piasecki (unpublished contract report) and Dubé et al. (1991) in the Cape Ray Fault Zone. The increase in strain is locally associated with significant retrogression and hydrothermal alteration, suggesting that the shearing is at least in part post-peak metamorphism in the Grand Bay Division.

Dubé et al. (1991) noted oblique reverse to subhorizontal movements, the latter associated with both dextral and sinistral shear sense indicators, and associated folding The

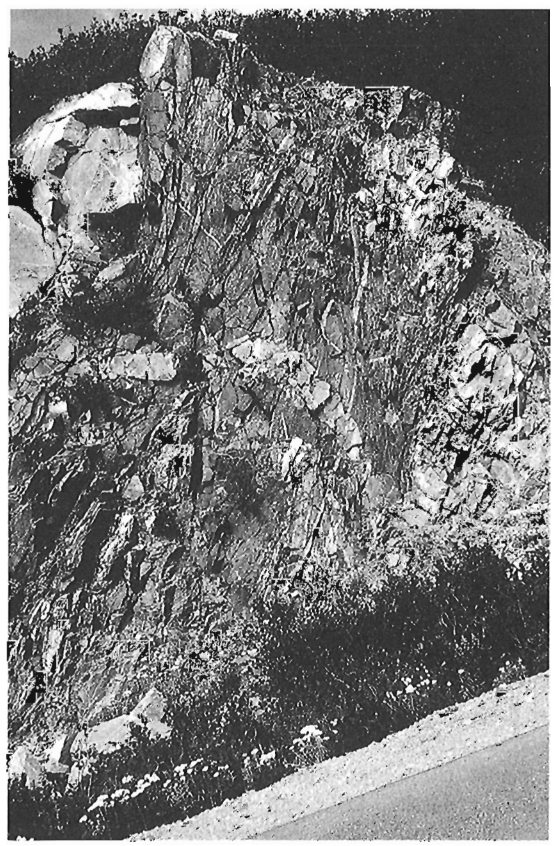

Figure 14. Weakly F2 folded apophyse of Rose Blanche Granite. Hammer for scale GSC 1991-567 R

latest movements and the north to northeast trending Z-shaped chevron folds and crenulations were interpreted as related to a dextral transpressive regime. However, we observed that the S-shaped folds are consistently overprinted by the Z-shaped folds (Fig. 5,13) in the Cape Ray Fault Zone along the coast (shown to us by B. Dubé) suggesting that these two foldstructures are due to two different, possibly unrelated strain increments. The S- and Z-shaped folds correspond in style and orientation respectively to the F3 and rare F4 folds in the Port aux Basques and Grand Bay divisions outside the fault zone and are tentatively correlated. Although more work is needed, we think that these folds record a change from a sinistral to a dextral transpressive regime.

STRUCTURAL AGES OF GRANITE INTRUSION

The Port aux Basques granite in the Port aux Basques Division (Fig. 1) has been deformed during D1 since it contains locally a well developed S1 but it cannot be ascertained whether the granite intruded before or during D1.

Near the village of Margaree in the Port aux Basques Division, pegmatites intruded along extensional shear bands and were also themselves deformed during the shearing, suggesting intrusion during sinistral shear. These pegmatites resemble the granite sheets deformed by sinistral shear in the Isle aux Morts Fault Zone, which also intruded during sinistral shear since they cut mylonitic fabrics developed in the surrounding supracrustal rocks.

The Rose Blanche granite sheets in the Otter Bay Division (Brown, 1976) show variable degrees of deformation. Evidence for the relative age of intrusion is well preserved in the the western sheet, which contains leucocratic garnet-bearing two-mica granite phases, east of Burnt Islands. The regional S2 schistosity is locally well developed in the granite, particularly close to the contacts with the country rock of the Otter Bay Division. Furthermore, the granite cuts the S1 foliation in the meta sediments and apophyses of the granite into the country rock are generally weakly folded during F2 (Fig. 14) suggesting intrusion during D2.

SUMMARY AND CONCLUSIONS

The results of our preliminary mapping shows that the Port aux Basques Complex can be subdivided into at least three units, of which the Grand Bay and Port aux Basques divisions contain a large percentage of mafic and ultramafic units. The latter are rare or absent in the Otter Bay Division. The Otter Bay Division is also characterized by a relative abundance of distinct calc-silicate rocks, which are rare or absent in the Grand Bay and Port aux Basques divisions. The Isle aux Morts Fault Zone may therefore represent a significant tectonic boundary, although the tectonostratigraphic affinities of the Grand Bay, Port aux Basques, and otter Bay divisions are not well understood at present. The Harbour Le Cou Group differs from the Otter Bay Division, contrary to the interpretation of Piasecki (1990), in that it contains fuchsite-rich layers, which are absent in the Otter Bay Division, and lacks the thick psammite beds and calc-silicate rocks typical of the Otter Bay Division. We agree with earlier workers that the Harbour Le Cou Group appears to represent a higher grade equivalent of the Middle Ordovician Bay du Nord Group.

The deformation history comprises four generations of folds. The kinematics of the first two generations of structures are not well understood at present, although D2 was accompanied by a clockwise P-T path and probably the product of the Silurian collision between the Avalon composite terrane and Laurentia (van Staal and Fyffe, 1991) The F3 and F4 folding are tentatively attributed to transpression-related deformation. Our preliminary data indicate that the sinistral and reverse shears predate the dextral shearing. The relationships between D1 and D2, and between D2 and younger deformation is at present unclear. The observed structures are therefore not necessarily all due to one progressive deformation.

ACKNOWLEDGMENTS

We benefited from discussions with Benoit Dubé, Steve Colman-Sadd, Ken Currie and Mark Piasecki who generously shared their knowledge of this area with us. Benoit Dubé gave us a fantastic fieldtrip in the Cape Ray Fault Zone, which helped considerably to improve our understanding of this part of Newfoundland. Ken Currie is thanked for critically reading this manuscript.

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Geological Survey of Canada Project 730044

Preliminary report on the metamorphic geology of the Port aux Basques Complex, southwestern Newfoundland[1]

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Abstract

This report summarizes initial results of a project concerned with unravelling the pressure-temperature-time-deformation history of the Port aux Basques Complex (PaBC). The Complex occurs between the Cape Ray Fault Zone and Rose Blanche, representing part of the Central Mobile Belt, and comprises the Grand Bay, Port aux Basques and Otter Bay divisions, separated by the Grand Bay Lineament and the Isle aux Morts Fault Zone, respectively. Rocks of all three divisions have been affected by polyphase deformation under variable metamorphic conditions. Structural analysis of the Port aux Basques Division reveals four phases of deformation, the first two of which occurred during high-grade metamorphism resulting in multiple stages of porphyroblast growth and matrix recrystallization. The metamorphism, which is of kyanite-sillimanite type and included development of migmatites by anatexis, is interpreted to be the result of crustal thickening as manifested in an overall clockwise pressure-temperature path.

Résumé

Le présent rapport résume les résultats préliminaires d'un projet visant à reconstituer l'évolution de la déformation selon le temps, la pression et la température du complexe de Port aux Basques (PaBC), qui se trouve entre la zone de failles de Cape Ray et Rose-Blanche; il fait donc partie de la zone mobile centrale et englobe les divisions de Grand Bay, de Port-aux-Basques et d'Otter Bay. Les deux premières divisions sont séparées par le linéament de Grand Bay tandis que les deux dernières le sont par la zone de failles d'Isle-aux-Morts. Les roches des trois divisions ont subi une déformation polyphasée dans des conditions métamorphiques variables. L'analyse structurale de la division de Port-aux-Basques met en lumière quatre phases de déformation, les deux premières ayant eu lieu pendant un épisode de métamorphisme intense; ce fait explique la croissance en plusieurs étapes des porphyroblastes et la recristallisation de la matrice. Le faciès métamorphique à kyanite et à sillimanite ainsi que la formation de migmatites par anataxie sont interprétés comme résultant d'un épaississement crustal, sur considération des variations dextrorsum d'ensemble de la pression et de la température.

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INTRODUCTION

Southwestern Newfoundland is an area of complex and poorly understood geology. Here, the Humber, Dunnage and Gander zones of Williams (1979) converge and tectono-stratigraphic correlations with areas to the north and south have been problematic. Rocks assigned to the Humber Zone traditionally have been separated from rocks assigned to the Dunnage Zone by the Long Range Fault. To the southeast, a definitive boundary which separates rocks of Dunnage and Gander Zone affinities has been controversial, so that the term Central Mobile Belt is preferred for this region.

The Cape Ray Fault Zone lies south of the Long Range Fault. It separates two contrasting geological domains and forms the northern boundary of the Hermitage Flexure (O'Brian et al., 1986). Northwest of the Cape Ray Fault Zone lies the Cape Ray Igneous Complex, which includes the Cape Ray Granite, the Cape Ray Tonalite, the Red Rocks Granite and associated remnants of the Long Range Mafic-Ultramafic Complex (P. Brown, 1973, 1975, 1976, 1977; Wilton, 1983, 1985). The Cape Ray Fault Zone is represented as a 1 km wide mylonite zone which extends through both the Cape Ray Igneous Complex and the Devonian Windsor Point Group, a deformed sequence of mainly volcanic and volcaniclastic rocks. Southwest of the Cape Ray Fault Zone lies the Port aux Basques Complex (PaBC) (Gillis, 1972). The Complex is composed of high-grade metasedimentary rocks of marine affinity with metabasic layers and felsic intrusive rocks. With the exception of late or post-tectonic intrusions, all of these rocks have been variably metamorphosed under at least amphibolite facies conditions.

Rocks within the Port aux Basques Complex have been interpreted several ways. P. Brown (1973) interpreted the Port aux Basques Gneiss (PaBC of this report) to represent Precambrian basement to the Gander Zone rocks. According to him, the Port aux Basques Gneiss became progressively reworked towards Harbor le Cou until the gneissic layering had become completely transposed to give to the rocks the monotonous appearance of semipelite. Three major and one minor phases of deformation were identified. P. Brown suggested that the metamorphism was intimately related to a suite of granitic sills, the Port aux Basques Granite, resulting in a prograde partial Barrovian sequence. This was overprinted by a second episode of porphyroblast growth during the third deformational event.

Wilton (1985) suggested that the Port aux Basques Gneiss (PaBC of this report) may be a sedimentary sequence underplated by Grenvillian Basement, based on the petrogenesis of granites emplaced within the Port aux Basques Gneiss some of which he argued were derived from a granulite source. M.A.J. Piasecki (unpublished report, 1990) divided the Port aux Basques Gneiss of Brown (1976, 1977) at the Isle aux Morts Fault Zone and redefined the rocks from Isle aux Morts to Harbor le Cou as Harbor le Cou Group. He inferred middle amphibolite grade for pelitic rocks near Port aux Basques, based on staurolite and kyanite bearing assemblages, and interpreted the Harbor le Cou Group to represent upper greenschist/lower amphibolite grade near Isle aux Morts, with grade decreasing to middle greenschist near Harbor le Cou. Piasecki speculated whether parts of the Port aux Basques Gneiss itself may

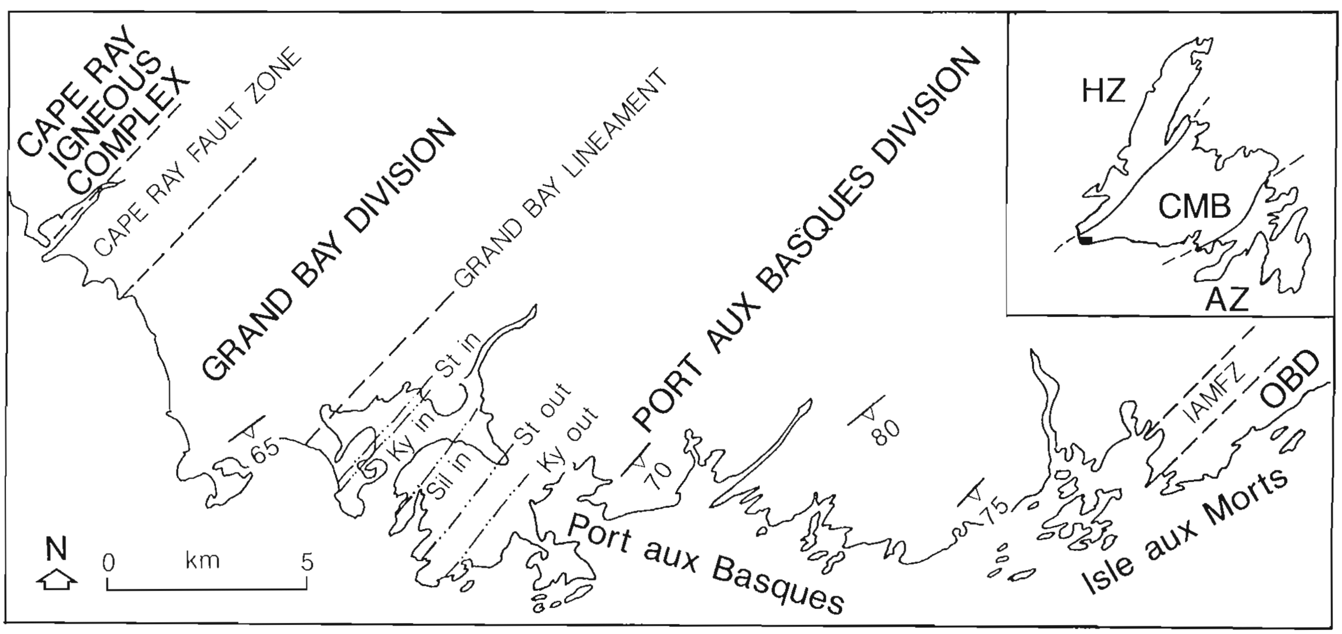

Figure 1. Outline of the geological map of the Port aux Basques region of southwestern Newfoundland. The Port aux Basques Complex includes the Grand Bay Division, Port aux Basques Division and Otter Bay Division (OBD). Dash lines represent tectonic contacts, IAMFZ = Isle aux Morts Fault Zone. Dash and double dot lines represent metamorphic isograds. Strike and dip symbols refer to the orientation of the main compositional layering. Inset shows location of the Port aux Basques region in southwestern Newfoundland, HZ = Humber Zone, CMB = Central Mobile Belt, AZ = Avalon Zone.

represent tectonic slivers of basement. Finally, it cannot be overruled that the Port aux Basques Complex is a supracrustal sequence that overlies a basement of both Gander and Dunnage Zone rocks.

A detailed structural and metamorphic study of the Port aux Basques Complex has been started to resolve its problematic status and to establish how it fits into an overall tectono-stratigraphic scheme for the Appalachian Orogen. The goal of this research is to integrate the history of deformation and metamorphism preserved in the rocks through field, petrographic, petrogenetic and geochronological studies to achieve a more thorough understanding of the pressure-temperature-time-deformation evolution of the Port aux Basques Complex. This will enable a better understanding of the tectonic evolution of the Appalachian Orogen.

Figure 2. Quartz-feldspar leucosomes and biotite-rich melanosomes developed in semi-pelitic Port aux Basques Division rocks and interpreted to represent the products of upper amphibolite facies partial melting. Location: North of RCMP Station, west of Port aux Basques.

Figure 3. Port aux Basques Division garnet-staurolite-kyanite-biotite schist. Location: Southwest of Port aux Basques, near Pole Rock.

GEOLOGICAL SETTING

At the extreme southwest of the Central Mobile Belt is the Port aux Basques Complex, a sequence of deformed sedimentary and volcanic rocks with mafic and felsic intrusive suites. Orogenesis in this part of the Central Mobile Belt has been attributed to a discrete phase of Silurian tectonothermal activity (Dunning et al. 1990). Titanite from an amphibolite in the Port aux Basques Complex yielded a post-peak cooling age of 412 ± 2 Ma (Dunning et al. 1990). It appears that tectonothermal activity of Devonian or younger age is limited to the emplacement of some plutons, exhumation and brittle faulting.

The Port aux Basques Complex is subdivided into three divisions on a lithological basis, separated by major shear zones (Fig. 1) (van Staal et al., 1992). At the northwestern edge of the study area, near the Cape Ray Fault Zone and adjacent to the Windsor Point Group, the Grand Bay Division (GBD) consists of garnetiferous psammites and semipelites with minor (<10%), metabasic rocks and occasional medium-grained granitic sheets. The rocks are at lower amphibolite grade and become progressively more flaggy and highly strained as the Cape Ray Fault Zone is approached.

To the southwest, across a steep, southeast dipping, ductile shear zone with a reverse sense of displacement to the north or northwest (van Staal et al. 1992), is the Port aux Basques Division (PBD). From approximately Grand Bay West to Isle aux Morts the Port aux Basques Division has a very distinctive and consistent compositional layering. This division is composed of dominantly felsic gneiss and associated metabasic lenses and layers, pelitic schists and rare calc-psammites. These rocks have been metamorphosed to upper amphibolite facies conditions, the highest grade in the Port aux Basques Complex.

Figure 4. Quartz-feldspar leucosomes associated with melanosome patches, developed in Port aux Basques Division amphibolitic unit of the Port aux Basques Division and interpreted to represent the products of upper amphibolite facies partial melting. Location: North of RCMP Station, west of Port aux Basques.

To the east and southeast at Isle aux Morts, another major lithologic and structural break occurs. The Isle aux Morts Fault Zone (van Staal et al., 1992) juxtaposes rocks of the Port aux Basques Division with rocks of the Otter Bay Division (OBD). The Otter Bay Division extends from Isle aux Morts, east to Rose Blanche and Harbour la Cou. Rocks in this group consist of feldspathic psammites interbedded with thin beds of semipelite which contain calc-silicate pods, and a few metabasic horizons, metamorphosed to amphibolite facies conditions.

The Port aux Basques Granite, as defined by P. Brown (1973, 1975, 1976, 1977), is a pink, commonly two mica-granitoid that intrudes the Port aux Basques Division. The granite was intruded before or during the earliest episode of ductile deformation (D_1). The Rose Blanche Granite (Brown, 1976) is a late tectonic garnet-bearing two-mica granite which intrudes the Otter Bay Division and adjacent terrains to the east. The Strawberry Hill and Isle aux Morts Brook granites are generally undeformed late syn to post tectonic granitoids that intrude both the Port aux Basques Complex and the Windsor Point Group. Wilton (1983) suggested an age of 362 ± 16 Ma (MSWD: 6.3) based upon a composite Rb/Sr whole rock isochron using samples from both granites. This age places an upper constraint on the age of deformation and metamorphism in the area.

STRUCTURAL HISTORY

We have subdivided the Port aux Basques Complex into three units, Grand Bay, Port aux Basques, and Otter Bay divisions. The deformational histories in these units are complex and may not be completely correlatable. The following describes the results of our preliminary work, concentrating on the Port aux Basques Division.

The composite gneissic layering displayed by the Port aux Basques Division is the most pronounced structural feature, it strikes northeast-southwest and dips steeply (60-85°) southeast. Polyphase deformation of the gneiss is referred to four folding events, excluding minor late brittle faulting that occurs sporadically across the area. The intensity of the earliest observed deformation largely has obliterated any previous history (pre-D_1) that may have been present, so that now original relationships within the supracrustal rocks have been transposed into a new tectono-lithologic layering.

The earliest observed deformation (D_1) involved tight to isoclinal folding (F_1) of compositional layering and the development of an axial planar schistosity, (S_1), only preserved sporadically due to the strong transposition associated with the second deformation (D_2). Evidence for D_1 is observed in numerous overprinting relationships produced as the result of the superposition of D_2 folds on D_1 structures. Fold interference patterns vary between type 2 and type 3 structures. The second deformation is intense and has resulted in the development of upright to recumbent, tight to isoclinal folds (F_2) and an associated crenulation cleavage (S_2), which largely has transposed S_1 and which is preserved throughout the study area as the dominant fabric in the rocks. F_2 hinge lines commonly are parallel to a stretching lineation defined by segmented grains or mineral alignment. An episode of intensive ductile shearing occurred late-syn- or post-D_2 which resulted in the Port aux Basques Division being juxtaposed with the Grand Bay Division. Cut-offs along F_2 fold limbs, shear bands and asymmetric boudinage are all consistent with an overthrusting motion to the north or northwest (van Staal et al., 1992). The third deformation (D_3) has developed kink or chevron style folds (F_3) that overprint the earlier structures. An axial planar crenulation cleavage (S_3), with a northwest trend and a moderate dip, is associated with these folds and is of variable intensity across the area. Evidence for a fourth deformation (D_4) is sporadic. D_4 has produced variably upright, plunging, open to tight folds and an associated crenulation cleavage that dies out away from the Cape Ray Fault Zone.

PORT AUX BASQUES COMPLEX

Grand Bay Division

The Grand Bay Division lies structurally beneath the Port aux Basques Division and above the Windsor Point Group. It comprises a sequence of well layered, predominantly feldspathic gneisses to semipelitic schists and gneissic granites with subordinate metabasic units and rare pelitic schists. The compositional layering within the Grand Bay Division is oriented consistently northeast-southwest and dips steeply to the southeast. Metamorphic grade for the unit as a whole is lower amphibolite facies as evidenced by hornblende in basic rocks and muscovite plus biotite in pelitic rocks.

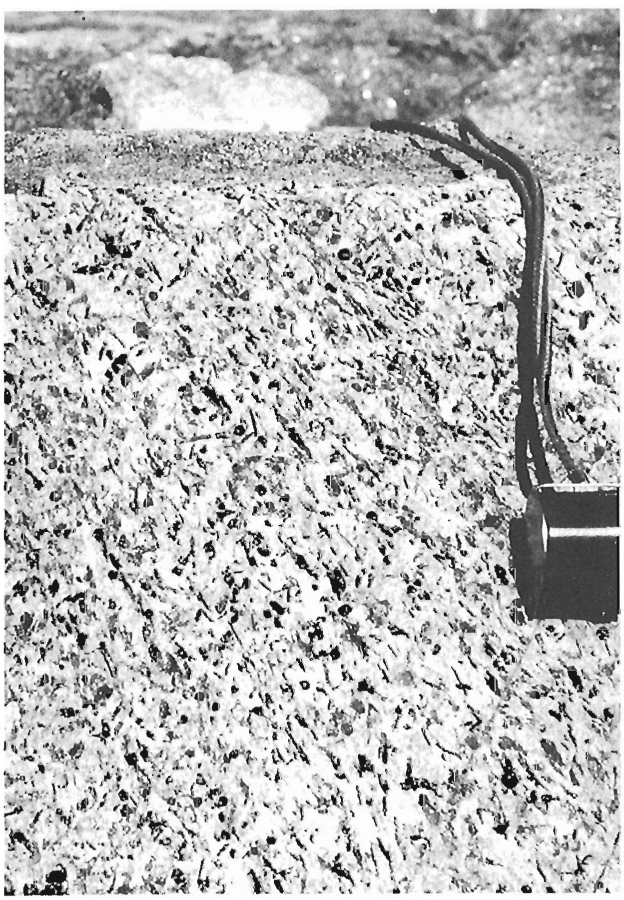

Figure 5. Port aux Basques Division calc-psammite layer showing hornblende and garnet in plagioclase-quartz matrix. Location: Southwest of Port aux Basques, near Pole Rock.

Figure 6. Sillimanite nodules parallel to S_2 in Otter Bay Division garnet-bearing biotite gneiss, east of Isle aux Morts.

Feldspathic gneiss

This unit comprises a sequence of grey-pink predominantly paragneisses with gradations to semipelite. Rocks in the unit are typically medium grained containing quartz (40-70 %), plagioclase (20-40%), biotite (10-30%), muscovite (0-5%), garnet (0-5%), and magnetite (0-2%). All varieties contain a foliation defined by biotite foliae or, more rarely, muscovite and biotite foliae. Layer thickness within the feldspathic gneiss units varies considerably from 0.5 m to several tens of metres. Some of the thicker homogeneous layers may represent orthogneisses.

Metabasic rocks

All of the metabasic rocks, which may be up to 2 m in thickness, have been strongly deformed generally to give the appearance of parallel layering with the enclosing feldspathic gneiss, but close inspection indicates that at least some of the basic rocks are discordant, and likely represent dykes. The rocks are composed of black hornblende (30-60%), biotite (20-30%), quartz and plagioclase, occasionally with subhedral dark red garnet porphyroblasts 1-5 mm in diameter and rare chrysotile, and often are quite schistose. The greenish tint characteristic of these rocks is given by minor epidote and variable amounts of Fe-rich chlorite.

Pelitic schists

In general, pelitic schists are uncommon in the Grand Bay Division and usually do not exceed 1 m in thickness. The silver-green pelites are primarily fine grained, flaggy muscovite schists with subordinate biotite (0-5%) plus plagioclase and quartz. In thin section, fine grained, aligned muscovite defines a well developed schistosity deformed by a weak crenulation cleavage. Larger green-brown biotite occurs in the plane of schistosity but also develops porphyroblasts that overprint the schistose fabric.

Figure 7. Sigmoidal inclusion tails in garnet which represent crenulations possibly of pre-S_1 fabric. A post-S_1 - pre-S_2 garnet rim now separates the discordant core inclusion trails from the external S_2 fabric. Long dimension of field of view is 2 mm. Location: Coastline of Port aux Basques.

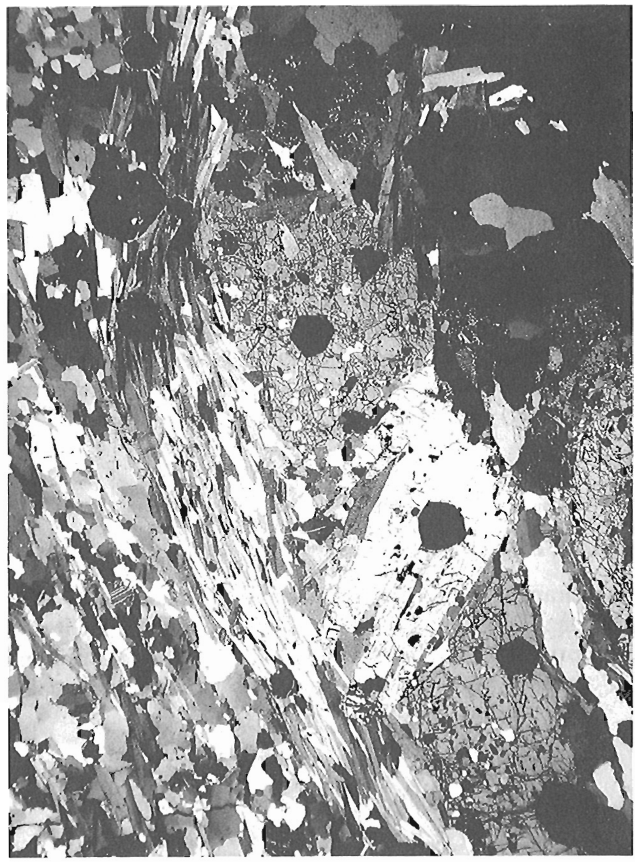

Figure 8. The S_2 schistosity is seen to wrap around one end of a large (20 mm long) single staurolite porphyroblast. The staurolite, which could be pre- or early syn-D_2, includes earlier kyanite, and both kyanite and staurolite include small, idiomorphic garnet grains. Idiomorphic garnet grains also occur in the matrix, but may be post D_2. Long dimension of field of view is 14 mm. Location: Southwest of Port aux Basques, near Pole Rock.

Port aux Basques Division

This division is an extensive group of paragneisses, orthogneisses, schists, and metabasic rocks that structurally overlies Grand Bay Division. The compositional layering within rocks of the group strikes consistently northeast-southwest and dips steeply to the southeast. The Port aux Basques Division records the most extreme metamorphic conditions in the area with upper amphibolite facies mineral assemblages and the development of mobilizates which represent both metamorphic segregations and partial melts. The unit as a whole is best recognized by the pervasive association of metabasic layers and lenses with felsic gneisses and schists (M.A.J. Piasecki, unpublished report, 1990). This interlayering between metabasic and felsic rocks is constant throughout the Port aux Basques Division and is noticeably absent in both the Grand Bay and Otter Bay divisions. The dominant lithologies are felsic gneisses, pelitic schists and metabasic rocks, with minor calc-psammites.

Felsic gneisses

The felsic gneisses, 1-3 m in thickness, encompass a wide range of rock compositions which contain variable proportions of quartz, plagioclase, biotite, muscovite, ± sillimanite, and ± garnet. The rocks are predominantly medium grained paragneisses, but coarse grained orthogneisses do occur, including highly sheared orthogneisses with large feldspar porphyroclasts near Margaree. The paragneisses with higher mica contents contain fibrolite/sillimanite typically as radiating masses seen on foliation surfaces. Leucocratic segregations are abundant, commonly with biotite selvedges (Fig. 2), and conform to the metatexites of M. Brown (1973). Although garnet is common, it is more conspicuous in the cores of some leucosomes.

Pelitic schists

Pelites are ubiquitous in the Port aux Basques Division, but the average thickness of units rarely exceeds 1 m. Geographically from west to east, the AFM mineralogy is as follows: (garnet + biotite), (garnet + staurolite + biotite), (garnet + staurolite + kyanite + biotite), (garnet + kyanite + staurolite + sillimanite + biotite), (garnet + sillimanite + staurolite + biotite), (garnet + sillimanite + biotite). Pink garnets are small (2 mm) and subhedral or larger (5 mm) grains with inclusions of quartz, plagioclase and biotite. Subhedral poilkioblastic staurolite (up to 40 mm) is a deep brown colour and may be twinned. The staurolite is smaller and amber coloured when it occurs with sillimanite. Large (10-40 mm) porphyroblasts of pale blue to green coloured

Figure 9. Staurolite with oriented quartz inclusions interpreted to represent S_1 preserved during early syn-D_2 growth of staurolite now wrapped by transposive S_2 schistosity. Location: West of junction of Trans Canada Highway and Rt. 470, Port aux Basques.

Figure 10. Relatively straight inclusion tails in the center of kyanite curve through the porphyroblast rim and are consistent with early syn-D_2 growth, a conclusion supported by the overall shape of the porphyroblast which exhibits small "tails" at the top left hand and bottom right hand corners. The dominant schistosity which encloses the kyanite porphyroblast is S_2. Long dimension of field of view is 6 mm. Location: Coastline of Port aux Basques.

kyanite are generally well developed. In thin section, kyanite may contain numerous inclusions of quartz, garnet, plagioclase and staurolite. Figure 3 illustrates a typical garnet, staurolite, kyanite and biotite assemblage. Also, kyanite can be found with quartz in veins and as a hydrothermal mineral on surfaces between lithologies. Sillimanite occurs in both the fibrolitic and prismatic form. Crystals are large (5-40 mm) and fibrolite varieties may form radiating masses partially pseudomorphed by muscovite or may form small aggregates, termed faserkiesel.

Metabasic rocks

Black to dark grey metabasic rocks, with occasional green epidosite nodules, range in thickness from a few centimetres to about 3 m. Foliation within the metabasic rocks is defined by biotite and hornblende crystals and trends generally to the northeast. The units are composed of hornblende, biotite, plagioclase, quartz, ± garnet, and ± clinopyroxene in variable modal proportions. Metamorphic segregations and mobilizates with irregular selvedges are common (Fig. 4).

Overall, the fabric in the basic rocks varies with the proportion of hornblende to plagioclase. Rocks with a high hornblende/plagioclase ratio develop a more schistose fabric. Subidioblastic hornblende can be seen to coexist with pale green clinopyroxene and garnet porphyroblasts. Garnet cores often contain oriented inclusions of quartz, plagioclase and hornblende, surrounded by an inclusion poor rim. Matrix plagioclase and quartz have a granoblastic texture. Rocks with lower hornblende/ plagioclase ratios develop a gneissic fabric. These rocks display subidioblastic hornblende, occasional poilkioblasts of garnet and an elongate quartz and plagioclase matrix. Accessories in both types include biotite, secondary chlorite, titanite, apatite and opaque grains.

A discordant relationship between the host gneiss and some of the metabasic rocks is evident from truncation of fine layers in the gneiss and markedly different plunges of hinge lines of same generation folds in gneisses and associated metabasic rocks (van Staal et al., 1992). This suggests that some of the metabasic rocks likely were originally dykes.

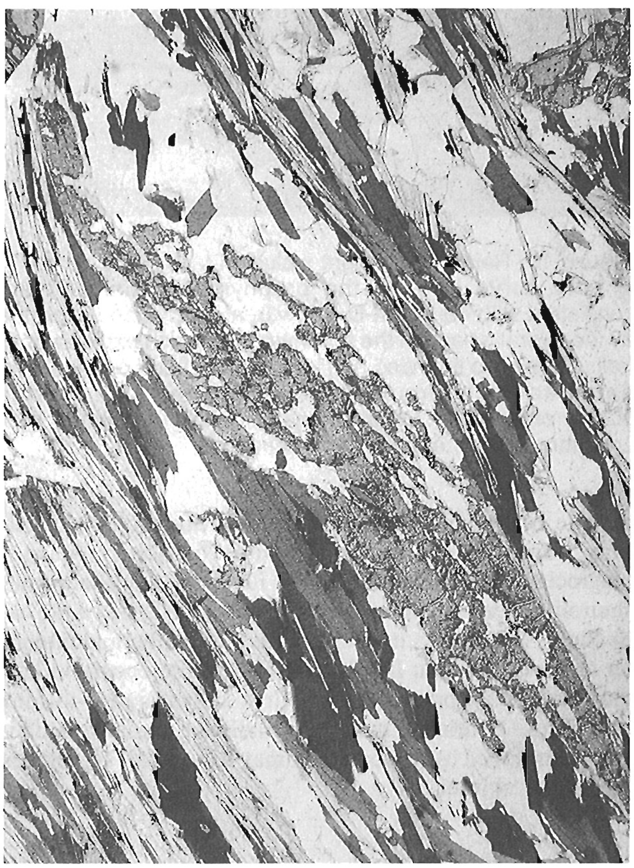

Figure 11. A kyanite porphyroblast nucleated and grown early during the D_2 folding and S_2 crenulation cleavage formation, which has been wrapped later by S_2, as the developing S_2 foliation transformed into a schistosity. The kyanite appears to have rotated during the development of S_2 and is now curved at each end by post-crystallization deformation. Long dimension of field of view is 6 mm. Location: Coastline of Port aux Basques.

However, other metabasic horizons which are not discordant are interpreted to be metavolcanic tuffs or flows (van Staal et al., 1992).

Ultrabasic pods surrounded by high strain zones occur in the vicinity of Western Island and Fox Roost. Three varieties have been identified. The most common is an essentially monomineralic rock composed of actinolite and minor talc. The second type occurs as a restricted layer approximately 10 cm thick between mylonitic gneisses. The mineralogy consists of small (1-2 mm) grains of epidote (70%) with sericitized poilkioblasts of plagioclase, quartz and minor titanite forming a granoblastic texture. The third type is an actinolite-spinel- clinopyroxene rock. The clinopyroxene occurs as large (5 mm) grains partially replaced by talc. The actinolite is medium grained and makes up the bulk of the rock. Spinel makes up 10% of the rock and is partly replaced by an opaque mineral.

Calc-psammites

Grey calc-psammite horizons up to 10 cm in thickness are a distinctive component within the Port aux Basques Division (Fig. 5). Mineralogy consists of radiating hornblende, with occasional clinopyroxene cores, and garnet set in a matrix of plagioclase and quartz. Epidote is present as small grains often with clinozoisite cores.

Otter Bay Division

The Otter Bay Division is a lithologically distinct, monotonous group of psammites, semipsammites and semipelites. Locally the units exhibit graded beds. The dominant structure is a northeast-southwest striking foliation that generally dips shallowly-to-steeply to the northwest. Metabasic rocks are sparse and appear intrusive in most cases; igneous textures found in low strain zones indicate a mafic igneous protolith (van Staal et al., 1992). The semipelitic rocks commonly contain both sillimanite and garnet. The sillimanite is present as prismatic and fibrolitic varieties, occasionally as ellipsoids within the dominant schistosity (Fig. 6). Muscovite and biotite define the schistosity, but late large (4-7 mm) muscovite grains cut across the fabric.

Calc-silicate pods are abundant in the thicker psammitic units, with diffuse, orange garnet prophyroblasts, biotite, hornblende and epidote. Highly sericitized plagioclase, quartz and titanite form the matrix.

MICROSTRUCTURES AND TIME RELATIONS OF MINERAL GROWTH IN THE PORT AUX BASQUES DIVISION

The dominant fabric in pelitic schists of the Port aux Basques Division is the pervasive S_2 schistosity, defined by medium- to coarse-grained muscovite and biotite, which is the thoroughly transposive end product of the crenulation cleavage process (Williams et al., 1977; Bell and Rubenach, 1983). In the western part of the Port aux Basques Division, porphyroblast minerals in pelites record several growth stages. Garnet wrapped by the fully developed transposive S_2 foliation, often may contain a weakly sinuous internal schistosity in the core that is oblique to the external foliation. These cores in turn are overgrown by a subidioblastic inclusion poor rim. Other garnets contain more sigmoidal inclusion trails (S_1), interpreted to be preservation of more extensively developed crenulations, which are oblique to the external S_2 fabric (Fig. 7). A post-D_1 - pre-end D_2 inclusion poor rim now separates the inclusion tails from the discordant external S_2 fabric. This suggests that S_1 in the pelites may be a crenulation cleavage of a pre-existing bedding-parallel (?) fabric. Small idioblastic garnets occur frequently within mica grains that define the S_2 fabric (Fig. 8). Periods of garnet growth are interpreted to be early, syn-D_1, between D_1 and D_2, and possibly later, although the relative age of the small idioblastic garnets is difficult to determine. Staurolite and kyanite occur together throughout most of the area and each displays multiple periods of growth. In the field staurolite, with an internal schistosity, may be seen with an external S_2 schistosity wrapped around it (Fig. 9), which suggests that the staurolite porphyroblasts grew early syn-D_2. Similarly, kyanite is seen to preserve quartz inclusion trails which curve

Figure 12. An F_2 micro-fold outlined by syn-D_1 sillimanite (sillimanite needles locally are pulled apart prior to D_2 folding) is preserved in a microlithon during the formation of the S_2 foliation and transposition of the S_1 fabric. Long dimension of field of view is 5 mm. Location: East of Port aux Basques, along coastline.

at each end into the external transposive S_2 fabric (Figs. 10, 11); we interpret this kyanite to be early syn-D_2. Figure 8 illustrates a large poilkioblastic staurolite with included kyanite, which itself contains a garnet. The S_2 schistosity is strongly flattened around the staurolite indicating growth before the end of D_2. In other thin sections, kyanite, with the S_2 schistosity wrapped around it, contains small rounded inclusions of relict staurolite, suggesting that some staurolite growth and resorption predates S_2. Some kyanite is present as inclusion-poor porphyroblasts the relative age of which is not well known at this stage. Locally, the regional fabric could be S_1. Sillimanite evidences growth at different stages in the history. Strained fibrolite and pulled apart sillimanite needles are preserved locally as relict microfolds of S_1 (Fig. 12). Also, sillimanite is found within and defining the S_2 schistosity, often associated with muscovite. Near Port aux Basques, sillimanite is most commonly manifested as a pseudomorphic replacement of syn-D_2 kyanite.

ISOGRADS AND METAMORPHIC CONDITIONS IN THE PORT AUX BASQUES DIVISION

Five isograds have been mapped (Fig. 1). These isograds correspond to the first and last appearance of staurolite and kyanite, and the first appearance of sillimanite in pelitic rocks. Due to the complexity of mineral growth during deformation, interpretation of these isograds needs to be approached with caution. The distribution of isograds indicates a regional increase in metamorphic grade from west to east in the Port aux Basques Division, with the eastern portion consisting solely of sillimanite + garnet + biotite bearing assemblages. We interpret this to be an area where sillimanite forms by more than one reaction. However, at the 'sillimanite in' isograd, kyanite begins to be reacted out by the polymorphic transformation of kyanite to sillimanite, but persists metastably to higher grade.

The microstructural interpretation described above indicates that metamorphic conditions were appropriate for sillimanite stability during both D_1 and D_2. The D_2 event was the more dominant and overprints D_1 throughout the area, increasing in intensity to the east. The polyphase growth of minerals appears to span both D_1 and D_2 and suggests a progressive metamorphism at elevated pressures and temperatures near the univariant kyanite to sillimanite transition. No evidence is seen that these rocks were involved in any substantial decompression and cooling prior to the D_2 event. The early syn-D_2 growth of kyanite supports this observation. Thus, it appears that the rocks remained at moderate pressures and temperatures during D_1 and D_2 and may have crossed from the kyanite stability field more than once. This may reflect successive stages in orogenic thickening.

Conditions of metamorphism may be estimated from a petrogenetic grid. The development of migmatites in felsic gneisses and metabasic rocks within the Port aux Basques Division, interpreted to represent partial melting, argues for conditions above the H_2O-saturated solidus for these rocks.

Biotite, garnet, staurolite, kyanite and sillimanite assemblages in higher grade pelitic rocks, are followed by the eventual breakdown of staurolite and kyanite to yield sillimanite, garnet and biotite assemblages. These parageneses indicate peak conditions in the range of 650-725°C at 8-10 kbar. A qualitative P-T trajectory for the progressive D_1 and D_2 events indicates an overall clockwise path in P-T space, supported by the kyanite to sillimanite transformation, that may cross the kyanite to sillimanite univariant reaction on more than one occasion, due to successive thickening events.

CONCLUSIONS

Our preliminary results indicate a complex metamorphic and deformational history for the Port aux Basques Complex. Rocks in the Grand Bay Division do not have a migmatitic aspect and appear to be of lower metamophic grade than the other two divisions. The Otter Bay Division contains sillimanite-bearing semipelites, which reflects amphibolite facies metamorphism. The Port aux Basques Division contains rocks which have been involved in four phases of progressive deformation with the peak metamorphic event apparently related to both the first and second phases of deformation. This metamorphism increases in intensity to the southeast, in the direction of the dip of compositional layering. Metamorphism in the Port aux Basques Division has been attributed to Silurian orogenesis (Dunning et al., 1990). High pressures and high temperatures in response to crustal thickening provide a reasonable mechanism for the prograde metamorphic sequence observed.

The tectono-stratigraphic affinities of the lithologies within the three divisions presently are not well understood. The rocks in the Grand Bay and Port aux Basques divisions contain dominantly marine metasedimentary rocks and a large percentage of metabasic horizons with some ultrabasic pods. The two latter lithologies are rare or absent in the Otter Bay Division, which is a distinctly different package of rocks. Thus, the Isle aux Morts Fault Zone may represent a major tectono-stratigraphic boundary as suggested previously (e.g. Colman-Sadd et al., 1990).

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Geological Survey of Canada Project 730044

Compilation cartographique et caractérisation des dépôts de surface de la région de Shawinigan – Trois-Rivières, Québec

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Résumé

La région de Shawinigan - Trois-Rivières est recouverte à environ 90% de dépôts meubles, les zones rocheuses se trouvant principalement dans les Laurentides. Les principaux types de sédiments sont le till, les sédiments fluvio-glaciaires, les argiles marines, les sables d'exondation, et les sédiments organiques. Si l'on fait exception de l'aspect spectaculaire de la moraine de Saint-Narcisse, les formes morainiques sont peu nombreuses. Quelques eskers, deltas, cordons littoraux, et dunes sont également présents. Le niveau maximum de la mer de Champlain dans la région se situe à une altitude d'environ 213 m. Les données disponibles indiquent que la direction d'écoulement des glaces au Wisconsinien supérieur était vers le sud-sud-est (135° à 170°), que la matrice du till est sablo-silteuse, que la composition du till reflète celle du substratum rocheux et n'est que peu reliée à la dispersion glaciaire.

Abstract

The Shawinigan - Trois-Rivières area is about 90% covered by surficial deposits, bare areas being located mainly in the Laurentians. The main sediment types are till, fluvioglacial sediments, marine clays, regressive sands and organic deposits. There are few morainic landforms in the area, the St. Narcisse Moraine being the most spectacular. Eskers, deltas, shoreline features and dunes are also present. Maximum level of the Champlain Sea was about 213 m a.s.l. Available data indicate that late Wisconsinan ice flow was to the south-southeast (135° to 170°), that till matrix is sandy-silty, and that the composition of till is related to the underlying bedrock and only slightly related to glacial dispersal.

INTRODUCTION

Bien que de nombreux travaux de cartographie des dépôts de surface existent au Québec méridional (Gadd, 1971; McDonald et Shilts 1971; Occhietti, 1980; LaSalle, 1989; voir Maurice, 1988 pour une bibliographie complète), la plupart des travaux récents (Lamothe, 1989; Occhietti, 1990) portent sur la stratigraphie des basses terres du Saint-Laurent. Le besoin d'uniformiser à l'échelle régionale les données cartographiques déjà existantes s'était déjà fait sentir (Ministère de l'Énergie et des Ressources du Québec, 1984a et 1984b). La Commission géologique du Canada a entrepris un projet de longue haleine, soit la compilation cartographique des données existantes sur les dépôts de surface au Québec méridional avec de nouveaux travaux sur le terrain (fig. 1). L'aspect principal du projet est la production de cartes régionales à 1:100,000 des dépôts de surface. L'exécution des travaux de terrains et les analyses de laboratoire sur les échantillons prélevés durant les campagnes de terrain permettront d'élaborer une caractérisation régionale des dépôts de surface, ainsi que leur cadre stratigraphique régional. Le présent article fait état des données et résultats partiels obtenus pour les cartes de Shawinigan (31I/NE) et Trois-Rivières (31I/SE).

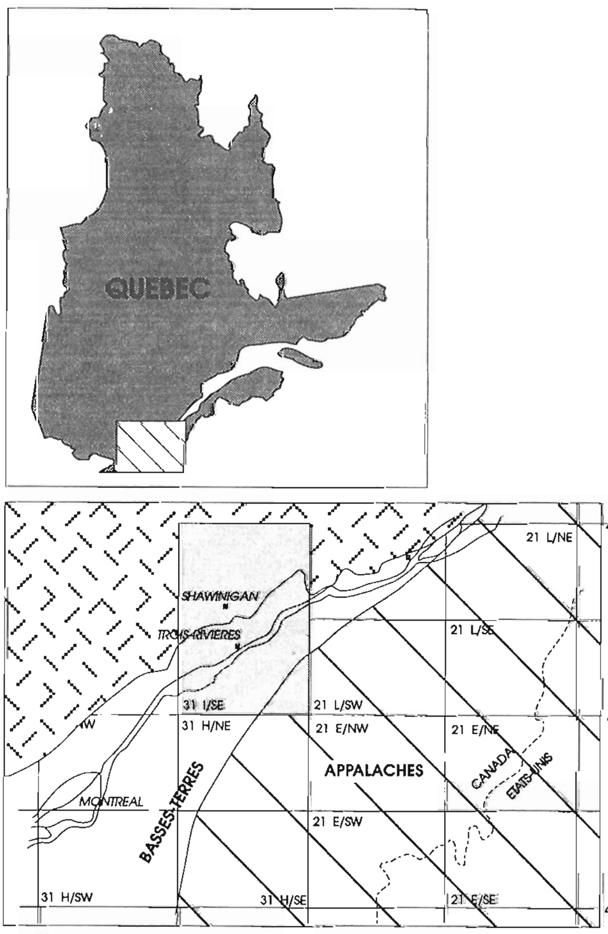

Figure 1. Localisation des travaux de compilation cartographique des dépôts de surface au Québec méridional. Les levés qui font l'objet du présent document ont été effectués en 1990 (31I/NE) et 1991 (31I/SE). Les grandes provinces physiographiques sont aussi indiquées.

LOCALISATION

La région étudiée est située entre les longitudes 072°W et 073°W et les latitudes 46°N et 47°N (fig. 1) et est couverte par les cartes SNRC 31I/01, 31I/02, 31I/07, 31I/08, 31I/09, 31I/10, 31I/15 et 31I/16. Un bon réseau routier dessert le milieu agricole localisé principalement dans les basses terres, alors que le réseau routier principal dans les Laurentides peut être complété par l'utilisation des chemins forestiers.

CADRE PHYSIOGRAPHIQUE

La région de Shawinigan - Trois-Rivières est située sur les hautes terres des Laurentides (secteur nord de la région), les basses terres centrales du Saint-Laurent, et le piedmont appalachien (secteur sud-est) (fig. 1). L'altitude varie entre près de 0 m le long du Saint-Laurent, et plus de 400 m dans les Laurentides. Au sud du fleuve, l'altitude maximale est de 120 m à la bordure appalachienne. Le secteur situé au nord du fleuve est drainé par les rivières Saint-Maurice, Batiscan et Sainte-Anne. Quant au secteur sud, il est drainé par les rivières Yamaska, Saint-François, Nicolet et Bécancour.

Le relief des hautes terres des Laurentides est surtout commandé par la topographie du socle, mais, dans les vallées, il est atténué par les dépôts quaternaires. Le relief plutôt plat des basses terres est principalement celui de la plaine argileuse et des sables d'exondation déposés par la mer de Champlain. Entre la rivière Batiscan et la limite est de la région, la démarcation entre le terrain plat, et les hautes terres des Laurentides correspond assez bien au contact structural entre le Bouclier et les basses terres du Saint-Laurent. Toutefois, entre la marge ouest de la région et la rivière Batiscan, la plaine s'étend jusqu'à 25 km au nord-ouest de la limite du Bouclier et des basses terres. Le piedmont appalachien dans la région est recouvert de sédiments de la mer de Champlain, et est topographiquement similaire aux basses terres.

CADRE GÉOLOGIQUE

Sur le Bouclier (fig. 2), le socle peut être séparé en cinq grands groupes: 1) le domaine de Mékinac composé principalement de gneiss charnockitiques; 2) des gneiss plutoniques et roches supracrustales hautement métamorphisées; 3) le groupe de Montauban, des métasédiments et métavolcaniques; 4) le complexe de la Bostonnais, des diorites, granodiorites et quartz-diorites; et 5) des granites (Nadeau et Corrigan, 1991; Nadeau, comm. pers. 1991). Dans les basses terres du Saint-Laurent, la plate-forme sédimentaire est composée de calcaire des groupes de Trenton et de Black River au nord, en bordure du Bouclier, et des shales, grès, calcaires et dolomies des groupes de Sainte-Rosalie, d'Utica, de Lorraine, et de Queenston (Avramtchev et al., 1989). Dans le coin sud-est de la région, les roches métasédimentaires appalachiennes sont composées principalement de schistes, ardoises et calcaires du groupe de Sillery et des formations de Bullstrode et de Melbourne (Avramtchev et al., 1989).

Les roches du Bouclier peuvent être utilisées afin d'évaluer une distance minimum de transport glaciaire à l'extérieur du Bouclier. La maille d'échantillonnage des tills n'est toutefois pas assez serrée pour documenter la dispersion de roches particulières. Dans les basses terres du Saint-Laurent et les Appalaches, certaines unités peuvent servir d'indicateurs. Entre autres, la formation de Deschambault du groupe de Trenton (Clark et Globensky, 1975; Lavoie, comm. pers.), qui affleure au contact entre le Bouclier et les basses terres du Saint-Laurent, ainsi que la formation de la montagne de Saint-Anselme du groupe de Saint-Roch (Avramtchev et al., 1989), qui a deux petites zones d'affleurements près de Saint-Léonard-d'Aston, ont un bon potentiel à titre d'indicateurs lithologiques. La formation de Deschambault est un calcaire à crinoïdes facilement reconnaissable dans les cailloux du till, tandis que la formation de la montagne de Saint-Anselme est une volcanite basique verte et blanche très distinctive.

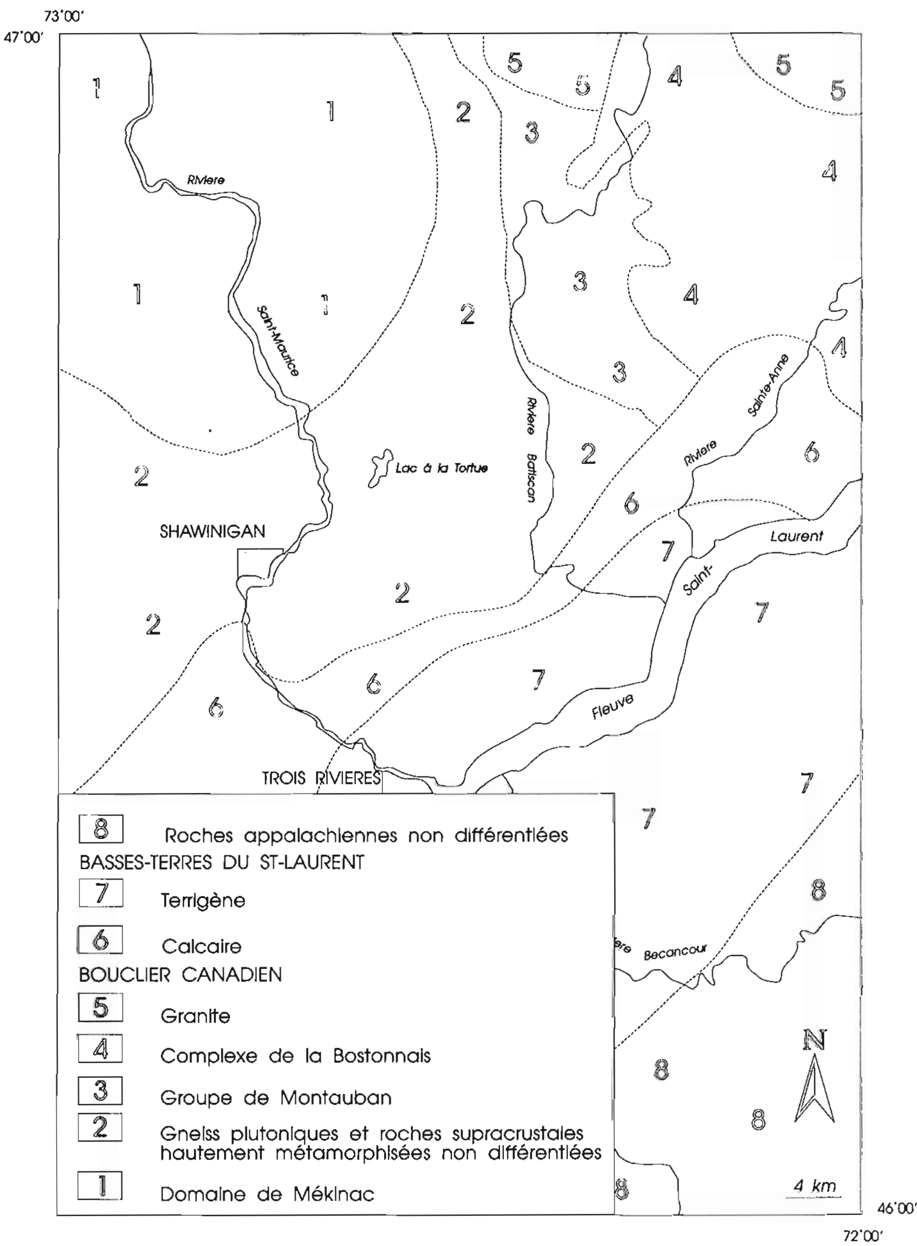

Figure 2. Géologie du substratum rocheux de la région étudiée, compilée à partir de Nadeau et Corrigan (1991), Béland (1961) et Pyke (1964) pour le Bouclier, et Avramtchev et al. (1989) pour les Basses-terres et les Appalaches.

GÉOLOGIE DU QUATERNAIRE

Dans la région étudiée, les sédiments meubles sont principalement attribués à l'action de l'inlandsis laurentidien du Wisconsinien supérieur, de même qu'à l'activité de la mer de Champlain. La stratigraphie quaternaire des basses terres du Saint-Laurent est complexe, et le cadre stratigraphique établi évolue rapidement (voir, entre autres, Parent et Occhietti, 1988; Lamothe, 1989; Occhietti, 1990; Bernier et Occhietti, 1991). L'objet du présent rapport est de présenter les nouvelles données relatives à la dernière phase glaciaire. Il ne sera donc que très peu question de la stratigraphie quaternaire; toutefois, un bref aperçu est présenté à l'aide d'une colonne stratigraphique composite de la région étudiée (fig. 3).

La plus vieille unité glaciaire des basses terres du Saint-Laurent, le till de Bécancour, a été attribuée à l'Illinoien (Lamothe, 1989). Le till de Bécancour est surmonté de rythmites qui indiqueraient une déglaciation en milieu glaciolacustre. Le sable de Lotbinière, qui repose en discordance sur ces rythmites a été attribué à la fin du Sangamonien (Lamothe, 1989). Ces sables fluviatiles indiquent une paléogéographie fluviale similaire à l'actuelle, quoique les débris organiques indiquent un climat plus froid que l'actuel. Une argile marine ancienne, l'argile de la Pérade, pourrait avoir été mise en place entre l'épisode glaciolacustre représenté par les rythmites et l'épisode fluviatile du sable de Lotbinière, mais une autre position stratigraphique, plus jeune, est aussi possible (Ferland et Occhietti, 1990).

Les varves de Deschaillons et le till de Lévrard forment une séquence continue surmontant le sable de Lotbinière, sans lacune de sédimentation. Cette séquence glaciaire daterait du Wisconsinien inférieur (Lamothe, 1989). Les sédiments de Saint-Pierre reposent en discordance sur le till de Lévrard, et indiquent une sédimentation non-glaciaire dans un climat plus froid que maintenant. L'autre position possible de l'argile de la Pérade se situe entre le till de Lévrard et les sédiments de Saint-Pierre (Ferland et Occhietti, 1990). Les sédiments de Saint-Pierre auraient été déposés durant le Wisconsinien moyen (Lamothe, 1989).

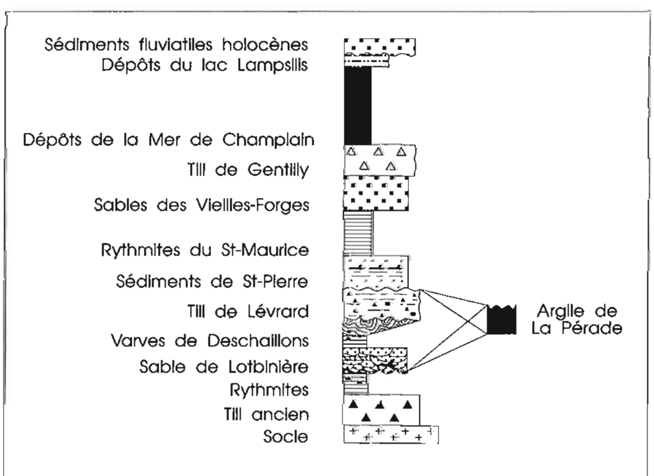

Figure 3. Colonne stratigraphique composite générale pour les basses terres centrales du Saint-Laurent (modifiée d'après Ferland et Occhietti, 1990). Au nord de la limite marine, dans les hautes terres des Laurentides, la formation de Mattawin (un équivalent latéral au till de Gentilly et des sédiments fluvio-glaciaires (Occhietti, 1980)) déposée lors du dernier épisode glaciaire est exposée.

Les rythmites du Saint-Maurice surmontent les sédiments de Saint-Pierre en discordance indiquant une sédimentation lacustre dans un climat plus chaud que celui indiqué par les sédiments de Saint-Pierre (Besré et Occhietti, 1990). Par-dessus les rythmites du Saint-Maurice, les sables des Vieilles-Forges indiquent une sédimentation dans un bassin peu profond, de type épandage ou deltaïque (Besré et Occhietti, 1990). Le contact entre les rythmites du Saint-Maurice et les sédiments de Saint-Pierre est interprété comme étant une lacune d'érosion (Lamothe, 1987) ou une lacune intraformationnelle sommitale (Ferland et Occhietti, 1990; Besré et Occhietti, 1990), et la séquence présentée par les sédiments de Saint-Pierre - rythmites du Saint-Maurice - sables des Vieilles-Forges n'est pas démontrée comme étant continue.

Le till de Gentilly, surmontant les sables des Vieilles-Forges, représente le dernier événement glaciaire dans les basses terres du Saint-Laurent, et le till en surface des hautes terres des Laurentides (formation de Mattawin, Occhietti, 1980) en est l'équivalent latéral (Parent et Occhietti, 1988). Les dépôts de la mer de Champlain sont en contact concordant avec le till de Gentilly. Des silts lacustres du lac Lampsilis (Parent et Occhietti, 1988) surmontent les sédiments de la mer de Champlain, et attestent d'une phase lacustre antérieure au rétablissement du système fluvial actuel, représenté par des sables fluviatiles.

TRAVAUX ANTÉRIEURS

Des travaux de cartographie ont été effectués sur des parties de la région de Shawinigan (Karrow, 1957; Allard, 1978; Occhietti, 1980; Gagnon et Morelli, 1986; LaSalle, 1989) de même que sur la totalité de la région de Trois-Rivières (Gadd, 1971). Les travaux de cartographie de Karrow (1957) ont été inclus dans l'ouvrage de Gadd (1971) et seront présentés dans ce contexte. La cartographie antérieure de la région étudiée a permis d'établir la localisation des principaux types de sédiments meubles, des zones rocheuses, ainsi que la direction du dernier écoulement glaciaire, principalement d'après l'examen des stries (figs. 4 et 5).

Au-dessus de la limite marine, dans les hautes terres des Laurentides, le till et les sédiments juxtaglaciaires de la formation de Mattawin (Occhietti, 1980) se trouvent principalement dans les vallées. Des moraines mineures indiquent des positions de retrait de la marge glaciaire (Occhietti, 1980). Trois axes principaux de drainage fluvio-glaciaire sont également identifiés, soit dans la région de Saint-Élie, Sainte-Thècle, et Notre-Dame-de-Montauban (Occhietti, 1980).

Sous la limite marine (environ 213 m, Occhietti, 1980), des dépôts argileux (sédimentation en eaux profondes) et sableux (exondation) couvrent la plupart des dépôts antérieurs, sauf les dépôts associés à la moraine de Saint-Narcisse, qui ont toutefois été remaniés par la mer de Champlain (Occhietti, 1980). La moraine de Saint-Narcisse témoigne d'un épisode de stabilisation du retrait glaciaire, ou peut-être de réavancée glaciaire (Parent et Occhietti, 1988) durant la phase d'inondation marine de la vallée du Saint-Laurent. Les structures de poussées dans la moraine indiquent que la glace est demeurée active durant cet épisode.

Au sud du fleuve Saint-Laurent, de nombreux affleurements de till avaient été attribués au till de Bécancour et au till de Gentilly (Gadd, 1971). Quelques corps de sédiments fluvio-glaciaires près de Grand-Saint-Esprit et de Notre-Dame-du-Bon-Conseil sont également présents (Parent et Occhietti, 1988). Des sédiments éoliens se trouvent en quantité mineure dans la région.

Les sédiments fluviatiles post-champlainiens (sables des hautes, moyennes et basses terrasses, Gadd, 1971; Occhietti, 1980) sont présents dans la vallée du Saint-Laurent, établissant la mise en place du système fluvial actuel.

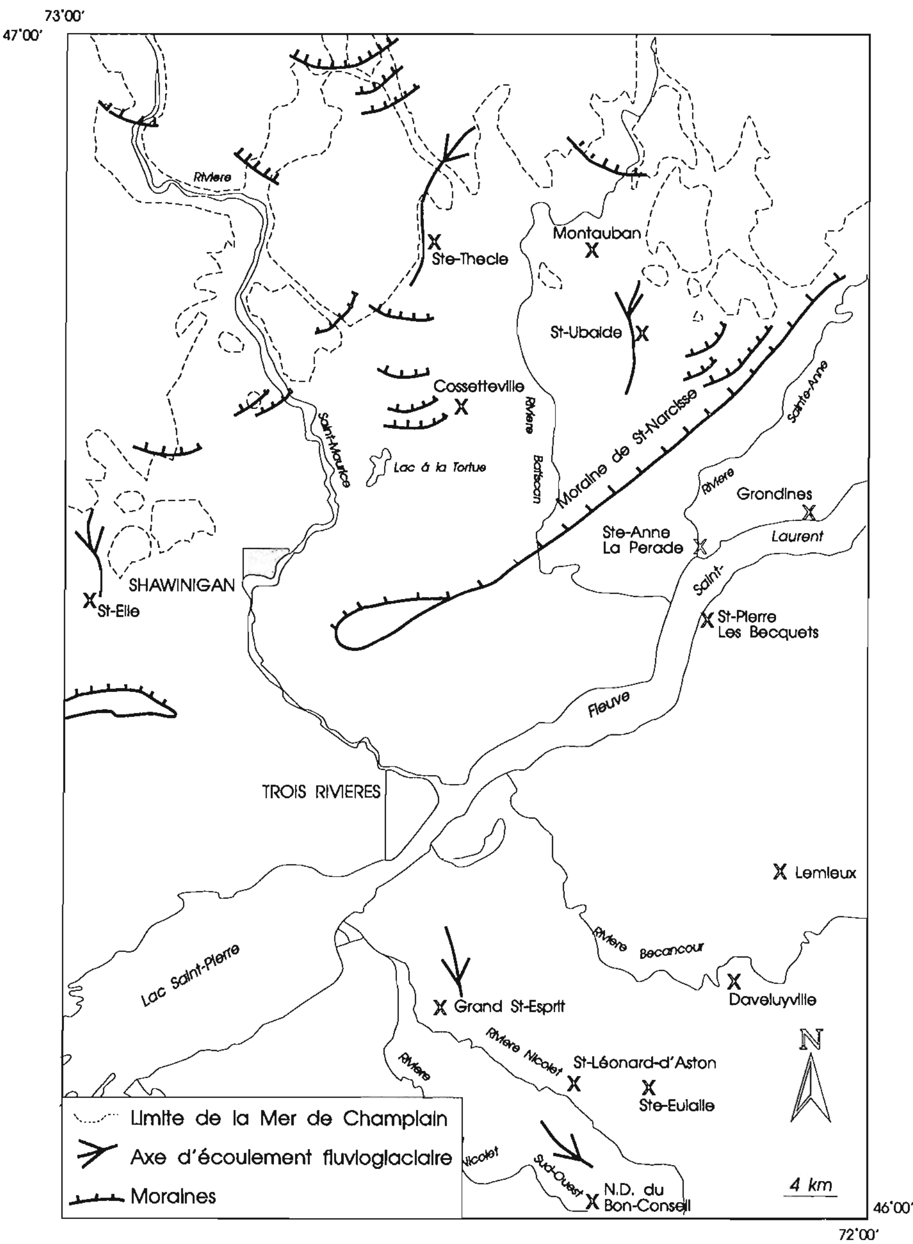

Figure 4. Sommaire de l'histoire quaternaire de la région étudiée. Position des moraines et eskers modifiée d'après Parent et Occhietti (1988); position de la limite marine à 213 m.

Finalement, des sédiments organiques sont présents dans toute la région, mais principalement au sud du Lac-à-la-Tortue et au nord de Grondines.

À quelques endroits d'aire très restreinte, des sédiments antérieurs au Wisconsinien supérieur affleurent en surface, surtout sur les flancs de vallées fortement encaissées. Il s'agit des régions de Grondines, Saint-Pierre-les-Becquets et le long de la rivière Saint-François (Gadd, 1971).

MÉTHODE

Outre les habituels levés de terrain, des échantillons de till ont été prélevés en essayant d'obtenir une couverture uniforme pour la région, pour fins d'analyse de composition, géochimique et lithologique, calcimétrique, et granulométrique. D'autres échantillons, par exemple de sites fossilifères ou des argiles, sont prélevés au besoin.

La photo-interprétation finale de la région visitée est effectuée après les levés de terrain. L'analyse des données de laboratoire s'effectue au rythme de leur disponibilité.

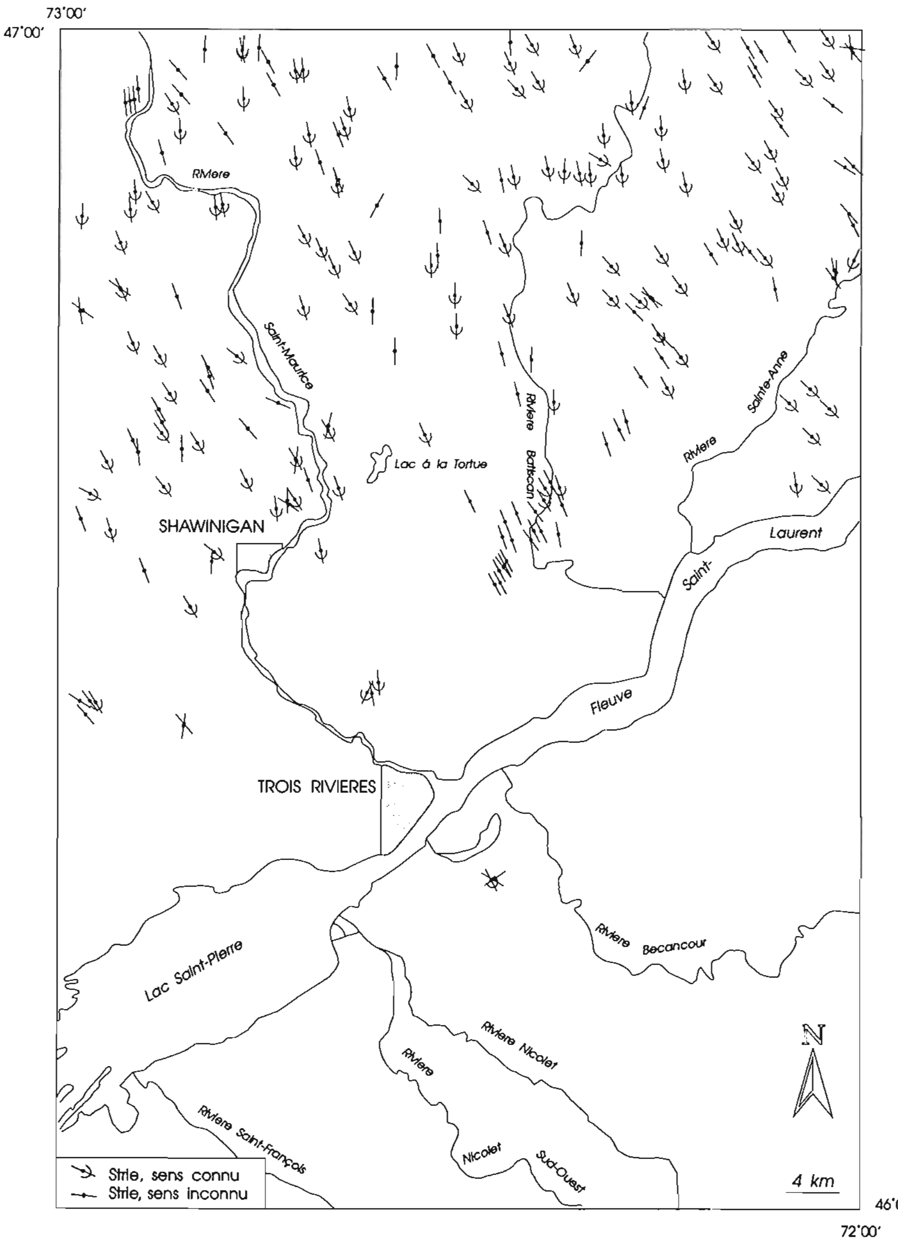

Figure 5. Distribution des stries dans la région étudiée (sources : Gadd et al., 1972; Occhietti, 1980; LaSalle, 1989; Bernier et Occhietti, 1991; cette étude).

RÉSULTATS ET DISCUSSION

Direction du dernier mouvement glaciaire

Les marques d'écoulement glaciaire relevées dans le cadre des travaux de terrain (fig. 5) indiquent un écoulement vers le sud-est (158°). Il n'est pas possible de déterminer l'âge des stries, ni d'établir une chronologie relative lorsque plusieurs familles de stries sont présentes sur le même affleurement. La direction des stries semble avoir été contrôlée par le dernier écoulement régional, modifié à certains endroits par un contrôle local topographique.

Des stries attribuées à un mouvement glaciaire antérieur au Wisconsinien supérieur (localisées sous un till ancien) ont été levées dans la région de Grondines (Bernier et Occhietti, 1991). Ces stries indiquent aussi un écoulement vers le sud-est. La concordance entre les directions d'écoulement glaciaire de diverses époques amène l'impossibilité d'attribuer avec certitude les stries à un mouvement glaciaire particulier. En l'absence de dépôts anciens superposés, la seule hypothèse valable est de présumer que les stries levées dans cette étude ont été formées lors du dernier épisode glaciaire dans la région.

Le relief quaternaire

La région située au nord de la limite marine n'est que partiellement recouverte de dépôts quaternaires, surtout du till et des sédiments juxtaglaciaires localisés dans les vallées. La plupart des sommets sont dénudés, ou en partie recouverts de till mince. La surface rocheuse est toutefois polie et striée, attestant du passage des glaces. Les formes morainiques sont peu nombreuses et sont localisées au nord de la moraine de Saint-Narcisse. De petites moraines sont présentes dans les vallées des Laurentides, et servent souvent à retenir l'écoulement des lacs. Sous la plaine argileuse déposée par la mer de Champlain, une série de moraines frontales mineures situées à l'ouest de Cossetteville (Béland, 1961; Occhietti, 1980) sont espacées d'environ 100 m les unes des autres, indiquant le retrait, peut-être annuel, du front glaciaire (Occhietti, 1980).

La moraine de Saint-Narcisse est sans contredit la forme de terrain la plus spectaculaire de la région. Elle traverse le coin nord-ouest de la carte de Trois-Rivières, et la carte de Shawinigan, selon un axe SW-NE. Elle est localisée au contact des Laurentides et des basses terres du Saint-Laurent, la rupture de pente associée au contrefort des Laurentides ayant joué un rôle dans la position de la moraine (Occhietti, 1980). Les matériaux qui composent la moraine sont variés: du till, des sédiments juxtaglaciaires et des sédiments glaciomarins. Les sites datés au ^{14}C de part et d'autre de la moraine indiquent que l'épisode Saint-Narcisse a eu lieu entre 11,000 et 10,800 BP (Occhietti, 1980).

Au sud du Saint-Laurent, Gadd (1971) avait identifié un axe SSW-NNE de 3 à 8 km de largeur, passant au nord-ouest de Sainte-Eulalie et Daveluyville, ayant un faible relief de moins de 8 m, rappelant une forme morainique, la moraine de Drummondville. Toutefois, Prichonnet (1984), et Parent et Occhietti (1988), entre autres, réfutent son existence. Prichonnet (1984), sur la foi d'une étude détaillée de 90 km du tracé de la "moraine", attribue les sables et graviers présents dans cette région à l'action littorale plutôt qu'à une activité glaciaire.

Les autres formes de terrain glaciaires présentes dans la région étudiée sont quelques eskers, situés au nord de la moraine de Saint-Narcisse, et au sud du Saint-Laurent. De plus, des terrasses marines et fluviatiles, de même que des cordons littoraux (surtout au sud du Saint-Laurent), attestent de la présence et du retrait de la mer de Champlain, de même qu'à l'incision du réseau de drainage actuel.

Les sédiments glaciaires

Dans la région étudiée, une attention particulière a été portée sur les caractéristiques physiques du till de surface: couleur, texture, et composition lithologique et géochimique. Les données disponibles indiquent que le till de surface varie selon le type de roche sous-jacente, et le facies présent.

La couleur

La couleur (code Munsell) du till est peu variable lorsque le till est localisé sur le Bouclier, tandis qu'elle est très variable dans les basses terres du Saint-Laurent et le piedmont appalachien. Sur le Bouclier, les tills sont en général de couleur gris olive à gris lorsque non-oxydés. Les tills oxydés sont de couleur brunâtre, particulièrement brun olive. Les échantillons prélevés dans les basses terres du Saint-Laurent sont de couleur plus variable, quoique les tills localisés sur la formation de Queenston (shales rouge et vert) sont de couleur brun rougeâtre à brun, indiquant un apport appréciable de shale rouge dans la matrice. Le till échantillonné dans les Appalaches est aussi de couleur variable, mais les teintes brunes semblent dominer. Les tills à teinte brune des basses terres du Saint-Laurent et des Appalaches ne sont pas nécessairement oxydés.

Lorsque la variété des couleurs du till de surface est mis en relation avec leur localisation, et la composante lithologique principale du till, il apparait clair que l'on ne peut différencier des nappes de till d'âge différent sur la seule foi de la couleur, comme cela semble avoir été le cas au début des années 1970 (Gadd, 1971). En effet, la couleur d'un till dépend essentiellement de sa composition lithologique, ainsi que de son degré d'oxydation. Des nappes de till d'âge différent, mais qui auraient traversé les mêmes unités lithologiques devraient avoir à peu près la même couleur. En l'absence de stratigraphie probante, les tills de surface de couleur rouge dans la région de Trois-Rivières ne sont pas considérés comme étant équivalents au till de Bécancour décrit et cartographié par Gadd (1971), mais ils sont plutôt associés au till de Gentilly.

Granulométrie

Les données sur la texture de la matrice des tills prélevés dans la région de Shawinigan montrent que le pourcentage d'argile ($< 2\ \mu m$) est inférieur à 10% dans 93% des échantillons. Les échantillons contenant plus de 10% d'argile ne semblent pas présenter d'autres caractéristiques communes (localisation, topographique ou par rapport au type de roche en place). Il s'agit sans doute plutôt d'un faciès différent.

En effet, la plupart des tills répertoriés dans la région de Shawinigan présente de minces (< 2 cm) passées sableuses gris pâle de quelques dizaines de centimètres de longueur. La présence systématique de ces passées sableuses, combinée à la matrice sablo-silteuse, indique que ces tills sont sans doute des tills d'ablation, alors que les tills plus "argileux" sont peut-être des tills de fond.

Dans la région de Trois-Rivières, les données granulométriques ne sont pas encore disponibles. Toutefois, la plupart des tills sont plus compacts que ceux prélevés dans la région de Shawinigan, principalement sur le Bouclier. Peu de sites présentaient des passées sableuses, indiquant que le

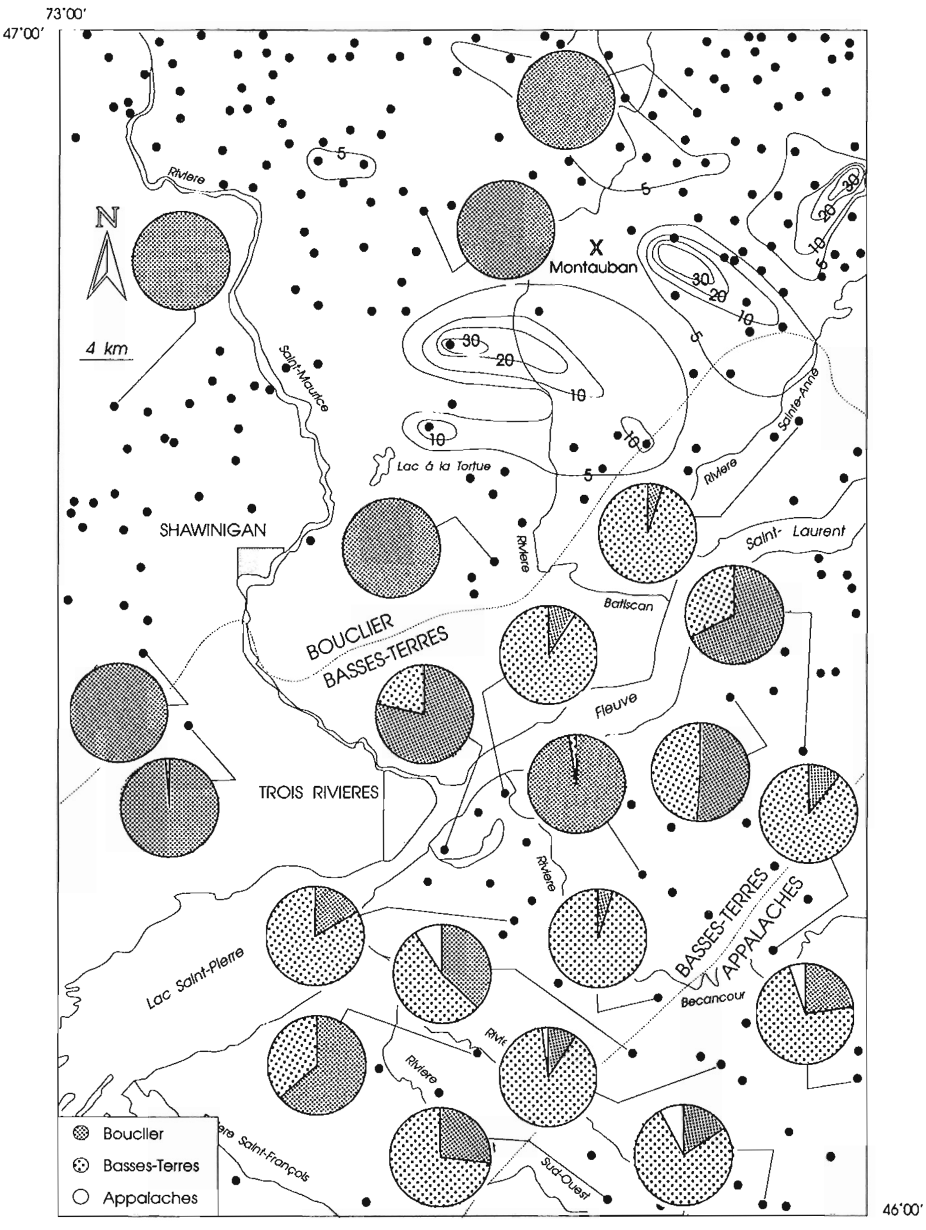

Figure 6. Composition lithologique et géochimique des tills de surface de la région étudiée. Sur le Bouclier, les tills sont composés à 100% de cailloux cristallins, alors que dans les basses terres du Saint-Laurent et les Appalaches, la composition lithologique reflète l'apport en cailloux du socle sous-jacent. La distribution de l'uranium dans la fraction argile (< 2 µm) des tills de la région de Shawinigan indiquent une concordance avec le socle sous-jacent. Une traînée de dispersion locale semble provenir de Montauban-les-Mines, et s'étend sur environ 15 km en direction sud-est.

till présent dans les basses terres du Saint-Laurent est sans doute un till de fond, remanié par endroit par la mer de Champlain, pour donner un till plus sableux. Lorsque ces sites sont excavés plus profondément, le till devient rapidement plus compact et plus argileux, sans doute le faciès till de fond.

Composition géochimique et lithologique

Dans la région de Shawinigan - Trois-Rivières, la composition du till semble refléter assez fidèlement celle de la roche sous-jacente. Ainsi, bien que des comptages lithologiques n'aient pas été effectués dans la région de Shawinigan, des observations qualitatives indiquent que les tills localisés sur le Bouclier ont 100% de cailloux cristallins, tandis que ceux localisés dans les basses terres du Saint-Laurent ont jusqu'à 95% de roches sédimentaires (fig. 6). Des comptages lithologiques sur la fraction >8 mm des tills de la région de Trois-Rivières montrent que les tills des basses terres du Saint-Laurent ont jusqu'à 100% de cailloux sédimentaires. Les cailloux d'origine appalachienne apparaissent dans le till dès que la limite entre les basses terres du Saint-Laurent et les Appalaches est franchie, mais sont en quantité minime. Toutefois, il n'a pas été possible de départager de façon satisfaisante les shales et grès pouvant provenir des basses terres du Saint-Laurent de ceux pouvant provenir des Appalaches; aussi, seuls les cailloux dont l'origine appalachienne ne faisait aucun doute ont été inclus dans cette catégorie.

Les distances et orientations du transport peuvent être évaluées plus précisément grâce aux indicateurs lithologiques. Ainsi, les roches du Bouclier sont présentes dans le till dans toute la région étudiée, indiquant un transport glaciaire minimal de l'ordre de 70 km, selon une orientation vers le "sud". Des cailloux provenant de la formation de Deschambault ont été trouvé dans un till situé près de Lemieux, pour une distance de transport d'environ 25 km selon une orientation sud-est. Finalement, des cailloux de la formation de la montagne de Saint-Anselme ont été trouvés à quelques sites à l'est-sud-est, et jusqu'à 12 km de leur lieu d'origine.

L'analyse des patrons de dispersion géochimique dans la fraction argile (<2 µm) des tills n'est pas complétée. Toutefois, sur une base préliminaire, il semble y avoir peu de dispersion à l'échelle de l'échantillonnage. L'élément le plus intéressant semble être l'uranium (fig. 6). Deux particularités de la carte retiennent l'attention.

Premièrement, il semble y avoir dispersion à l'aval glaciaire du gîte de Montauban. Bien qu'il ne semble pas y avoir beaucoup d'uranium dans le gîte (2 à 25 ppm; Morin, 1987), l'aire de dispersion est bien délimitée, et commence à environ 7,5 km au sud-est du gîte. Avec une maille d'échantillonnage beaucoup plus serrée autour du gîte de Montauban; Pelletier et Beaumier (1990) ont également détecté une dispersion vers le sud-sud-est qui commence à environ 6 km du gîte. Les hautes teneurs en uranium trouvées dans la présente étude pourrait venir d'un granite situé immédiatement à l'amont glaciaire des échantillons à valeurs anomales.

Deuxièmement, il semble y avoir une certaine concordance entre les zones à faible teneur en uranium (en général < 3 ppm) et le domaine de Mékinac, les gneiss plutoniques et les granites du Bouclier, et les basses terres du Saint-Laurent. Toutefois, sur le groupe de Montauban et le complexe de la Bostonnais, les teneurs en uranium sont plus élevées (en général, > 3 ppm). La raison de ce patron est encore à établir.

Bien que l'interprétation géochimique et lithologique ne soit pas complétée, les grandes lignes qui se dégagent semblent indiquer, dans un premier temps, que la composition du till de surface est directement liée au type de roche sous-jacente. Toutefois, à quelques endroits, il est possible de détecter l'action glaciaire suivant un patron de dispersion classique, soit haute teneur à la tête de la traînée, près de la source, et une décroissance progressive dans la direction d'écoulement.

CONCLUSION

Dans la région de Shawinigan - Trois-Rivières, les sédiments quaternaires consistent en du till, principalement d'ablation au nord, et principalement de fond au sud, des sédiments fluvio-glaciaires, des dépôts de la mer de Champlain, et des sédiments organiques. L'écoulement des glaces s'est effectué vers le sud-est (158°). Les formes de terrain principales sont la moraine de Saint-Narcisse et la plaine argileuse de la mer de Champlain. La moraine de Drummondville n'a pu être identifiée.

Les tills du Bouclier sont caractérisés par une couleur gris à olive, une matrice sablo-silteuse et de nombreuses passées sableuses. Dans les basses terres du Saint-Laurent et les Appalaches, le till est de couleur variable, selon la roche sous-jacente et la matrice semble être plus argileuse. Bien que peu de dispersion glaciaire ait été détectée dans le till de surface, il est toutefois possible d'estimer un transport glaciaire minimum de l'ordre de 70 km pour les cailloux provenant du Bouclier, et de 12 à 25 km pour des cailloux des Appalaches et des basses terres du Saint-Laurent, selon une direction générale vers le sud - sud-est.

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Factor analyses as a method of evaluating sediment environmental quality in Halifax Harbour, Nova Scotia

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Abstract

Statistical factor analyses have been employed to identify communalities between 31 geochemical variables measured in 274 samples of surficial sediments from Halifax Harbour. Concentrations of contaminant metals, such as Zn, Pb, Cu, Cd, Ni, Hg, Cr, and Mn, have been governed by dominant physical and chemical processes in various parts of the Harbour system. Five primary sediment types have been identified from these analyses. Maps illustrating the distribution of the dominant sediment types can be used to identify sediment sources, as well as the area of influence of important diagenetic processes. The primary factor accounts for 41% of the variance in our data. Sediments represented by this factor contain high concentrations of contaminant metals and organic matter derived from anthropogenic sources. Other significant factors relate to fluvial inputs from surface drainage and diagenetic processes that enrich or deplete metal concentrations in the sediments.

Résumé

Des analyses statistiques factorielles ont été utilisées pour relever les traits communs entre 31 variables géochimiques mesurées dans 274 échantillons de sédiments de surface prélevés dans le port de Halifax. Les concentrations de contaminants métalliques, comme Zn, Pb, Cu, Cd, Ni, Hg, Cr et Mn, sont dues aux principaux processus physiques et chimiques ayant cours dans diverses parties du port. Cinq principaux types de sédiments ont été identifiés à partir de ces analyses. Les cartes illustrant la répartition des types de sédiments dominants peuvent être utilisées pour déterminer les sources des sédiments, ainsi que la zone d'influence des principaux processus diagénétiques. Le facteur primaire est à l'origine de 41 % de la variance observée dans les données. Les sédiments représentés par ce facteur contiennent de fortes concentrations de contaminants métalliques et organiques provenant de sources anthropiques. Les autres facteurs significatifs sont liés aux apports fluviatiles amenés par le drainage superficiel et à des processus diagénétiques qui enrichissent ou appauvrissent en concentrations métalliques les sédiments.

INTRODUCTION

Halifax Harbour has been receiving untreated sewage effluent and solid waste since the founding of Halifax in 1749. Metals such as zinc, lead, copper, mercury, and cadmium have been concentrated in Harbour sediments up to 20 times normal background levels (Buckley and Winters, manuscript in preparation). With planned future development of a major metropolitan sewage treatment facility, research was conducted to examine the stability of contaminated deposits, and the impact metals may have on the future environmental quality of the entire estuarine system.

In this paper we identify relationships between analytical variables, using factor analyses techniques, and we interpret significant factors that may indicate dominating physical and chemical processes in various parts of the Harbour system. The basic assumption of factor analyses is that underlying dimensions, or factors incorporated in data matrices, can be used to explain complex phenomena (Norusis, 1990). The most distinctive aspect of factor analyses is that it provides a means of grouping data having common derivation. Given an array of correlation coefficients for a set of variables, factor-analytic techniques (R-mode) are used to determine whether some underlying pattern of relationships exists between groups of variables. Data may be rearranged or reduced to a smaller set of factors or components that may be taken as source variables accounting for the observed interrelations in the data (Cattell, 1965a,b; Harman, 1967; Nie et al., 1970). Observed correlations between variables are due to the their sharing of common factors. Since one of the goals of factor analysis is to reduce a large number of variables to a smaller number of factors or components, factor scores can be estimated for each case, and subsequently used to represent the value of the factors. Our objective is to demonstrate that factor analyses can be used to map the spatial effects of dominant geochemical and sedimentary processes in Halifax Harbour.

SEDIMENT SAMPLING AND ANALYTICAL METHODS

From surficial sediment samples collected during three surveys between 1986 and 1988, we analyzed 274 samples taken from all parts of Halifax Harbour (Fig. 1). Field methods and sampling information are available in reports by Prouse and Hargrave (1987), Buckley et al. (1989) and Winters et al. (1991). Details of laboratory methods are provided in Winters et al. (1991) and Buckley and Winters (manuscript in preparation).

Carbon analyses were carried out using a Leco combustion analyzer. Organic carbon, C_{org}, was determined after removal of the inorganic carbon by treatment with 1 M HCl. Calcium carbonate was calculated from the quantity of inorganic carbon.

Total elemental analyses (M_T) for Si, Al, Ti, Ca, Mg, K, Li, Fe, Mn, Cu, Zn, Ni, Pb, Cr, and Cd were by the modified Buckley and Cranston (1971) total decomposition method, using a HF-H_3BO_3 in a teflon bomb, followed by atomic absorption spectroscopy. Total mercury was determined using flameless cold-vapour atomic absorption spectroscopy as adapted from Brandenberg and Bader (1967).

Potential labile metals, including Fe, Mn, Ca, Cu, Zn, Ni, Cr, and Pb were determined from the sediment samples by a series of three sequential leaches and a separate oxidizing leach. The first sequential leach, M_{WA}, was with weak (25%) acetic acid adjusted to pH 2 for 16 hours, as described in Chester and Hughes (1967). This leach was designed to extract metals from carbonates, some hydrated sulphides, and weakly bound adsorbed metals on mineral surfaces. The second leach, M_{HA}, with 1 M hydroxylamine hydrochloride ($NH_2OH \bullet HCl$) for 16 hours was to extract potentially reducible metals, as described by Chester and Hughes (1967). The third and final sequential leach, M_{HHA}, was with heated (80°C) 0.04 M hydroxylamine hydrochloride at pH 2, and was designed to extract strongly adsorbed metals from mineral surfaces (Tessier at al. 1979). The sum of metal extracts by these three leaches is operationally defined as labile metal and is identified as M_{Labile}.

Metals associated with organic matter, M_{org}, were estimated by carrying out an oxidizing leach using a solution of 10% hydrogen peroxide and 25% acetic acid at pH 2 for 24 hours. The acetic acid was necessary to retard hydrolysis of metals as they were extracted. Because this leach also extracted metal from weak acid soluble forms, it was necessary to correct for this component by subtracting the results from the first sequential leach. Although this adjusted result may correct for some forms of metal not bound to organic matter, there are other metastable mineral phases that are likely to be extracted in the oxidizing leach. The resultant concentration is operationally defined as organically-bound metal, although these results may overestimate metals in this form. Analyses of this type were carried out for Fe, Mn, Cu, Zn, Ni, Pb, and Cr.

The principal axis method of factor extraction (as described in Norusis, 1990) was used to define the "good fit" criterion. Principal axis factoring proceeds in a similar manner as principal components analysis, except that the diagonals of the correlation matrix are replaced by estimates of the communalities (Nie et al. 1970). At the first step, squared multiple correlation coefficients were used as initial estimates of the communalities. Based on these, the requisite number of factors were extracted. The communalities were then re-estimated from the factor loadings, and factors were again extracted, with the new communality estimates replacing the old. Reiteration continued until negligible change occurred in the communality estimates.

In the factor extraction phase, the number of common factors needed to adequately describe the data was determined. Although the factor matrix obtained in the extraction phase indicated the relationship between the factor and the individual variables, it was usually difficult to identify meaningful factors based on this matrix. To enhance the interpretability of the factors, the varimax method of orthogonal rotation was used to transform the initial matrix. This method minimized the number of variables that had loadings on a factor and maximized the variance of squared loadings in each column (Nie et al., 1970). During the rotation, the Kaiser method of normalization was used to discarded the trivial and uninterpretable factors (Kaiser, 1973). When the variance for a variable was not described by the common factors, its uniqueness permitted it to be removed from the analyses. Finally, factor scores were used to determine the dominant factor for each sample and the spatial variation of factor scores was mapped.

RESULTS AND DISCUSSION

Five common factors describing 64% of the total variance in 250 samples were identified, and the factor matrix loadings were sorted by principal variables and listed in Table 1. Four of these samples were unique and were separated from the factor analyses. Variables not described by any of the common factors were identified as unique and removed from the factor analysis. Of the original 274 samples, 24 samples did not include analyses for one or more of the variables, thus requiring their exclusion from the factor analysis.

The dominant factor (four factors only) for individual samples was mapped (Fig. 2). The occurrence of the four most significant statistical factors provides a basis for determining the areas influenced by the dominant sedimentological and geochemical processes.

Factor 1

Variables included in this factor account for 41% of the total variance. These are high loadings for all analytically determined phases of Zn and Pb, total and organically bound Cu, total cadmium, some labile forms, and organically bound Ni and Cr, as well as organic carbon (Table 1).

The distribution of samples dominated by Factor 1 is shown in Figure 2. These include samples taken near major sewage outfalls in the central Harbour area, Bedford Bay, and The Northwest Arm. This distribution pattern indicates the influence of sewage discharge on accumulation of organic-rich and metal-contaminated sediments. Some dispersion of these contaminated sediments into the central area of Bedford Basin is also indicated.

Factor 1 portrayed the anthropogenic input and accumulation of Zn, Cu, Pb, Cr, Ni, Cd, and organic matter. Large proportions of each of these metals were in strong association with the organic matter and may have entered the Harbour as organometallic complexes. Some metal may also have been adsorbed to the organic complexes after entering the marine environment. This factor clearly identified the zones of high urban or industrial effluent input and the most prominent settling basins where this organic-rich mass has been accumulating.

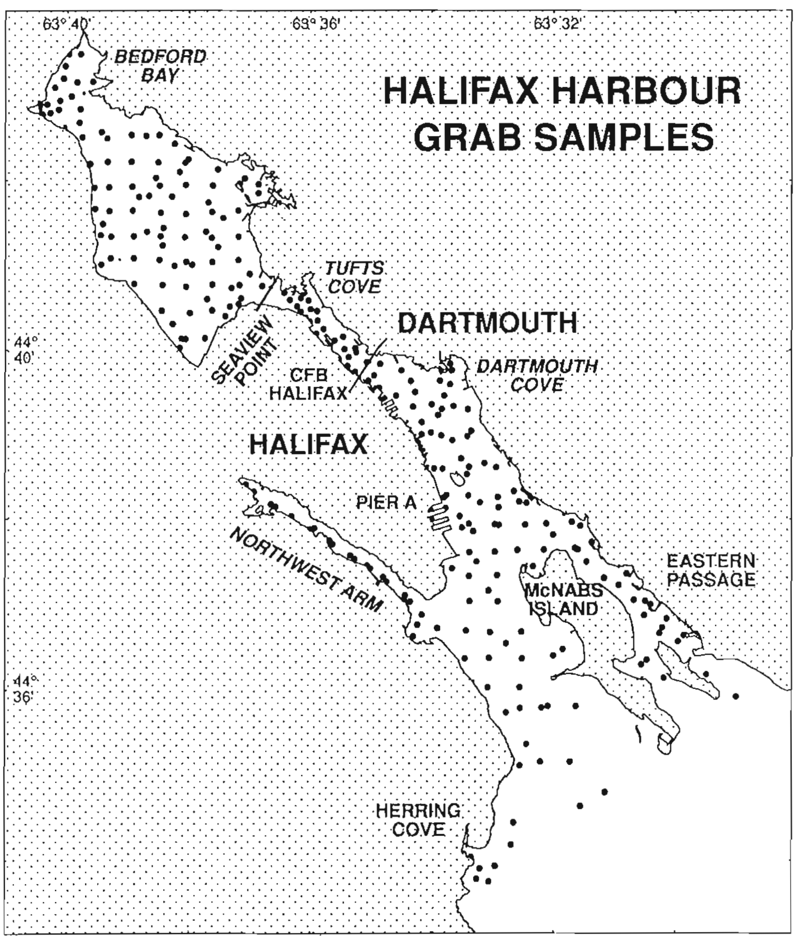

Figure 1. Locations of surficial sediment samples.

There is a general tendency for the finer sedimentary particles on the bottom to be moved toward the head of the Harbour (i.e., Bedford Basin) by inward flowing tidal currents following the deeper channels (Fader and Petrie, 1991). These currents are strongest in constricted areas such as The Narrows near the entrance to Bedford Basin. The current-borne sediments are deposited in regions of very weak flow.

Factor 2

Variables included in this factor account for 8% of the total variance and samples contain high and unique loading for total Al, Mg, K, and Li (Table 1). Moderate secondary loadings also occur for organic carbon, as well as organically-bound Pb, Ni, and Cr, and total Fe. Unique moderate loadings also occur for organically-bound Fe and Mn.

The high Al, Mg, K, Li, Fe, and clay content of the sediments in this factor group and the negative loading for total Si indicates the geochemical and sedimentological character of the sediments. These are fine grained, phyllosilicate-rich sediments with a high Li content which indicates a significant influence of mica-type clay minerals (Buckley and Cranston, 1991). Sediments with a high silt and sand content also have a high Si content. Thus, it follows that the Si content of the sediment will decrease as the clay content increases.

Some organically-bound metals are associated with this sediment type. This may be caused by the relatively high loading for adsorbed organic matter on the detrital components of these fine grained sediments.

The spatial distribution of samples dominated by Factor 2 is shown in Figure 2. These samples dominate marine areas adjacent to the main surface drainage systems. The main influences are: the Sackville River and the Paper Mill Lake flume in Bedford Bay and the northern part of Bedford Basin (accounting for 50% of total fluvial discharge into Halifax inlet); Fairview Cove storm drainage in southwestern Bedford Basin; Banook Lake flume and Halifax City storm drainage in the central Harbour area; a combination of Chocolate Lake Brook, Frog Lake Brook, and Williams Lake Brook in the Northwest Arm; and Purcells Lake Brook at the southern entrance to the Northwest Arm. This pattern of factor 2 loadings clearly demonstrates the significant influence that land surface drainage has on the sedimentological and geochemical characteristics of adjacent marine sediments. It also demonstrates how fine grained, detrital particles are dispersed from point sources, with sediments from the Sackville River plume and the Banook Lake plume being dispersed southward.

Factor 3

Variables included in this factor accounted for 6% of the total variance and samples contain high loadings for all labile forms of Cu, weak acid and reducible forms of Pb, and weak acid extractable Zn. Also, moderate loadings are observed for reducible Zn, total Pb, and weak acid extractable Ni (Table 1). Samples dominated by Factor 3 (Fig. 2) are located in Bedford Basin, the central Harbour, the North West Arm, and the Harbour approaches.

Table 1. Factor analysis matrix after varimax rotation (Varimax converged in 7 iterations) and Kaiser normalization. Only factor loadings greater than 0.35 are listed.

	FACTOR 1	FACTOR 2	FACTOR 3	FACTOR 4	FACTOR 5
Zn_{org}	**0.92**	...	...	...	...
Zn_T	0.86	...	...	...	...
Zn_{HA}	0.53	...	0.36	...	...
Zn_{HHA}	0.86	...	...	...	...
Cu_{org}	0.86	...	...	...	...
Cu_T	0.84	...	...	...	...
Pb_{org}	**0.63**	**0.42**	...	...	...
Pb_T	0.75	...	0.47	...	...
Pb_{HHA}	0.83	...	...	...	...
Cd_T	0.73	...	...	...	...
C_{org}	**0.69**	**0.53**	...	...	...
Ni_{org}	0.68	0.37	...	...	...
Ni_{WA}	0.65	...	...	...	...
Ni_{HHA}	0.61	...	...	0.42	...
Cr_{org}	0.71	**0.43**	...	...	...
Cr_{WA}	0.62	...	...	...	...
Cr_{HHA}	0.65	...	...	...	...
Ca_{HHA}	0.41	...	...	...	...
Al_T	...	**0.83**	...	...	...
Si_T	-0.44	-0.79	...	...	...
Mg_T	...	**0.78**	...	...	...
K_T	...	0.73	...	...	...
Li_T	...	**0.66**	...	...	...
Fe_T	0.48	0.55	...	...	...
Fe_{org}	...	0.52	...	...	...
Mn_{org}	...	**0.42**	...	...	...
Zn_{WA}	**0.57**	...	0.58	...	...
Pb_{HA}	0.40	...	0.65	...	...
Pb_{WA}	0.46	...	0.65	...	...
Cu_{HA}	...	...	0.63	...	...
Cu_{WA}	...	...	0.63	...	...
Cu_{HHA}	...	...	0.56	...	...
Mn_{HA}	...	...	...	0.80	...
Mn_{HHA}	...	...	...	0.77	...
Mn_T	...	...	...	0.42	...
Ca_{HA}	...	...	...	0.58	0.39
Ni_{HA}	...	...	0.43	0.55	...
Ti_T	...	...	...	0.49	-0.40
Ca_T	...	...	...	...	0.78
Ca_{WA}	...	...	...	...	0.74
$CaCO_3$	...	...	...	...	0.70
MnWA	...	...	...	...	0.37

These parameters were not associated with the organic matter and elevated concentrations may have been the result of authigenic processes. Labile metal fractions included weak acid leachable metals which were bound as carbonates or oxyhydroxides, easily reducible metals which were bound to Fe-Mn oxides or oxide coatings on other minerals, and metals which were bound as mature oxides (Chester and Huges, 1967; Tessier et al., 1979; and Fitzgerald et al., 1987). Factor 3 was indicative of areas where the dominant processes favoured the accumulation of these metal fractions near the water-sediment interface after being mobilized from more deeply buried sediments. When Fe and Mn are effluxed from the sediments the change of oxidation state as well as pH precipitate them from solution as oxyhydroxides and oxide coatings on particles at the bottom interface and on suspended particles. Cu, Pb, and Zn may also be co-precipitated with the Fe and Mn. Zones containing high concentrations of these metals were observed adjacent to dump sites (i.e. the old Halifax City dump at Sea View Point).

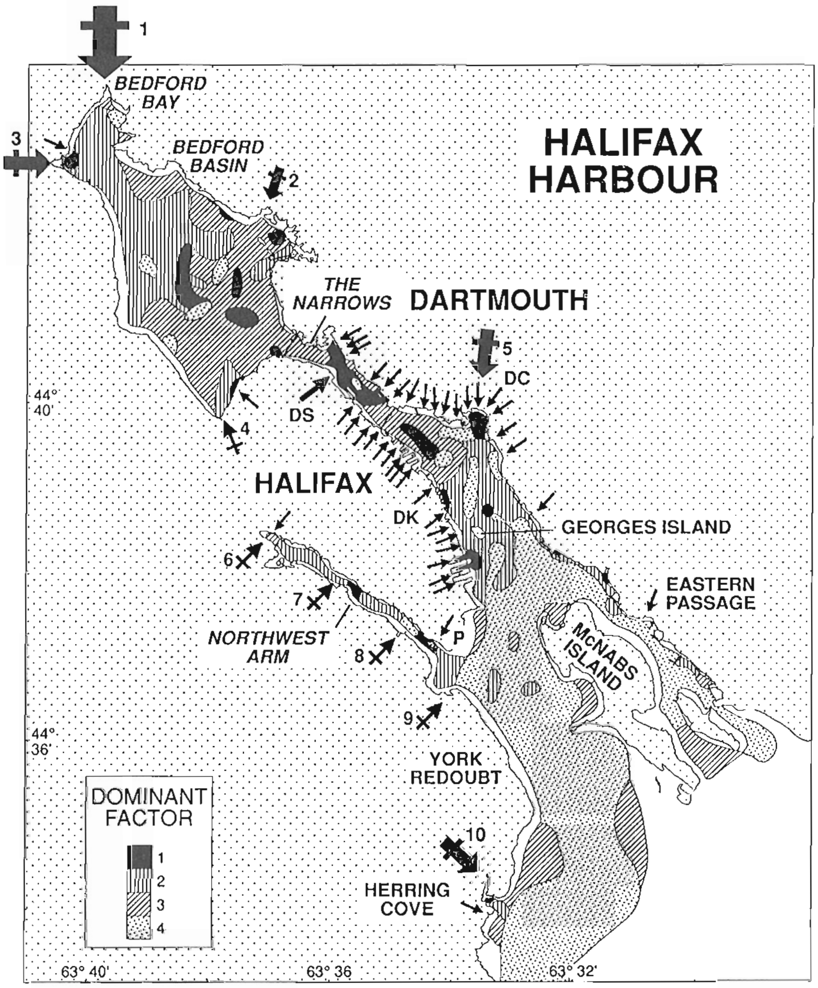

Figure 2. Spatial distribution of dominant factors. Scaled vectors represent relative magnitudes for the fluvial ("t" shaped vector) and sewage effluent inputs. Fluvial inputs are numbered from north to south and east to west: 1. Sackville River, 2. Wrights Brook, 3. Paper Mill Lake flume, 4. Fairview Cove storm drain, 5. Banook Lake flume, 6. Chocolate Lake brook, 7. Frog Lake brook, 8. Williams Lake brook, 9. Purcells Lake brook, and 10. Powers Pond flume. Some of the sewers are labelled with abbreviations: "TC" for Tufts Cove, "DC" for Dartmouth Cove, "DS" for Duffus Street, "DK" for Duke Street, and "P" for Point Pleasant.

Regression analyses of organically bound copper concentrations versus total copper concentrations reveals that 82% of the total copper in the surfical sediments was organically-bound. Paulson et al. (1991) observed that the changes in partitioning between dissolved and particulate copper (in the marine environment) suggested a rapid release of copper from organic matter simultaneously with a slower chemical re-equilibration between the dissolved and particulate phases. Fischer et al. (1984), from studies in the Pacific Ocean, speculated that dissolved copper from sediment was readsorbed on to particles near the bottom. Thus it appears that geochemical conditions at the water-sediment interface favours an elevation in the concentration of the labile forms of copper in the surfical sediments. Because of the inclusion of some labile forms of lead and zinc in this factor grouping (Table 1), it appeared that they may be subject to some of the same processes as noted for copper.

Factor 4

Variables included in this factor account for 5% of the total variance. The variables which showed high loadings on this factor (Table 1) included concentrations of Mn_{HA} and Mn_{HHA}. Total Mn was also associated principally with this factor. Other fractions including Ni_{HA}, Ni_{HHA}, Ca_{HA}, and Ti_T also belonged to this grouping. The factor scores of individual samples showed high loadings for sediments overlying the drowned river channel of the Sackville River in the outer Harbour, as well as in some sedimentary basins and zones of higher sediment accumulation. The dominant factor map shows that in the central Harbour, Bedford Basin, and Northwest Arm, Factors 1 and 2 dominate the surface sediments, although processes indicated by Factor 4 are occurring over large areas. This factor dominated (Fig. 2) the drowned river channel in the outer Harbour as well as smaller zones in the central Harbour and Bedford Basin.

Factor 4 metal fractions were representative of leachable metals which were present as Fe-Mn oxides or oxide coatings on other minerals, and mature oxides (Chester and Huges, 1967; Tessier et al., 1979; Fitzgerald et al., 1987). Oxide coatings also contained the easily and moderately reducible fraction of Ni.

Sediments in the outer Harbour had a silty-sandy texture, however in the area dominated by Factor 4, they were underlain by the estuarine muds of the relict river channel of the Sackville River (Fader et al., 1991). These buried muds contain methane gas, indicating very reducing conditions that would readily remobilized Mn and allow it to diffuse upwards towards the water-sediment interface. Oxidizing conditions at or near the interface creates an oxidation front. This results in the precipitation of manganese oxide coatings on particle surfaces (Buckley and Cranston, 1988; Buckley et al., 1989). In Bedford Basin where surficial sediments are dominated by Factor 4, the fine grained sediments are underlain by sediments containing methane gas (Fader et al., 1991). Observations in the intermediate and bottom waters of Bedford Basin (Petrie and Yeats, 1990) indicate that Mn concentrations were elevated in the bottom waters and that it was probably effluxing from the bottom sediments, which were in a reducing environment. The dissolved manganese was being oxidized in the upper layers of the water column and converted to the particulate form. Observations by Delziel et al. (1989) supported these finding. A model describing this continuous process was reported by Calvert and Price (1977) and Yeats et al. (1979) and quantified by Sundby and Silverberg (1985). These studies showed a Mn cycle in which particulate Mn was buried, reduced and mobilized, and effluxed out of the sediment into the bottom water, only to be resedimented and buried. In some areas in Halifax Harbour, this process has been continuously concentrating Mn within 1 cm of the sedimentwater interface. A similar process was also suggested by Delziel et al. (1989) to explain a 3 times enrichment of iron in suspended sediments taken from bottom water.

Factor 5

This is the least significant of all the factors and describes the variance (4% of the total variance) due to calcium carbonate concentration. Associated with this factor are total and weakacid leachable calcium, $CaCO_3$ (from inorganic carbon), and weak-acid leachable Mn (Table 1). This factor is of little significance in the central and inner areas of the Harbour where $CaCO_3$ content is generally less than 3%, and is of minor importance in the outer reaches of the Harbour, where marine shell fish are viable in shallow water areas. Weak acid leachable Mn is associated with this carbonate material.

Unique samples

Four sites have total lead concentrations varying between 5 and 10 times the mean for the inlet. The organic fractions for these samples are low, and varied between 0% and 15%. From regression analyses we found that for Halifax inlet, an average of 40% of the total lead is organically-bound (n=247, r=0.846). Because the locations of 3 of these samples are near marine facilities (a marina, a naval dockyard, and an ocean terminal) the source of these elevated concentrations of lead may be marine paints and antifouling compounds.

A sample from the west-central Bedford Basin also contains an elevated copper, concentration (at 8 times the mean for the inlet), of which only 26% is organically-bound as compared to 82% for the inlet. The cadmium concentration is also elevated (5 times the mean for the inlet); this sample was taken near a dump site that appears to contain materials rich in lead, copper and cadmium.

CONCLUSIONS

Factor analyses have been used to classify the geochemical characteristics of surficial sediments in Halifax Harbour. Although the natural and anthropogenic inputs from many sources have been altered by physical and chemical processes, the spatial effects of the dominant processes could be mapped with the aid of factor analyses. Five dominant types of sediments have been identified as resulting from depositional and diagenetic processes:

(1) Metal and organic-rich sediment having an anthropogenic origin and dominating areas immediately adjacent to main sewer outfalls, as well as some small depositional basins;

(2) Sediment derived from surface land drainage having moderate organic carbon content as compared with sediment type 1;

(3) Diagenetically altered sediment of an anthropogenic origin containing remobilized Cu, Zn, and Pb from subsurface sediments, and from solid waste contamination of surface sediments;

(4) Sediments altered by geochemical processes that are similar to type 3 sediments, however, enriched in Mn precipitate and commonly located over areas where subsurface strata are gas-charged;

(5) Sediments containing above average calcium carbonate from shell fragments; however calcium carbonate concentrations were low and this factor was of relatively little significance.

This application of factor analyses demonstrates significant advantages for environmental assessment research. In many of these types of applications there are large data bases derived from multiple types of analyses, making it difficult to identify significant or dominant processes. This example from Halifax Harbour is outstanding in that it helped us to clearly identify significant processes and impacts in a very complex system.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 890045

A new database of magnetic observations from the Arctic and North Atlantic oceans

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Macnab, R., Verhoef, J., and Srivastava, S., 1992: A new database of magnetic observations from the Arctic and North Atlantic oceans; in current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 173-178.

Abstract

Coherent and well adjusted magnetic data are crucial for quantitative plate tectonic investigations and the automated production of accurate maps. To create a digital data base that will meet these needs, a compilation team at AGC has been assembling magnetic observations from the Arctic and North Atlantic Oceans and adjacent land areas. Begun in late 1988, the project will finish in early 1993 with public distribution of grids, maps, and documentation describing constituent data sets and the procedures used to meld them into the data base. The data assembly phase of the project is now essentially complete, with enough observations acquired from many international organizations to cover most of the project area. The next phase will involve rigourous error checking of all data sets, followed by adjustment, merging and gridding. Numerous internal and external checks are built into the process to minimize errors in the final data base.

Résumé

Il est crucial de pouvoir compter sur des données magnétiques cohérentes et corrigées permettant des analyses quantitatives de la tectonique des plaques et la production automatisée de cartes précises. Pour créer une base de données numériques qui répondent à ces besoins, un équipe de compilation du CGA a rassemblé les observations magnétiques faites dans les océans Arctique et Atlantique Nord et dans les zones continentales adjacentes. Entrepris à la fin de 1988, ce projet se terminera au début de 1993, date à laquelle seront diffusés dans le grand public des quadrillages, des cartes et des documents décrivant les ensembles de données et les procédés utilisés pour les intégrer dans la base de données. La phase de rassemblement des données est maintenant essentiellement terminée et les observations obtenues de nombreux organismes internationaux sont suffisamment nombreuses pour couvrir la grande partie de la zone à l'étude. La prochaine phase consistera à vérifier minutieusement tous les ensembles de données afin d'y relever les erreurs pour ensuite les ajuster, les fusionner et les reporter sur des quadrillages. De nombreuses vérifications internes et externes sont intégrées au processus pour minimiser le nombre d'erreurs dans la base de données finale.

INTRODUCTION

Plate tectonic investigators are developing increasingly sophisticated techniques for analyzing plate motions over large regions, and for restoring ancient continents to their pre-drift configurations. Reflecting the advent of powerful yet affordable laboratory computer systems, these new approaches feature the application of advanced combinations of mathematics and software to large data sets (Verhoef et al., 1990a).

The preparation of the data sets used in these manipulations is usually a major task in itself, involving the assembly and merging of observations from many sources, with various specialized procedures to ensure coherence and homogeneity of the final database.

Since late 1988, a team at the Atlantic Geoscience Centre has been developing such a database, containing magnetic observations from the Arctic and North Atlantic Oceans and adjacent land areas (Macnab et al., 1990; Verhoef et al., 1990b). The objective of this operation is to produce an assemblage of high quality data suitable for quantitative tectonic and other related investigations, and for the automated production of accurate maps.

TIMETABLE, PRODUCTS AND REVIEW PROCEDURES

The project timetable for the 1990-92 period is illustrated in Figure 1. Final completion is scheduled for the end of 1992, at which time three sets of products will be released to the geoscientific community:

Data grids

Grids defining the regional magnetic fields of the Arctic and North Atlantic, produced from the final database after all assembled data have been adjusted and merged; where necessary, the data will be filtered to protect contributors' proprietary interests. Final grids will be distributed through established data centres.

Maps

Large scale maps produced from the final database, portraying the regional magnetic fields of the Arctic (scale 1:6 000 000, minimum latitude 64°N) and the North Atlantic (scale 1:5 000 000, from 35°N to 80°N), will be published by the Geological Survey of Canada.

Project report

A comprehensive report will describe the assembled data sets and their sources and explain procedures used in managing and processing the different data sets.

All operations will feature careful internal checking of results to ensure that errors are neither retained nor inserted in the final database through faulty procedures. As an additional precaution, external advisors have been invited to review results and procedures at key intermediate stages.

About mid-1992, organizations that have contributed data to the compilation will receive advance copies of the grids and maps for their exclusive use. These first-time users will also be asked to provide feedback on detected errors and other observed problems in the maps and database. The present intention is to convene a review meeting several months after the advance release, which will allow contributors an opportunity to comment on the database and to suggest improvements.

CURRENT STATE OF THE DATABASE

To date, the magnetic database contains over 40 million contributed observations consisting of about 1300 separate sets of digital data in the form of (a) *original observations* along ships' tracks and aircraft flight lines or (b) *grids* created at AGC or elsewhere from previously merged digital data or from digitized contour maps. So far, data have been provided or offered by 41 organizations in 15 countries.

Figure 2 illustrates the content and coverage of the database in its present condition. This shaded relief diagram was produced by plotting a preliminary grid developed from data sets merged without adjustment or re-levelling. Some idea of the density of coverage of the database is given in Figure 3, which illustrates the distribution of profiles and grids obtained to date for the North Atlantic.

Complementing the GSC's extensive holdings in the Arctic and Northwest Atlantic, major infusions of original digital observations have come from the US National Geophysical Data Center (NGDC) in Boulder, Colorado, and the Naval Research Laboratory (NRL) in Washington DC.

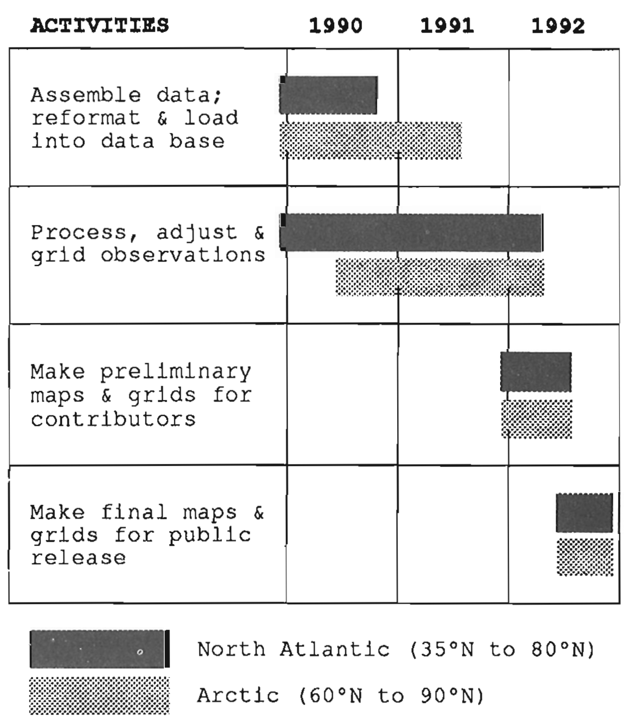

Figure 1. Project timetable, 1990-92.

Significant sets of gridded data covering North America and the Soviet Union have been obtained, respectively, from the Decade of North American Geology (DNAG) magnetic compilation (Committee for the Magnetic Anomaly Map of North America, 1987), and from NGDC (National Geophysical Data Center, 1991).

Important new compilation maps portraying magnetic anomaly contours in the central Arctic Ocean and the northern continental shelf of the USSR were provided by the laboratories of SEVMORGEOLOGIA and VNIIOkeangeologia in St. Petersburg, Russia (Shimaraev, 1990; Verba et al, 1990). Contours on these maps were digitized and converted to grids at AGC.

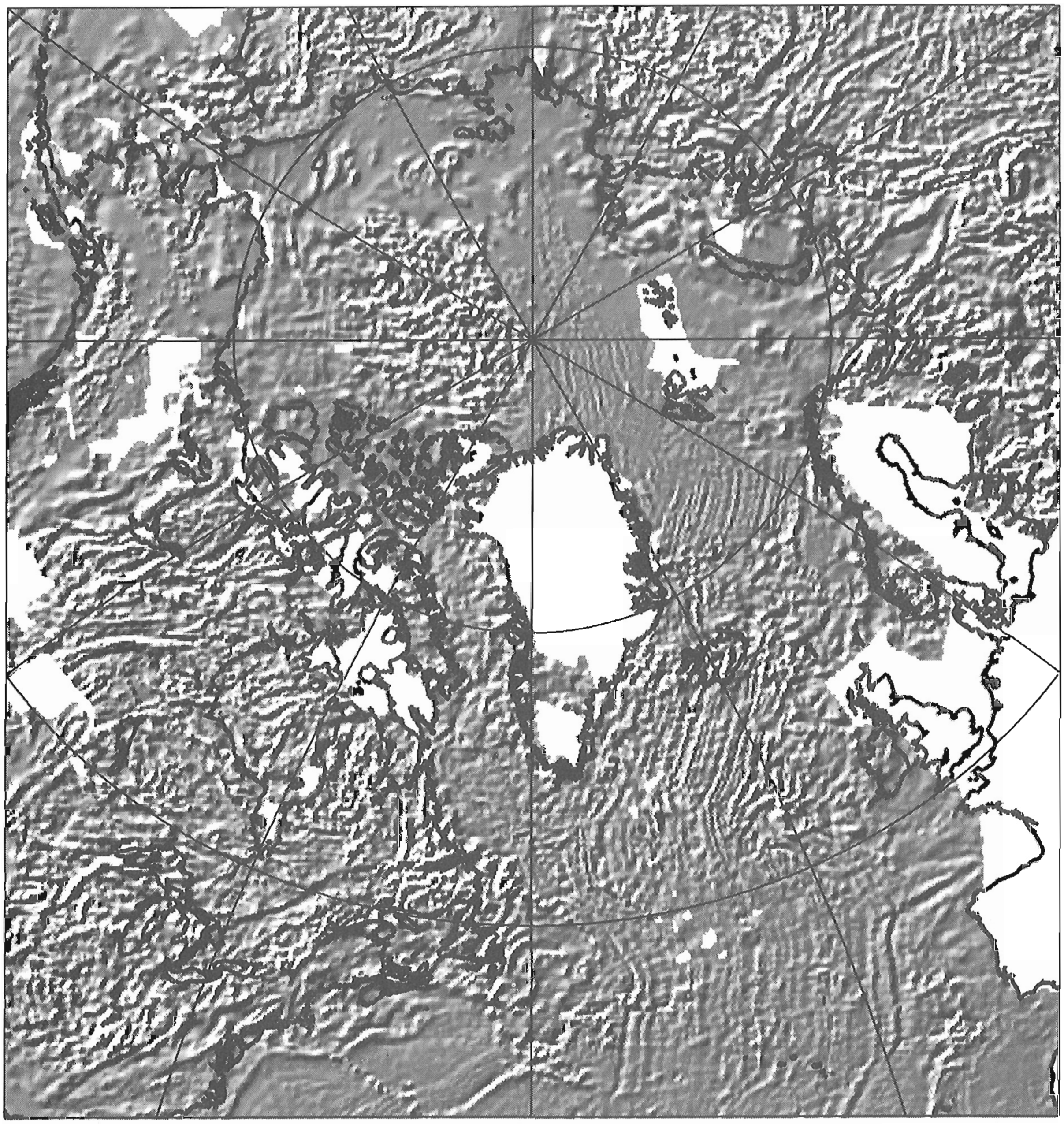

Figure 2. Preliminary shaded relief representation of the database, produced from a provisional grid developed by merging data sets without adjustment or re-levelling. Projection is Transverse Mercator with central meridian 50°W.

PROCEDURES FOR PROCESSING AND MERGING THE DATA SETS

Figure 4 outlines the major steps involved in processing and rationalizing the data sets to produce a coherent database. Initially the data are treated in three separate categories, according to their format upon arrival at AGC: (1) *grids* of observations that have been averaged and interpolated over a matrix of cells with varying and non-uniform dimensions; (2) *contours* drawn on maps; and (3) *profiles* representing original observations along ships' tracks and aircraft flight lines.

While the methods may vary according to the category of data, some procedures are common to all three. For instance, all data sets are catalogued to facilitate their management, and documented to describe the circumstances of their acquisition along with relevant technical and administrative information necessary for subsequent processing and evaluation.

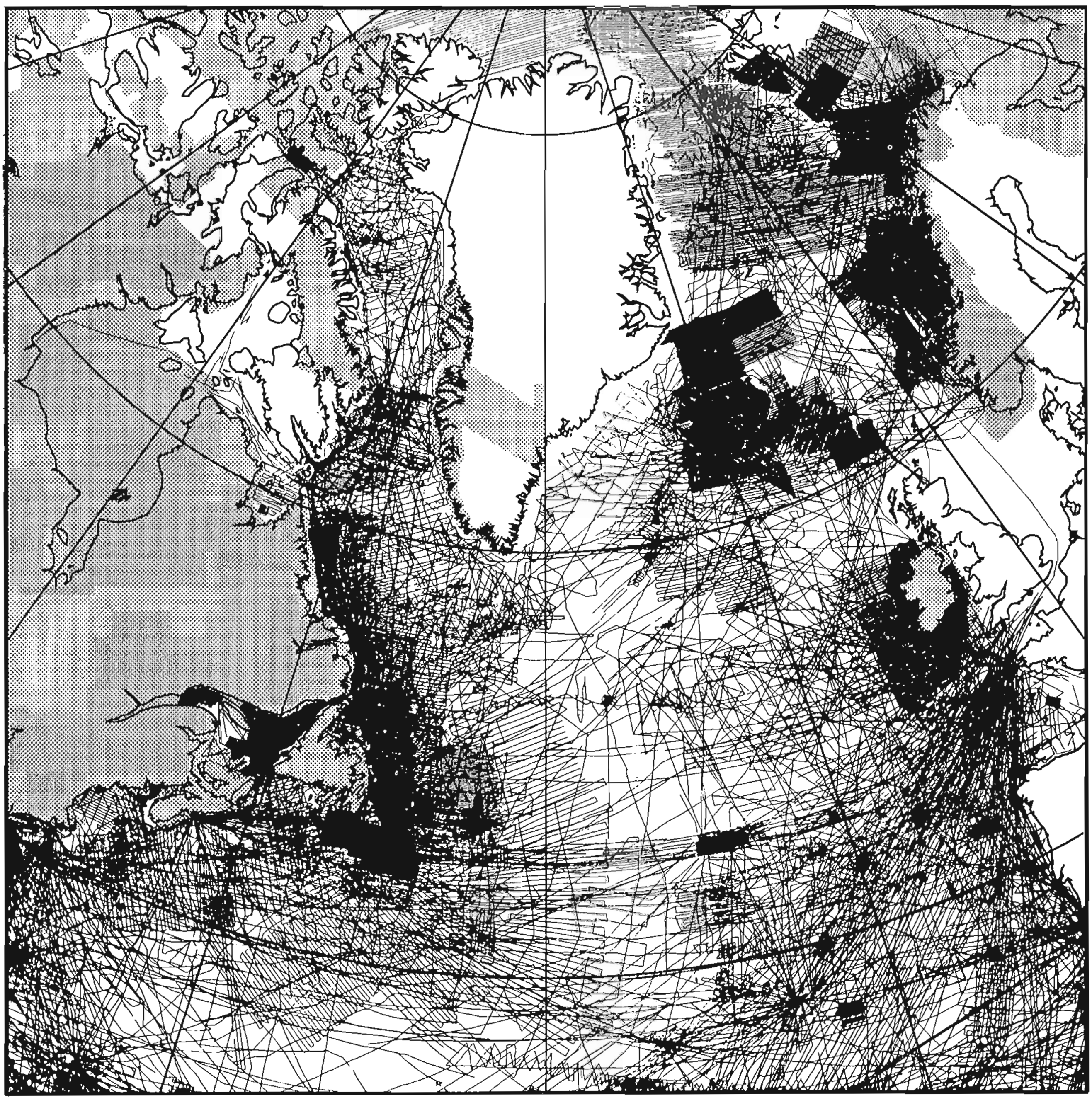

Figure 3. Data distribution in the North Atlantic component of the data base. Observations made along ships' tracks and aircraft flight lines are shown as lines, while gridded data sets are represented by shaded areas.

Grids and profiles are next converted to standard internal formats to permit processing with a uniform set of programs; for a given data set, this may be as simple as re-arranging data fields, or it may require major sorting and re-organization. Map contours are traced manually on a digitizing table to extract X-Y coordinates which are then converted to geographic coordinates by projection inversion.

Once their formats have been standardized, all data sets are subjected to preliminary quality control. Grids are scanned for singular points and other irregularities; digitized contours are re-plotted for comparison with original maps; profile data are examined to detect spikes and unlikely gradients. All errors identified during these processes are corrected.

As the profiles represent the most important fraction of the database and as they tend to be less 'refined' than the grids and contours, their analysis and adjustment consume the major part of the subsequent handling and processing. A crossover analysis is performed to check the internal consistency of the data sets by determining errors at profile intersection points and deriving correction factors that can be applied to the data sets (Verhoef and Macnab, 1987). The adjusted data are subjected to further quality control to improve the overall agreement between profiles through re-adjustment, or by discarding spurious observations; this process usually takes more than one iteration and can be extremely consumptive of computer resources on account of the large number of data points that must be manipulated.

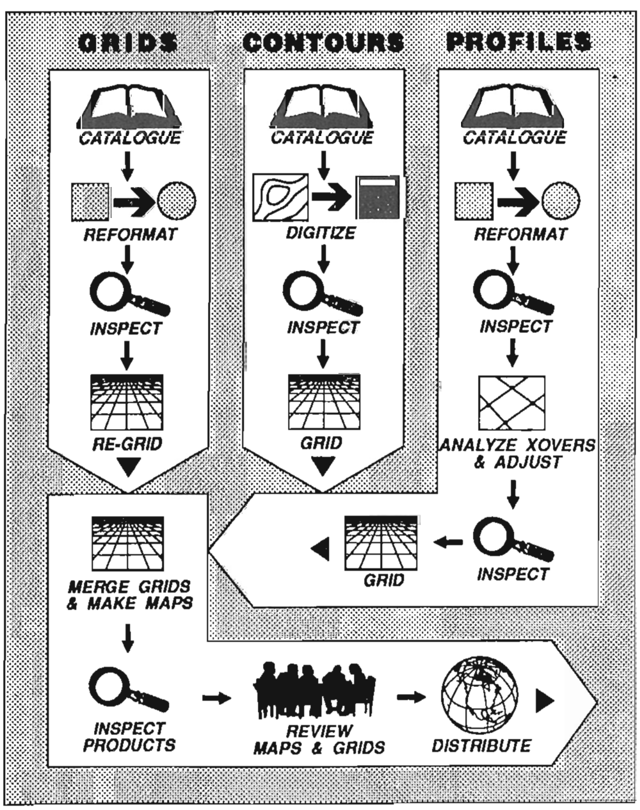

Figure 4. Outline of procedures for incorporating three categories of contributed data sets (grids, contours, profiles) into a coherent data base. The process provides for examination and correction of intermediate results at several stages.

The separate processing streams are concluded by reducing all three classes of data to a common grid. The final parameters of this grid will be determined upon examination of all data sets; for computational convenience it could consist of a single grid that covers both the Arctic and North Atlantic study regions, with nominally uniform cells throughout to minimize distortion.

Likewise, the final choice of gridding procedure will depend upon an analysis of the data sets. For profiles and contours, a common approach is to define a minimum curvature surface under tension, followed by interpolation to calculate surface elevations at pre-defined matrix intersections. For existing grids, re-sampling and filtering is often adequate, but some cases may require special treatment to compensate for biases or other irregularities stemming from the original data distribution, or from prior handling.

When all data sets have been reduced to a common grid format, they will be combined with appropriate adjustments, including levelling and tapering, to minimize edge effects and multiple wavelength contaminations. Together with prototype magnetic anomaly maps, copies of this first combined grid will be distributed to data contributors for review and assessment; their feedback will be weighed to determine revisions and improvements that have to be incorporated in the final grid and maps prior to their public distribution.

RELATED ACTIVITIES

The data compilation project serves as a focus for other initiatives related to the analysis and interpretation of magnetic observations. For the most part, these additional tasks involve investigators from other institutions, with interests that are complementary to ours.

For example, the merging of Arctic multi-year data sets requires an accurate method for modelling secular variations in a region where magnetic observatories are sparsely distributed. The technique of spherical cap harmonic analysis (Haines, 1985) has been suggested as an effective solution to the problem; this has led to an investigation of its applicability, and to preparatory work by staff of the GSC's Geophysics Division.

Approached as an assemblage of marine magnetic observations from the past several decades, the compilation's database has been analyzed in order to extract quantitative information describing secular variation in the open ocean (Williams et al., 1991). This provides a useful tool for evaluating the effectiveness of the International Geomagnetic Reference Field (IGRF) in modelling the time-varying component of the main geomagnetic field.

The ultimate objective of the compilation project is to develop a data resource for scientific analysis. Already, a number of joint investigations have focussed on portions of the database and their significance in terms of regional geological structures in several regions: the Celtic Sea (Murphy, 1991); the Norwegian and Barents Seas (Skilbrei et al., 1991); the Arctic Ocean (Shimaraev et al., 1991; Srivastava et al., 1991); the North Atlantic (Bocharova et al., 1991; Roest et al., in press); and the USSR (Zonenshain et al., 1991). Further studies are in progress or planned.

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Geological Survey of Canada Project 890041

Quaternary seismic stratigraphy of the inner shelf region, Eastern Shore, Nova Scotia

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Stea, R.R., Fader, G.B.J., and Boyd, R., 1992: Quaternary seismic stratigraphy of the inner shelf region, Eastern Shore, Nova Scotia; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 179-188.

Abstract

Initial mapping of seismic profiles from a transect across the inner shelf in the Sheet Harbour area revealed zones of distinct acoustic surface morphology and seismic sequences. These include (landward) : 1. Eastern Shore Moraine 2. Zone of Basin Fill- Emerald Silt and LaHave Clay 3. Truncated Emerald Silt 4. Minor Moraines 5. Inter-Moraine Basin 6. Zone of Mounds-Ridges 7. Zone of Valleys 8. Zone of Bedrock. Zone 6 has a unique acoustic morphology not yet described on the inner Scotian Shelf. It is characterized by steep, irregular mounds or ridges 5-25m in height and 100 to 300m wide scattered in depths of 80 to 115m. South of Sheet Harbour the seabed at the same depth is denuded of Quaternary cover. We interpret the seismic sequences and zones in the Sheet Harbour transect to represent ice marginal deposition from a landward-retreating ice mass, punctuated by stillstands or slight readvances. On land, ice centres were changing from an ice divide which straddled the axis of the Nova Scotia peninsula (Scotian Ice Divide) to remnant ice caps largely confined to northern Nova Scotia. Sea level lowering of at least 70m followed ice retreat from the region.

Résumé

La cartographie initiale des profils sismiques d'un transect traversant la plate-forme interne dans la zone de Sheet Harbour a révélé des zones de morphologie de surface et des séquences sismiques acoustiquement distinctes. Ce sont notamment (vers le continent) : 1. la moraine d'Eastern Shore 2. la zone de remplissage de bassin – le silt d'Emerald et l'argile de LaHave 3. le silt d'Emerald tronqué 4. les moraines secondaires 5. le bassin inter-morainique 6. la zone des monticules et crêtes 7. la zone des vallées 8. la zone du socle. La zone 6 présente une morphologie acoustique particulière non encore décrite sur la plate-forme Néo-Écossaise interne. Elle est caractérisée par des monticules irréguliers abrupts de 5 à 25 m de hauteur et de 100 à 300 m de largeur disséminés à des profondeurs de 80 à 115 m. Au sud de Sheet Harbour, le fond marin à la même profondeur n'est pas recouvert de sédiments quaternaires. Selon la présente interprétation, les séquences et les zones sismiques dans le transect de Sheet Harbour représentent une sédimentation proglaciaire par un glacier en recul vers le continent, dont le déplacement a été ponctué de positions stationnaires et de légères réavancées. Sur le continent, la nature des centres glaciaires s'est modifiée, passant d'une ligne de partage glaciaire qui chevauchait l'axe de la presqu'île de la Nouvelle-Écosse (ligne de partage Néo-Écossaise) à des calottes glaciaires résiduelles en grande partie confinées au nord de la Nouvelle-Écosse. Un abaissement du niveau de la mer d'au moins 70 m a suivi le recul glaciaire dans la région.

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INTRODUCTION

This report provides an initial description and interpretation of the seismic stratigraphy on a transect of the inner continental shelf of Nova Scotia, Canada. The transect extends from the Eastern Shore Moraine (King and Fader, 1986) to Sheet Harbour, 90 km northeast of Halifax. This contribution is part of a seismic mapping and sampling program whose objectives include: 1) establishment of lithic and seismostratigraphic criteria for the recognition of equivalent terrestrial till units in the offshore; 2) determination of the offshore extent of ice flow phases identified on land by glacial erosional features, drumlins and erratic dispersal; 3) sampling and dating of the Quaternary units to establish a temporal framework for glacial events in the region; 4) identification of sedimentary units in the offshore by seismic, biostratigraphic, compositional and geotechnical properties for future engineering and mineral resource studies; 5) determination of the extent of the Holocene transgression across the inner shelf, including low sea level stands, residual beaches and erosional unconformities; and 6) evaluation of the placer potential and aggregate resources of the inner Scotian Shelf. The research is co-sponsored by the Atlantic Geoscience Centre, Geological Survey of Canada, and Dalhousie University Centre of Marine Geology (Ph D. Thesis-R. Stea) and funded through the Canada-Nova Scotia Cooperation Agreement on Mineral Development 1990-92, Dalhousie University, ESSO Resources Canada Limited, and an EMR-NSERC research agreement.

This preliminary assessment will augment previous studies of the inner shelf (eg. Piper et al., 1986; Hall,1985; Forbes et al., in prep.) and provide data to better integrate the terrestrial and marine stratigraphic models (Grant and King, 1984; Stea et al., in press). The regional distribution of surficial formations in this area of the inner shelf was originally mapped by King (1970).

The Huntec Deep Tow high resolution seismic system was used to generate reflection profiles. Seismic reflection data of lower resolution but greater penetration were obtained using a sleevegun system. Bathymetric data were obtained using a hull mounted 3.5 KHz acoustic profiler and previously published charts. Seafloor topography/reflectivity were obtained with BIO sidescan (long range 1-2 km) and Klein sidescan sonar systems (shortrange high-resolution). Cores of the seafloor were obtained with the newly-designed AGC vibracorer with a 3m barrel. Seafloor samples were obtained with a large volume IKU grab.

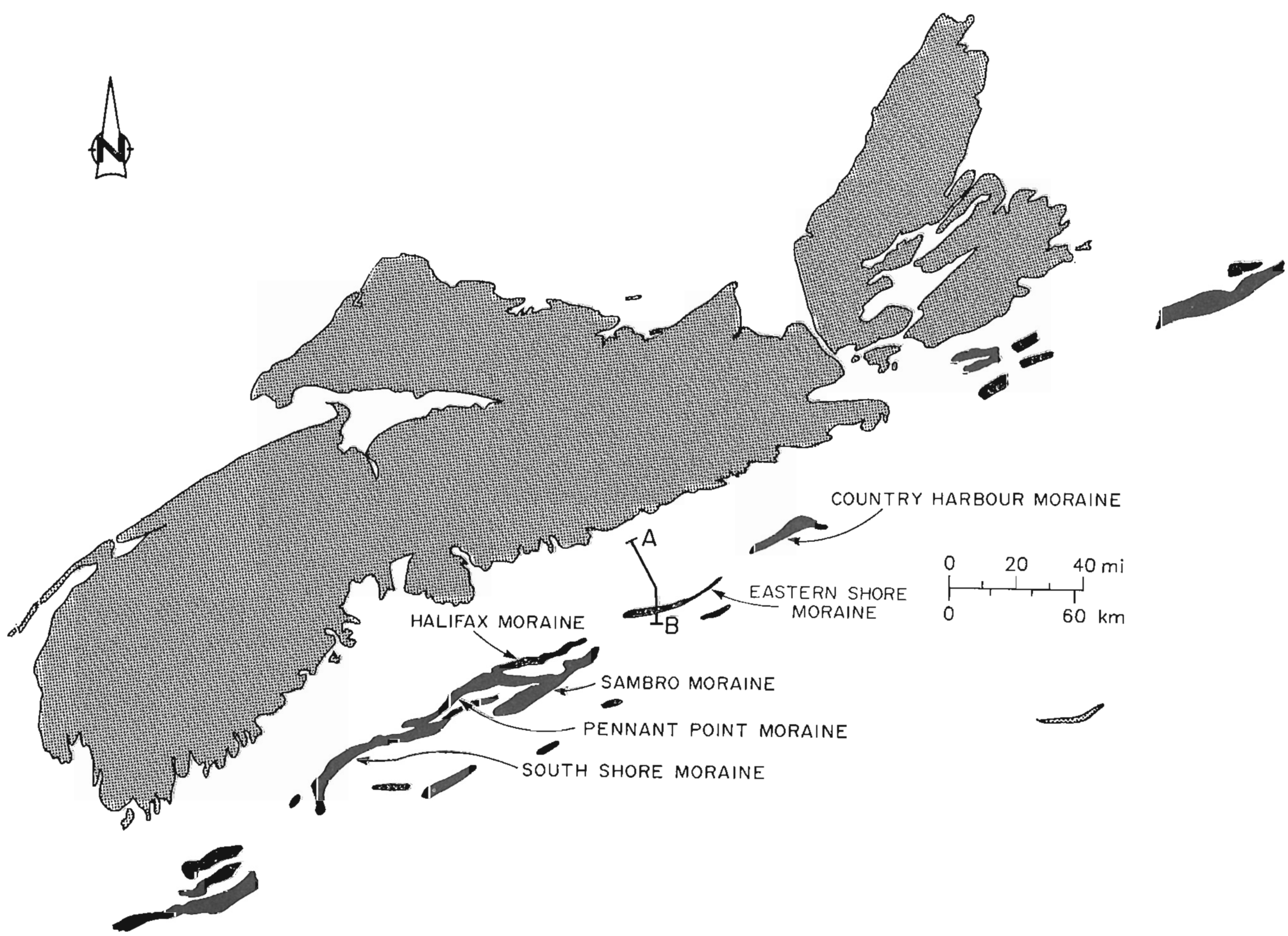


Figure 1. Location of the seismic transect and moraines of the inner shelf region (after King et al., 1972).

SEISMOSTRATIGRAPHIC MAPPING

Stratigraphic analysis of seismic profiles is based on seismic sequence stratigraphy (Vail, 1987). The method involves recognition of unconformity-bound packages of strata termed seismic sequences. Boundaries between seismic sequences are detected on seismic records by terminations of acoustic reflections. The boundaries are either erosional, lapout (hiatal) or conformable. The seismic properties of the reflections within each sequence (continuity, amplitude/frequency) are an indication of the lithic properties of the strata. Initial mapping of the seismic profiles from the Sheet Harbour area revealed zones of distinct acoustic surface morphology and seismic sequences. Forbes et al. (in prep.) also recognized the presence of cross-shelf zones on the inner Scotian Shelf off Halifax. The zones are a result of glacial erosional and depositional processes and sea level transgression and regression during the Holocene. Eight zones were defined across the Sheet Harbour transect (Fig. 1, 2). From south to north these are:

1. Eastern Shore Moraine
2. Zone of Basin Fill- Emerald Silt and LaHave Clay
3. Truncated Emerald Silt
4. Minor Moraines
5. Inter-Moraine Basin
6. Zone of Mounds-Ridges
7. Zone of Valleys
8. Zone of Bedrock

1. Eastern Shore Moraine

The Eastern Shore Moraine is part of the Scotian Shelf Moraine System (King et al., 1972). These moraines are large ridges that occur 30-40 km offshore and parallel the present day Nova Scotian coastline (Fig. 1). The Eastern Shore moraine is 5-7 km wide and 45 to 70 m high. The surface of the moraine is undulatory and furrowed. It has a marked asymmetry, thicker and steeper on the north (landward) side and thinner on the south (Fig. 2). The moraine consists of acoustically incoherent material, in contrast to the well-defined, coherently-laminated, acoustic units adjacent to the moraine (Emerald Silt). On the moraine's seaward side, a wedge-shaped projection (till-tongue) interdigitates with the laminated unit. Below the till-tongue the Emerald Silt appears to grade into the acoustically massive unit. On the seaward side above the till tongue, and on the landward side the acoustic laminae are draped on the moraine. IKU grab samples from the top of the moraine revealed an olive-grey sandy diamicton, with angular to subrounded clasts. The pebbles are predominantly metagreywacke of the Cambro-Ordovician Meguma Group (>90%) with a minor granitoid component. Some clasts are striated and facetted with flat-iron shapes indicative of glacial erosion. The surface of the moraine is armoured with a boulder-cobble lag.

2. Basin Fill

Zone 2 consists of a wide basin landward of the Eastern Shore Moraine. The basin is 7 km wide and occurs in water depths greater than 145 m. Five seismic units were recognized above acoustic basement interpreted as Meguma Group bedrock. The lowest unit (Zone 1; Unit 1; Fig. 2) consists of hummocky reflections of moderate continuity and low to moderate amplitude. These reflections are draped over the underlying basement topography and conformably overlain by an acoustic unit (2) with higher reflector continuity and amplitude. There is some evidence of an onlap relationship between the two units towards the basin margins. Infilling broad hollows in unit 2 is a unit of moderate reflector amplitude and continuity (Zone 1, Unit 3; Fig. 2).

There is clear evidence of onlap/fill relationships between units 2 and 3. These units drape over the underlying topography, to a lesser degree for each successively younger unit. The total thickness varies from 12 to 14 m.

Units 1 and 2 are equivalent to the glaciomarine Emerald Silt facies "A" of King and Fader (1986), whereas unit 3 is clearly equivalent to facies "B". These units are truncated by a major unconformity (Fig. 2). Lying in undulations on the unconformity is a 1-2m thick unit with horizontal reflections of moderate to high amplitude (unit 4). This is overlain by 3-7 m of material with low reflectivity (unit 5). Units 4 and 5 are equivalent to the post-glacial LaHave Clay described by King (1970).

3. Truncated Emerald Silt

This zone consists of units 1, 2 and 3 (Emerald Silt) truncated by an erosional surface at a depth of 143m (Fig. 2). About 5 to 10 m of sediment appear to have been removed. Several channels were cut into the remaining eroded Emerald Silt (Zone 3, unit 1; Fig. 2) and later infilled with units 4 and 5 (LaHave Clay).

4. Minor Moraines

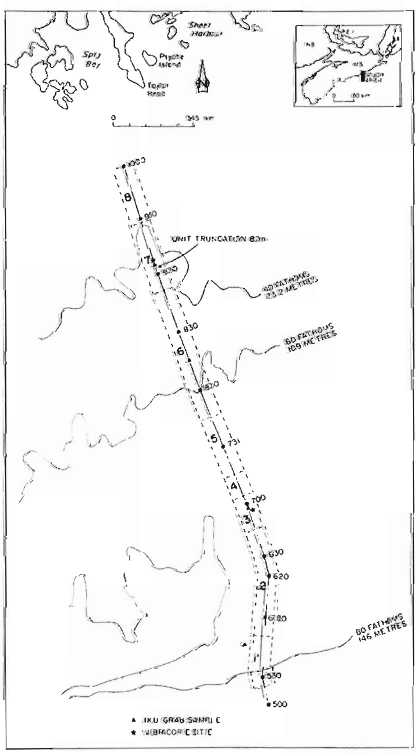
This zone includes several mounds of acoustically incoherent material. The largest feature (moraine) is 30 m high x 150 m wide with eroded Emerald Silt on its seaward side (Fig. 2). It is found in 124-135 m water depth. A till-tongue extends from the thickest and steepest part of the moraine into units 1 and 2. Unit 2 appears to drape over the tongue. Emerald Silt decreases in thickness from 11m on the seaward side of the feature to 6m on the landward side. Sidescan records show a high intensity backscatter suggesting a gravel-boulder cover. A core was collected within the Emerald Silt just south of the till tongue.

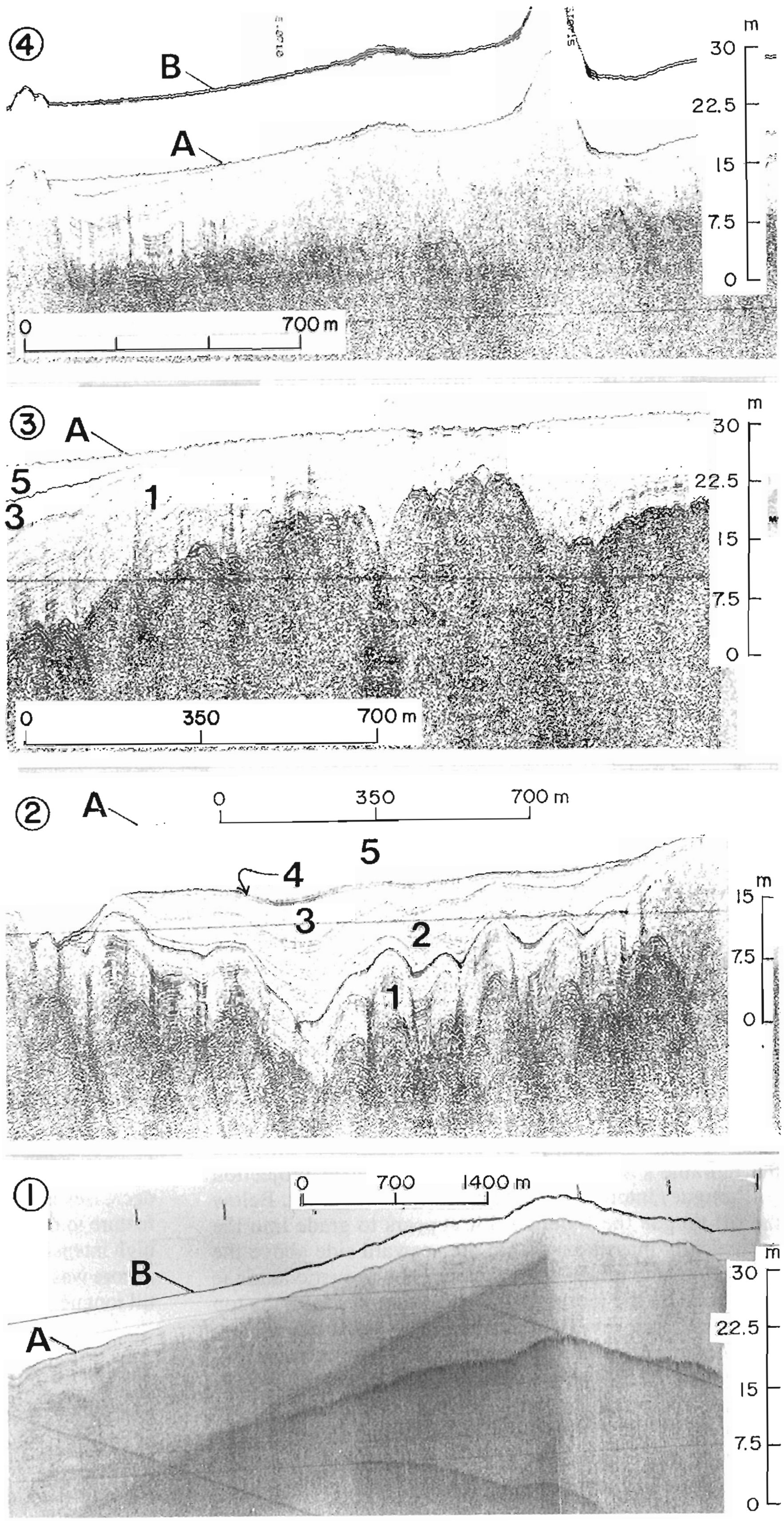
5. Inter-Moraine Basin

Units 1 and 2 (Emerald Silt) pinch out on the landward side of the "moraines" (Fig. 3). Where acoustic basement lies close to the seabed the Emerald Silt is truncated by an unconformity at the sea floor. Unit 5 (LaHave Clay) overlies Emerald Silt in basins between topographic highs. The zone

Figure 2

Detailed map of the transect area and photos of the Huntec seismic profiles within each designated zone. Circled numbers 1-8 refer to the seismic zones. The surface reflectivity line is designated as "B". The seabed trace marked as "A". Seismic sequences 1-5 are marked on the photos. In Zone 7 the arrow marks the contact between units 1 and 2.





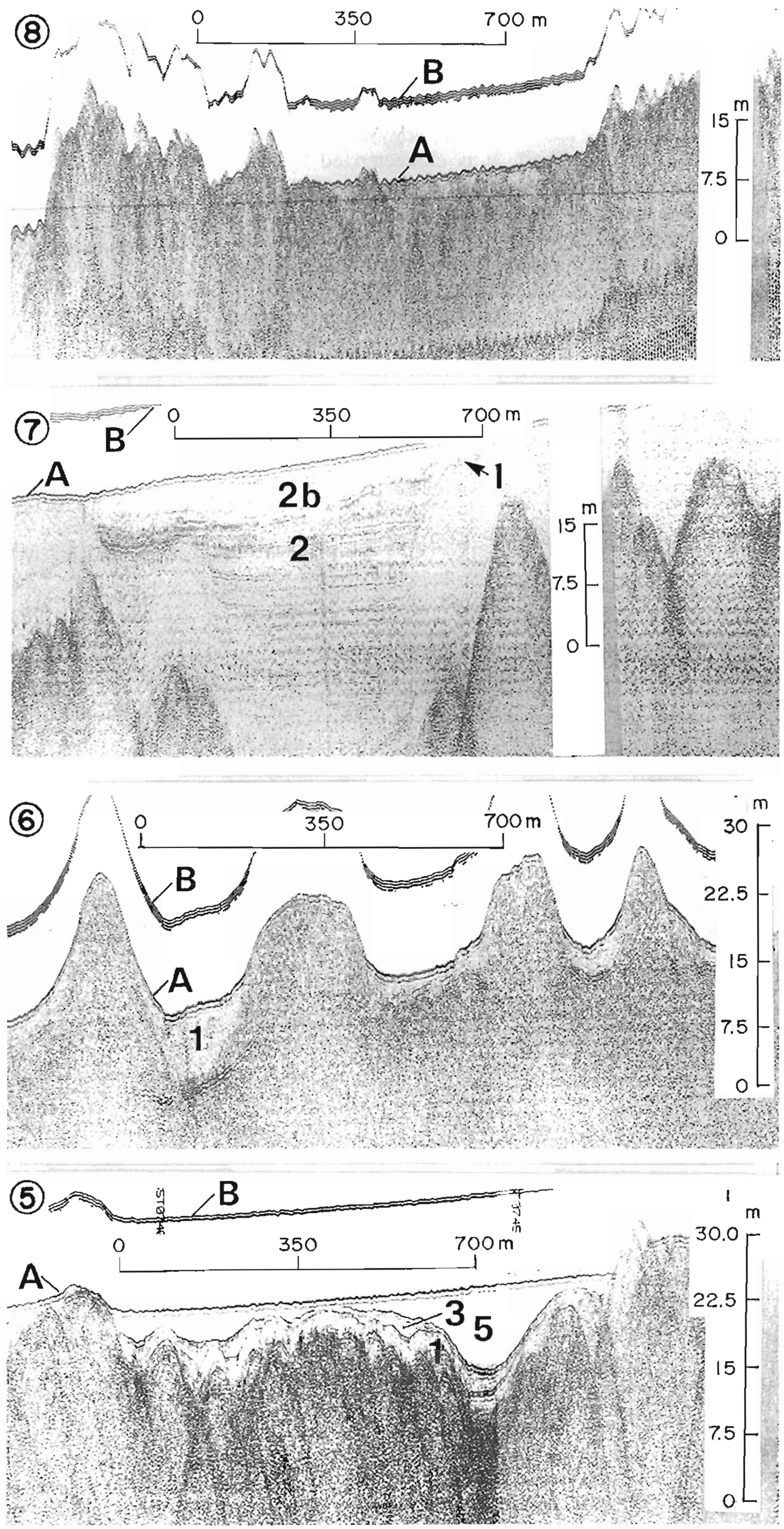
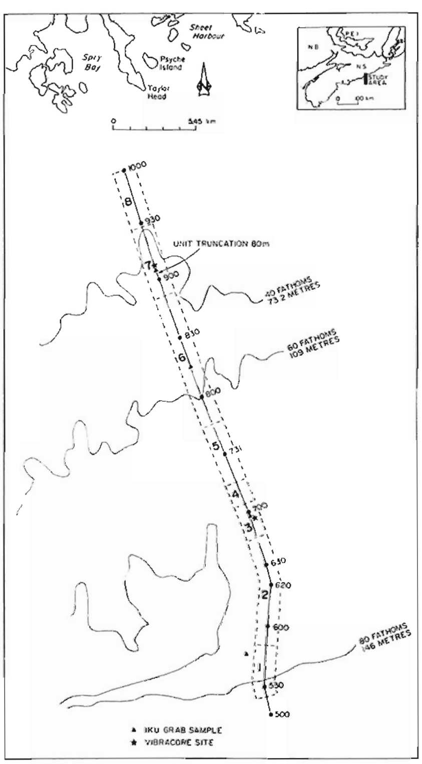


Figure 3

Detailed map of the transect area and photos of the Huntec seismic profiles within each designated zone. Circled numbers 1-8 refer to the seismic zones. The surface reflectivity line is designated as "B". The seabed trace marked as "A". Seismic sequences 1-5 are marked on the photos.



ranges from 137 to 124 m water depth. At the boundary between zones 4 and 5, unit 5 (LaHave Clay) pinches out abruptly against a rise in acoustic basement (Fig. 3).

6. Zone of Mounds-Ridges

This zone has a unique acoustic morphology not yet described on the inner Scotian Shelf. It is characterized by steep, irregular mounds or ridges 5-20 m in height and 100 to 300 m wide (Zone 6; Fig. 3). They are scattered throughout a zone which ranges in depth from 80 to 115 m. These features may be linear ridges or hummocks; further analysis of data from a sidescan sonogram mosaic across the area will address this problem. The "mounds" are acoustically homogeneous (Fig. 3). Reflection traces cannot be resolved within the "mounds". The sleevegun record (Fig. 4), however, shows a series of reflections which appear to continue through the features.

Between the "mounds" are two acoustic units (units 1, 2; Zone 6; Fig. 3, only unit 1 is illustrated). Unit 1 consists of low amplitude, draped reflections truncated by an onlap fill unit (2) of moderate to high amplitude. The lateral relationships between the strata in unit 1 and the mound materials are uncertain.

Sidescan sonograms of the mounds reveal high intensity backscatter (gravelly-bouldery) areas of irregular shapes. Boulders can be seen scattered randomly throughout. A grab sample of this unit produced subangular to subrounded cobbles and boulders, in an olive-grey muddy-sand matrix.

7. Zone of Valleys

This zone consists of a series of valleys and channels incised into acoustic basement (Meguma Group bedrock; Fig. 3). The valleys are 700 to 1000 m wide and 40 to 70 m deep. The zone ranges from 80 m to 65 m water depth. Two main units make up the valley fill (Zone 7; units 1, 2; Fig. 3). Unit 1 consists

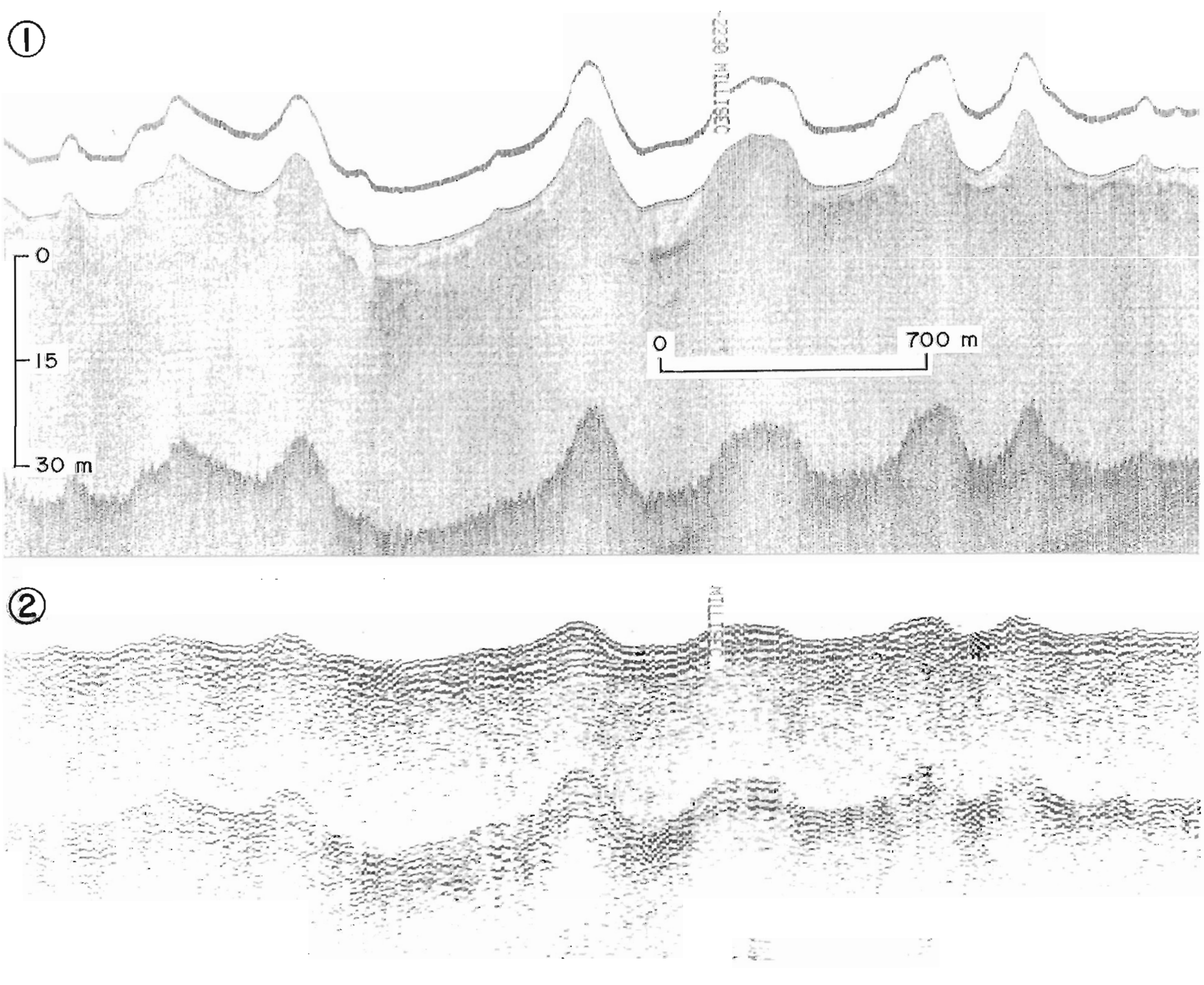


Figure 4. Comparison of the Huntec seismic profile (1) and the equivalent sleevegun record (2) in the Zone of Mounds-Ridges (Zone 6). Note that reflections appear to continue through the mounds in the low-frequency record (2). Some of these reflections, however, are multiples of the "bubble pulse".

of conformable, draped reflections of low amplitude and continuity which resemble the deeper water Emerald Silt (Zone 1; unit 1; Fig. 2). Channels and valleys lined with unit 1 are filled in an onlap relationship with horizontal reflections of moderate to high amplitude (unit 2). Unit 2 grades upward to a more indistinctly stratified seismic unit (2b). The valley fill units are truncated by the sea floor at the boundary between zones 7 and 8.

Grab samples were obtained (Fig. 3) near the contact with Zone 8. Unit 1 revealed an olive-grey sandy diamicton with abundant metagreywacke clasts. A sample taken near the valley-fill truncation zone (Fig. 3) produced well-rounded gravel with discoid clasts in a well sorted sand matrix.

8. Bedrock Zone

This zone is characterized by ubiquitous bedrock outcrop at the seabed. Distinct ridges are visible on sidescan sonograms. The hollows between ridges are filled with gravel and/or till.

INTERPRETATION

The Eastern Shore Moraine is part of the main Scotian Shelf Moraine System. These moraines were originally interpreted as buoyancy-line deposits relating to the Late Wisconsinan margin (King et al., 1972), and later reinterpreted as a grounded ice shelf deposit (King and Fader, 1986). The asymmetry of the Eastern Shore Moraine (steep landward side), landward onlap and distal interfingering of the Emerald Silt suggests that it was deposited near the margin of an ice sheet. The ridge represents an accumulation of debris while the glacier was grounded for a period of time. The moraine's steep landward slope probably reflects deposition in contact with the ice. On the seaward side of the moraine, subaqueous fan and mass flow deposition and surging processes can be expected to form interdigitating diamicton and laminated glaciomarine deposits (Boulton, 1986). Till tongues found on the seaward side may have formed as a response to grounding line fluctuations (King and Fader, 1986). Barnett and Holdsworth (1974) suggested that subaqueous, asymmetric moraines (similar in shape to the Eastern Shore Moraine) form at the grounding line in a wedge-shaped space between the lake bed and an extensive ice ramp in water depths greater than 40 m.

The moraine material appears similar in texture and composition to stony ground moraine of the Eastern Shore called the Beaver River Till (Grant, 1989). This till was deposited by ice stemming from the Scotian Ice Divide in Nova Scotia (Stea et al., 1989; Phase 3; Fig. 5). The main characteristics of this diamicton are local derivation, clast angularity and a high clast/matrix ratio. Tills formed by earlier ice sheets (Lawrencetown and Hartlen Tills; Fig. 5) are found in drumlin fields along the Eastern Shore region (Stea and Fowler, 1979). These have significant percentages of erratics derived from source regions in the North Mountain, Cobequid and Antigonish Highlands 50-100 km to the north (Stea et al., 1989). The Lawrencetown Till is distinctly reddish in colour whereas all samples of the moraines in the offshore appear olive-grey. Some samples, however, did reveal red zones and lenses in the matrix associated with higher percentages of allochthonous clasts. These may represent the reworked component of earlier tills.

If the correlation with the Beaver River Till is valid then it would suggest that flow from the Scotian Ice Divide (Fig. 5) overrode or formed these moraines and was more extensive than generally believed (Piper et al., 1986; Grant and King, 1984, Grant, 1989). This interpretation is consistent with striation data that implies extensive flow from the Scotian Ice Divide across highland regions in northern Nova Scotia above an elevation of 200-300 meters (Stea and Finck, 1984). Offshore seismic evidence links the ice divide with ice streams adjacent to the Cape Breton Highlands (Heiner Josenhans pers. comm, 1990).

The conformable basin fill units (Zone 2; units 1, 2, 3; Fig 2) are equivalent to the Emerald Silt of King and Fader (1986). King et al. (1991) interpreted the onlap-fill relationships of the Emerald Silt facies as intermittent deposition between rapid retreats and stillstands of a former ice margin. Unit 1 was probably deposited close to the former ice margin. Coarser and more rapid sedimentation, proximal to the ice margin, may explain the lower continuity of unit 1 relative to the younger units. The zone of minor moraines is a strong indication of a still-stand phase of the retreating glacier. Behind these moraines Emerald Silt thins landward and is truncated by an erosional unconformity at the sea floor.

Zone 6 (mounds-ridges) is perhaps the most difficult zone to interpret. The acoustically incoherent zone within the topographic highs may be an artefact rather than the acoustic expression of the true nature of the sediments. Continuation of reflections through the landforms on the sleevegun record implies some continuity of deposition (Fig. 4). The landforms resemble the lift-off moraines of King and Fader (1986) without the cover of Emerald Silt and LaHave Clay. Lift-off moraines are thought to be till-cored but there is no sample data to corroborate this interpretation (King and Fader, 1986, p. 53). The landforms in Zone 6 may represent a transition between grounded ice and ice proximal glaciomarine sedimentation. These features may form by rain-out and slump rather than horizontal accretion. Barnett and Holdsworth (1974) describe irregular landforms analogous to the "mounds" that are produced by this process (Fig. 6).

The mechanism of erosion which truncated the Emerald Silt at depths as great as 140 m is uncertain. A substantial thickness of overlying deposits was removed. Sea level lowering (Fader, 1989) would not appear to be a factor (Fader, 1989) because uneroded depositional features (Zone 6) are found at much shallower depths. Glacial erosion is another possibility, but one might expect an overlying acoustically incoherent unit (diamicton). King and Fader (1988) reported till overlying the truncated surface of the Emerald Silt north of the Country Harbour Moraine. The ice readvance that deposited the minor moraines may have overridden and eroded the Emerald Silt, but more likely, the erosion was accomplished by enhanced proglacial meltwater flow.

Evidence of sea-level lowering can be seen near the landward end of Zone 7. The valley fill units are truncated and there is a marked increase in surface reflectivity. A grab

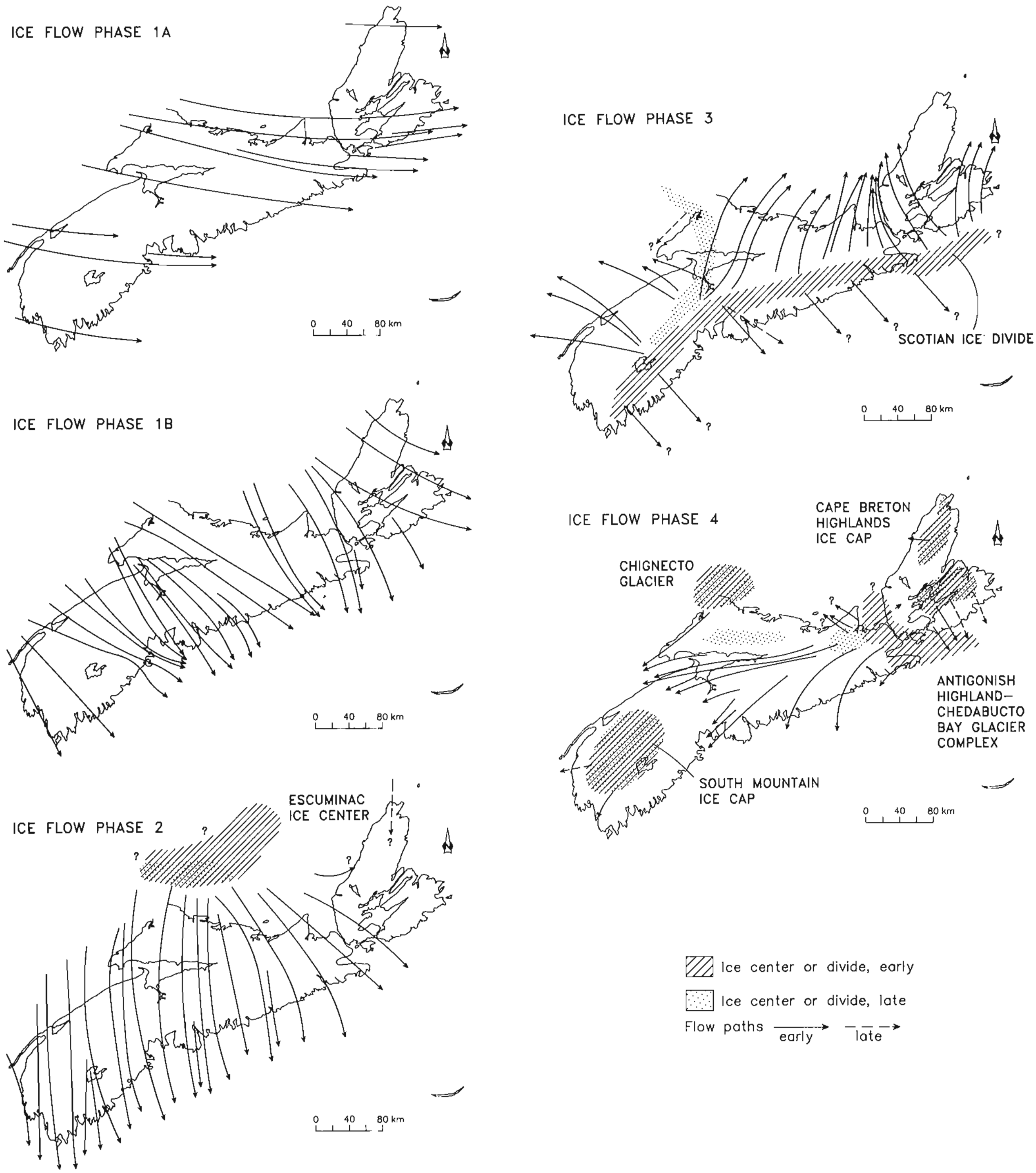


Figure 5. Ice flow phases in Nova Scotia. Phases 1a, 1b, stem from Appalachian/Laurentide ice centres and formed the Hartlen Till on the Eastern Shore. Phase 2 emanates from the Escuminac Ice Centre situated over a vast area of Carboniferous red-beds. The Lawrencetown Till, red in colour, was formed during this ice flow event. During Phase 3 an extensive ice divide formed across Nova Scotia. Ice flow was northward and southward from this divide. The Beaver River Till was deposited during this phase.

sample at 70m revealed discoid, well-rounded clasts indicative of a beach zone. The inferred relative sea level history from the acoustic facies is one of gradually falling sea levels during ice retreat until portions of the inner shelf become subareally exposed. In the Halifax inner shelf region, Forbes et al. (in prep.) noted that basin fill sedimentation

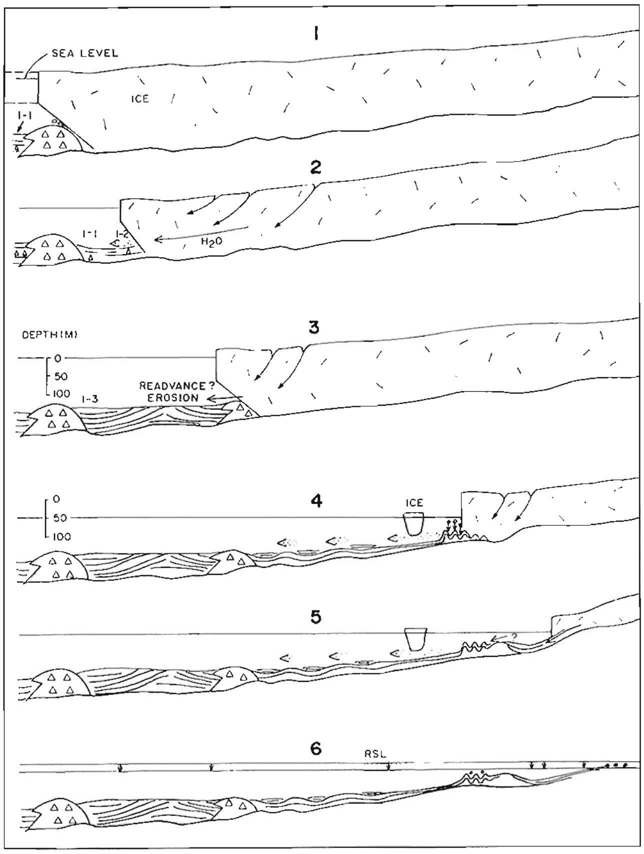


Figure 6. Pictorial summary of the proposed deglaciation and sea level events.

Stage 1: Ice ramp front at the Eastern Shore Moraine, relative sea level high.

Stage 2: Landward retreat of the ice front. Debris is fed to the glaciomarine environment by meltwater emanating from the glacier sole. Formation of units 1, 2 (Emerald Silt) landward of the Eastern Shore Moraine.

Stage 3: Stillstand or readvance of the ice front to form the minor moraines (Zone 5). Erosion of the Emerald Silt in the ice proximal area (Zone 3). Deposition of unit 3 in the deeper basin.

Stage 4: Retreat of the ice front to Zone 5. Formation of the mounds-ridges in shoaling water (depths 40m?) and synchronous deposition of unit 3 in the adjacent basinal areas (Zone 4).

Stage 5: Further retreat of the ice front with relatively high RSL. Glaciomarine? deposition in the valley zone.

Stage 6: Drop in sea level to 70 m. Subareal exposure of glaciomarine sediments; formation of beach zone.

styles changed from glaciomarine to glaciofluvial and concluded that the ice margin surfaced sometime during retreat in the Halifax area.

There appears to be significant differences in the character of the inner shelf near Halifax (Forbes et al., in prep.) and Sheet Harbour. The outcrop zone of Forbes et al. (in prep.) extends offshore in water depths from 70 to 100 m. At these depths off Sheet Harbour, acoustic basement is overlain by thick morainal deposits in Zone 6. Forbes et al. (in prep.) suggested that the outcrop zone formed by wave reworking and erosion on the emergent shelf during deglaciation. If this model is correct then extensive morainal deposits at similar depths in the Sheet Harbour area imply the persistence of ice in that area.

TIMING

The timing of glacial events in the inner shelf regions is not well constrained because of the lack of direct dates on glacial sediments. Forbes et al. (in prep.) estimated that ice retreat and sediment deposition in the Halifax inner shelf zone occurred in the period between 15 and 13 ka. Piper and Fehr (1991) determined that a "Late Till" was deposited before 14.3 ka in the inner shelf off Lunenburg. Till-tongue deposits along the northern margin of Emerald Basin have a radiocarbon age of 15-15.5 ka (King and Fader, 1988; Gipp and Piper, 1990). Stea and Mott (1989) estimated that the ice retreated to the present coastline by 13 ka. Retreat rates of the Laurentide Ice Sheet in New England have been estimated to be 50-70 m/a (Gustavson and Boothroyd, 1987). Retreat rates in the marine environment would be considerably greater. Assuming the New England rate of ice retreat, the inner shelf would have been deglaciated by 14.5 ka and all of Nova Scotia by 13 ka. Southern Nova Scotia appears to have followed this timetable of retreat (Piper and Fehr, 1991; Stea and Mott, 1990). Dates on glaciomarine deltas, and minimum dates on lake sediments and buried organic horizons in northern Nova Scotia, however, imply residual ice until at least 12 ka and in some highland areas, probably much later (Stea and Mott, 1989; 1990). These data suggest that the retreat of the ice in northern Nova Scotia after 15 ka was slow or interrupted by several stillstands and possible readvances. On land, Stea and Finck (1984) documented a late-glacial readvance in northern Nova Scotia, as yet undated, stemming from the Antigonish Highlands (Phase 4; Fig. 5). This ice sheet flowed southwestward and southward into the adjacent lowlands. Along the Eastern Shore the margin of this ice mass could not be defined and was presumed to be offshore. The Sheet Harbour transect indicates two stillstands and possible readvances after the deposition of the Eastern Shore Moraine. Ice stemming from the Scotian Ice Divide (Phase 3; Fig. 5) may have formed or overrode the Eastern Shore Moraine. The remnant Antigonish Highland ice cap complex (Phase 4; Fig. 5) may have formed the inner shelf morainal zones. Minimum dates on lake sediments suggest that the advance occurred sometime between 12 and 13 ka. A possible correlative event is the Robinsons Head advance of Newfoundland (Grant, 1989) dated ca. 12.6 ka. Stea and Mott (1989) proposed that ice may have covered part of the Eastern Shore during the Younger Dryas. A lake core in the region, however, penetrated the sedimentological "oscillation" that marks the Younger Dryas (11 000-10 000 BP) elsewhere in Nova Scotia (R. J. Mott pers. comm, 1991). This implies that much of the Eastern Shore region was ice free during the Younger Dryas. There is strong evidence, however, of Younger Dryas ice build up in areas north of the Cobequid Highlands and in offshore areas adjacent to Chedabucto Bay (Stea and Mott, 1990; King and Fader, 1988).

CONCLUSION

We interpret the seismic sequences and zones in the Sheet Harbour transect to represent grounded ice and ice marginal deposition from a retreating ice mass. This retreat was interrupted by stillstands and slight readvances. On land, ice centres were changing from an ice divide which straddled the axis of the Nova Scotia peninsula to remnant, highland ice caps. The events on the inner shelf can be summarized in 6 stages (Fig. 6).

Stage 1: Ice ramp front at the Eastern Shore Moraine, relative sea level high.

Stage 2: Landward retreat of the ice front. Debris is fed to the glaciomarine environment by meltwater emanating from the glacier sole. Formation of units 1, 2 (Emerald Silt) landward of the Eastern Shore Moraine.

Stage 3: Stillstand or readvance of the ice front to form the minor moraines (Zone 5). Erosion of the Emerald Silt in the ice proximal area (Zone 3). Deposition of unit 3 in the deeper basin.

Stage 4: Retreat of the ice front to Zone 5. Formation of the mounds-ridges in shallowing water (depths <40m?) and synchronous deposition of unit 3 in the adjacent basinal areas (Zone 4).

Stage 5: Further retreat of the ice front with relatively high RSL. Glaciomarine? deposition in the valley zone.

Stage 6: Drop in sea level to 70 m. Subareal exposure of glaciomarine sediments; formation of beach zone.

ACKNOWLEDGMENTS

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Geological Survey of Canada Project 870052

Inner shelf Quaternary sediments off northeast Newfoundland[1]

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Abstract

The characteristics of four northeast Newfoundland inner-shelf seabed environments – Baie Verte, La Scie, Halls Bay and Hamilton Sound - are discussed in relation to the potential of the region to host marine placer gold deposits. The shallow, relatively high-energy, coastal environments at the mouth of Baie Verte, seaward of La Scie and within Hamilton Sound appear to be the most favourable environments for concentrating particulate gold. These areas contain coarse clastics, are exposed to wave action, and have yielded samples showing anomalous gold concentrations. In contrast, the deeper, quiescent setting of Halls Bay appears to be an unfavourable site for placer development.

Résumé

Les caractéristiques de quatre milieux marins de plate-forme interne au nord-est de Terre-Neuve - Baie Verte, La Scie, la baie Halls et le détroit de Hamilton - sont traitées en fonction de la possibilité de découvrir des gisements aurifères marins. Les milieux littoraux peu profonds relativement agités à l'embouchure de la baie Verte, au large de La Scie et dans le détroit de Hamilton semblent être des milieux très favorables à la concentration de particules d'or. Ces zones contiennent des roches clastiques grossières, sont exposées à l'action des vagues et ont donné des échantillons dans lesquels les concentrations en or étaient anormales. Par contre, le milieu calme, plus profond de la baie Halls semble être un endroit non favorable à la mise en place de gisements alluvionnaires.

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INTRODUCTION

Under the aegis of the Canada-Newfoundland Cooperation Agreement on Mineral Development (MDA) a program was initiated in 1990 to determine the placer gold potential of inner-shelf environments along the northeast Newfoundland coast. Numerous onshore mesothermal quartz vein deposits on the Baie Verte Peninsula, the Springdale Peninsula, and along western White Bay (Solomon and Emory-Moore, 1989; Dubé, 1990) have provided incentive for offshore exploration. The prerequisites for identifying inner-shelf placer gold deposits are a knowledge of the character and distribution of Quaternary sediments on the inner shelf, an understanding of Late Quaternary relative sea-level changes, and a knowledge of how contemporary processes such as waves, currents and iceberg scour rework the seabed. Recent surveys carried out thus far under the agreement were designed to improve our understanding in these areas.

The major objective of this paper is to present new information from four areas of the northeast Newfoundland inner shelf with contrasting Quaternary sediment types and seabed conditions. These areas are Baie Verte, the La Scie region, Halls Bay and Hamilton Sound (Figure 1). The secondary objective is to suggest how this new information affects the search for offshore marine gold placers in the region.

STUDY AREA

Geological setting

Coastal bedrock exposures within the study area comprise alternating northeast trending belts of Cambrian, Ordovician and Silurian marine siliciclastic and volcanic rocks, except the western margin of the Baie Verte Peninsula which is comprised of Proterozoic basement rock. To the east these exposures are within the Gander Zone; to the west they are within the Dunnage Zone and include the Ordovician Davidsville and the Silurian Botwood Groups (Evans, 1991). These two groups host gold occurrences and have been the recent focus of onshore mineral exploration activity.

Seaward of Baie Verte, upper Proterozoic rocks are in contact with lower Carboniferous rocks. To the east, in Notre Dame Bay, these lower Carboniferous rocks are in fault contact with Cambrian-Ordovician and Ordovician-Silurian rocks (Fader et al., 1989; Haworth et al., 1976). Airgun data recently collected on CSS Dawson cruise 91-026, along the eastern margin of the study area, northeast of Fogo Island, show a wedge of stratified sediment overlying basement rock. This appears to confirm the relationship between Devonian granite and overlying Cretaceous/Tertiary siliciclastics given in Fader et al. (1989).

The character and distribution of Late Quaternary sediments on the northeastern shelf were studied by Dale and Haworth (1979). They mapped five seismic facies as follows: an acoustically transparent layer; an acoustically-stratified unit; a semi-stratified unit; a non-stratified unit interpreted as till, and bedrock. Prior to the recent MDA work (Shaw and Wile, 1990; Shaw et al., 1990a; Shaw, 1991) virtually no information existed on the distribution of Quaternary sediments on the inner shelf of this region, except for some limited work on surficial sediments in Halls Bay (Slatt, 1975; Slatt and Gardiner, 1976).

Physiography and oceanographic setting

The study area comprises the inner shelf between the Baie Verte Peninsula and Hamilton Sound (Figure 1). The coastline is generally characterized by steep, rocky headlands, rising to several hundred metres within a kilometre of the sea, contiguous to southwest-northeast trending fiord-type embayments. The relief diminishes toward the east and the character of the coastline changes. The so called "Straight Coast" between Hamilton Sound and Cape Freels consists of numerous sandy beaches and coastal dunes in small, rock-bounded compartments (Shaw and Forbes, 1990).

The seafloor of the inner shelf forms a moderately sloping platform out to the 400 m bathymetric contour. Numerous narrow embayments reaching depths of 200 to 500 m dissect this platform. The four parts of the inner shelf described in this report represent the range of physiography in the region: Baie Verte is a small embayment on the north coast of the Baie Verte Peninsula; the La Scie area is an exposed, shallow, inner-shelf environment off the northeast coast of the peninsula; Halls Bay is a long, deep fiord basin; and Hamilton Sound is a shallow, relatively open embayment in the eastern part of the study area.

Glaciation and relative sea-level changes

Quaternary glaciation has left a deep imprint on the inner shelf, both through erosion and the deposition of glaciomarine sediments. The Late Wisconsinan stadial maximum in Newfoundland occurred between 14 000 and 13 000 yrs BP (Grant, 1989; Dyke and Prest, 1987). According to Brookes (1970), Grant (1972; 1974) and Rogerson (1982) Newfoundland was not overridden by the Laurentide ice sheet during the Late Wisconsinan glacial period but supported an independent ice cap located in central Newfoundland.

The early recessional stages of the Late Wisconsinan glaciation and final ice retreat occurred between 14 000 and 10 000 yrs BP. These resulted in extensive postglacial shore level change along the northeast coast dominated by the effects of crustal recovery. Grant's (1989) isobase map of delevelled paleoshores depicts net emergence in the study area. A maximum marine limit of approximately +90 m on the Baie Verte Peninsula is part of an eastward decreasing continuum with a lower marine limit of +75 m in Halls Bay (Tucker, 1974) and postglacial emergence of only +43 m (Grant, 1980) in Hamilton Sound.

METHODOLOGY

During cruise 90-013 (Shaw and Wile, 1990), data were collected from CSS Hudson in White Bay, around the Baie Verte Peninsula, and in Green Bay, Halls Bay and Notre Dame Bay (Figure 1). Survey lines overlapped those previously reported by Dale and Haworth (1979) to facilitate regional coverage. Huntec DTS high resolution seismic records, air gun records, 3.5 and 12 kHz data, 5 gravity cores, 3 IKU grab samples and 28 van Veen grab samples provided the data base from this cruise. Samples were analyzed by C-CORE under contract. Results are reported in an unpublished report (C-CORE, 1990) and in Shaw et al. (1990b).

On cruise 90-035, data gathered from CSS Navicula within Baie Verte, the approaches to La Scie Harbour, Halls Bay and Hamilton Sound (Figure 1) provided more extensive and detailed coverage of the inner shelf. Data were collected using a Seistec high-resolution seismic reflection system, a Datasonics Bubble Pulser seismic reflection system, a Klein 595 sidescan sonar towfish, the ship's 30 kHz echosounder, a van Veen grab sampler, an Alpine gravity corer and a bottom camera. The sampling program resulted in a total of 98 van Veen grab samples, 98 accompanying seabed photographs and 9 gravity cores (Shaw et al., 1990a). Samples from cruise 90-035 were analyzed by C-CORE under contract. Results and interpretations are contained in Open File Report 2417 (Emory-Moore, 1991). Data from CSS Dawson cruise 91-026 (Shaw, Russell, Sherin and Atkinson; unpublished data) are not yet fully analyzed and are not discussed in this report.

BAIE VERTE

Baie Verte is a 15 km long, 2 km wide, fiord-like embayment on the north coast of the Baie Verte Peninsula. Because of its proximity to numerous gold occurrences onshore, it has been a high-priority target in the current MDA program. Tracks and sample locations for cruises 90-013 and 90-035 are shown in Figure 2A. The inner bay, landward of Duck Island, is relatively shallow (Figure 2B) with a moderate-relief coast. Small sand and gravel beaches are common, particularly on the south side of the bay, whereas the north coast is characterized by steep, rocky slopes. The Baie Verte River forms a delta at the head of the bay.

In the inner bay Quaternary sediment thickness averages 5 m with a maximum thickness of 25 m (Figure 2C). Several seismo-stratigraphic units are defined. Acoustic basement is bedrock, recognized by its strong surface reflection. Bedrock exposures are few in the inner bay (Figure 2D). The stratigraphically lowest Quaternary unit is acoustically incoherent, and while generally thin, forms three ridges up to 25 m thick, transverse to the bay. Where exposed at the seabed the unit has a veneer of boulders, gravelly mud or muddy gravel. This unit is interpreted as glacial diamict. Overlying the glacial diamict is an acoustically stratified, highly conformable unit, 2 to 5 m thick. Cores 90-035-133, -134, and

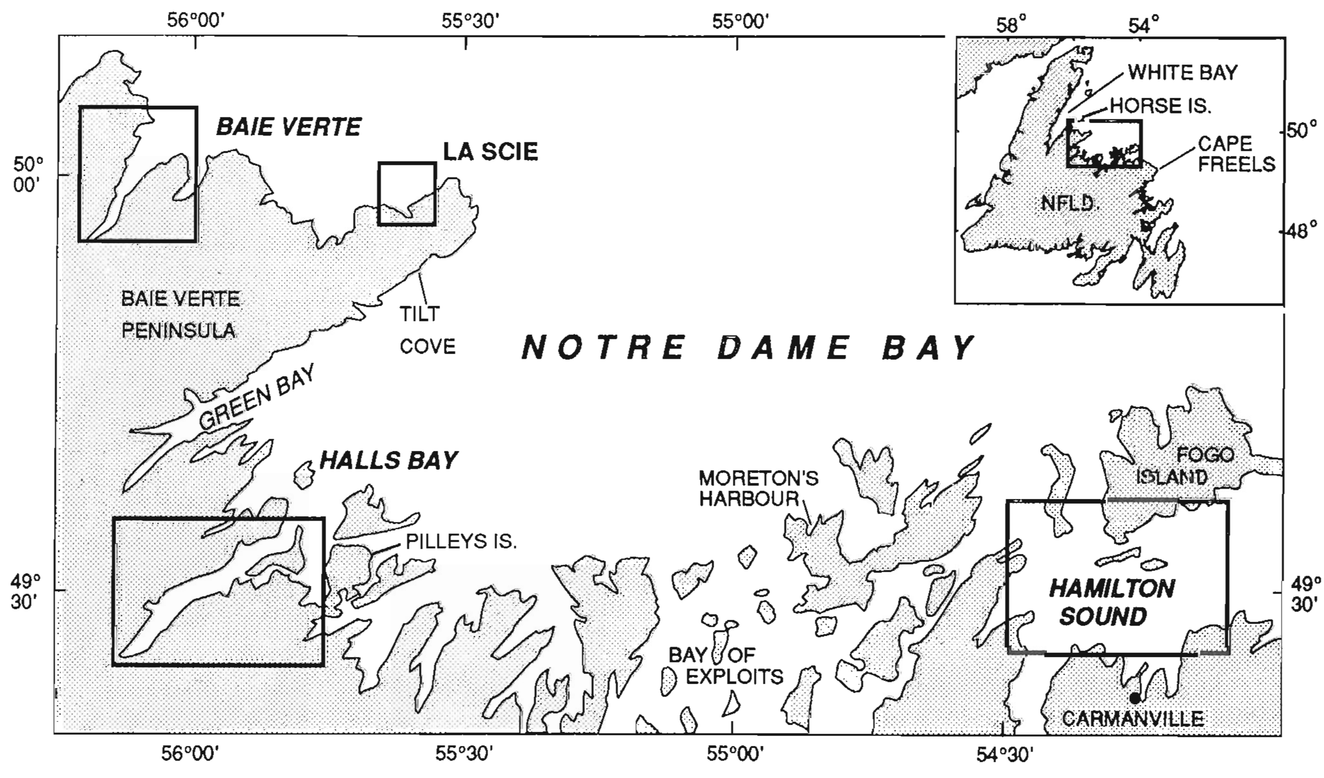


Figure 1. Location of study areas on the northeast Newfoundland inner shelf.

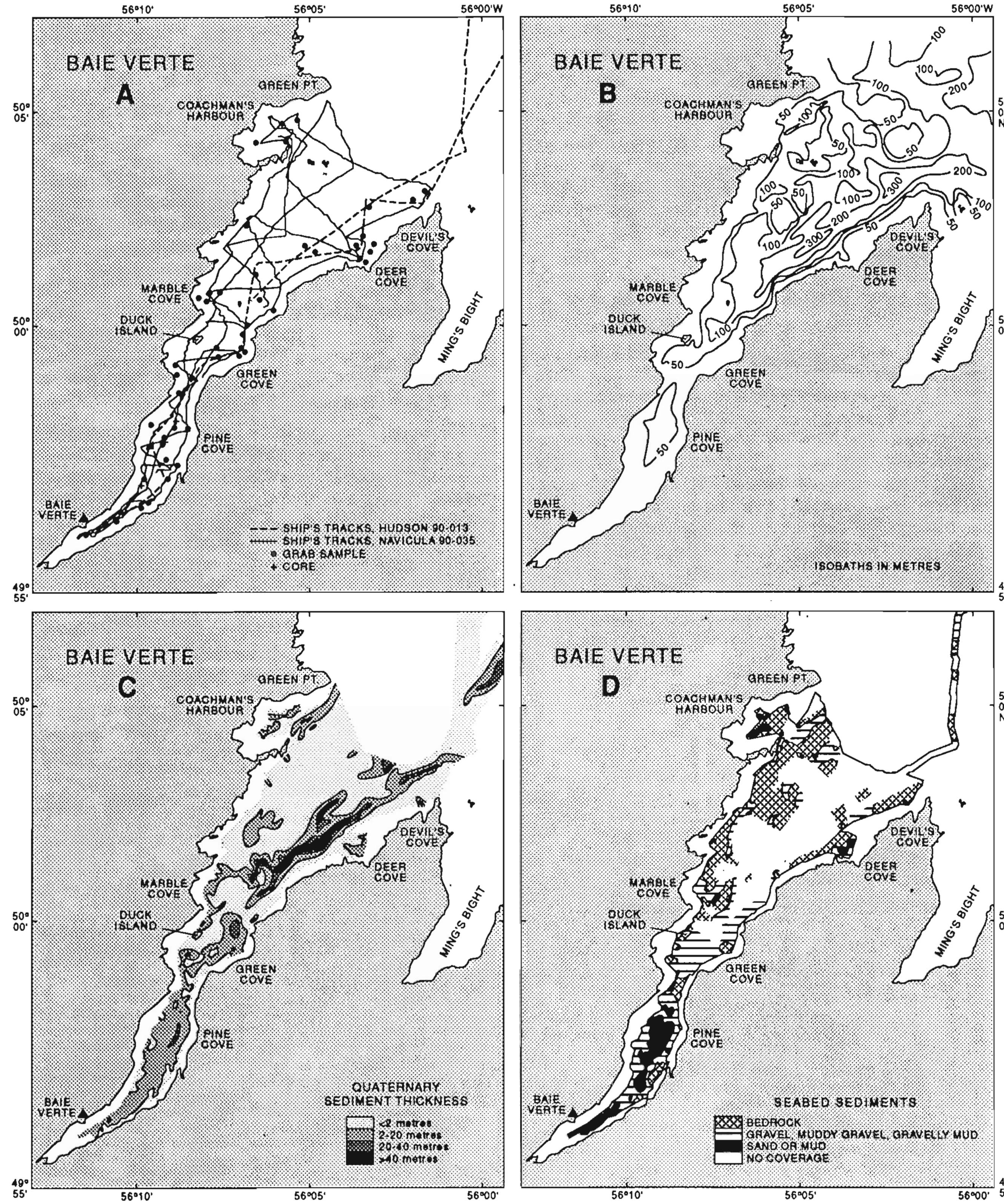


Figure 2. Baie Verte: **A**) tracks and sample locations; **B**) bathymetry; **C**) Quaternary sediment thickness; **D**) seabed sediment types interpreted from sidescan sonograms.

-135 show this unit to be grey clay with grit and pebbles (Shaw; unpublished data). The uppermost unit is acoustically transparent and is interpreted as ponded postglacial mud with minor ice-rafted gravel. It reaches a maximum thickness of 24 m in the central corridor of the bay. On sidescan records the seabed expression is a featureless or mottled surface, which, in the vicinity of the harbour, is heavily incised with curvilinear and plumose drag marks. In water depths between 17 and 46 m this unit displays iceberg furrows and pits. Profiles of acoustically incoherent postglacial sediments along the steep sides of the inner bay coincide with a botryoidal pattern on sidescan sonograms. This is interpreted as creep or slump features. A sample comprising two in situ specimens of the gastropod *Tachyrhynchus erosus* found at the base of this unit in core 90-035-134 were dated (AMS) at 5510 ± 160 yrs BP (TO-2402; adjusted for 410 yr reservoir effect).

The outer bay (Figure 2B) is characterized by a more irregular bathymetry with numerous shoals and islands and a progressive seaward shoaling. The average depth in the central part of the outer bay is 100 m, apart from a deep trough which reaches a maximum depth of 325 m. Bedrock is extensive in the outer bay and Quaternary sediments are thin (Figure 2C), with several exceptions. The trough contains up to 70 m of acoustically stratified sediment, most of which is believed to be glaciomarine mud. The uppermost sediments in the trough have weaker acoustic stratification and are probably either ice-distal glaciomarine or of postglacial age. Accordingly, the thickness of Holocene sediments in the trough is uncertain. Pockets of thicker (usually up to 10 m) surficial sediments are located in relatively shallow coastal embayments of the outer bay – Marble Cove, Coachman's Harbour, Devil's Cove and Deer Cove – separated from adjacent basinal muds by steep, bedrock sidewalls. These deposits are largely postglacial in age, and comprise sand and gravel. There is evidence of iceberg impacts (pits and furrows) in these exposed outer bays, and of considerable sediment mobility (gravel ripples at one location).

Outside the bay the exposed inner shelf north of the Baie Verte Peninsula is characterized by extensive exposures of bedrock and a thin, discontinuous cover of acoustically-stratified glaciomarine and postglacial mud reaching a maximum thickness of 10 m south of Horse Islands (Shaw and Wile, 1990). In water depths shallower than 180 m the Quaternary sediment is acoustically incoherent and is comprised of mixtures of gravel and mud. This cover coincides with a zone of modern iceberg scour and has been interpreted by Shaw and Wile (1990) as an iceberg turbate.

The variation in the distribution and thickness of Quaternary sediments within Baie Verte is primarily due to bathymetry and former glacial ice positions. Most sedimentation appears to have been limited to the narrow, deep trough. Limited sedimentation in the outer bay and the area seaward probably results from the combined influence of a shallow depositional environment and energetic oceanographic conditions. In terms of processes capable of winnowing sediments to produce gold placers, parts of the outer bay are favourable settings. Indeed, anomalous levels of gold, including visible particles, were found in localized, shallow pockets of coarse postglacial sand proximal to local till sources, in Deer Cove and elsewhere (Emory-Moore, 1991).

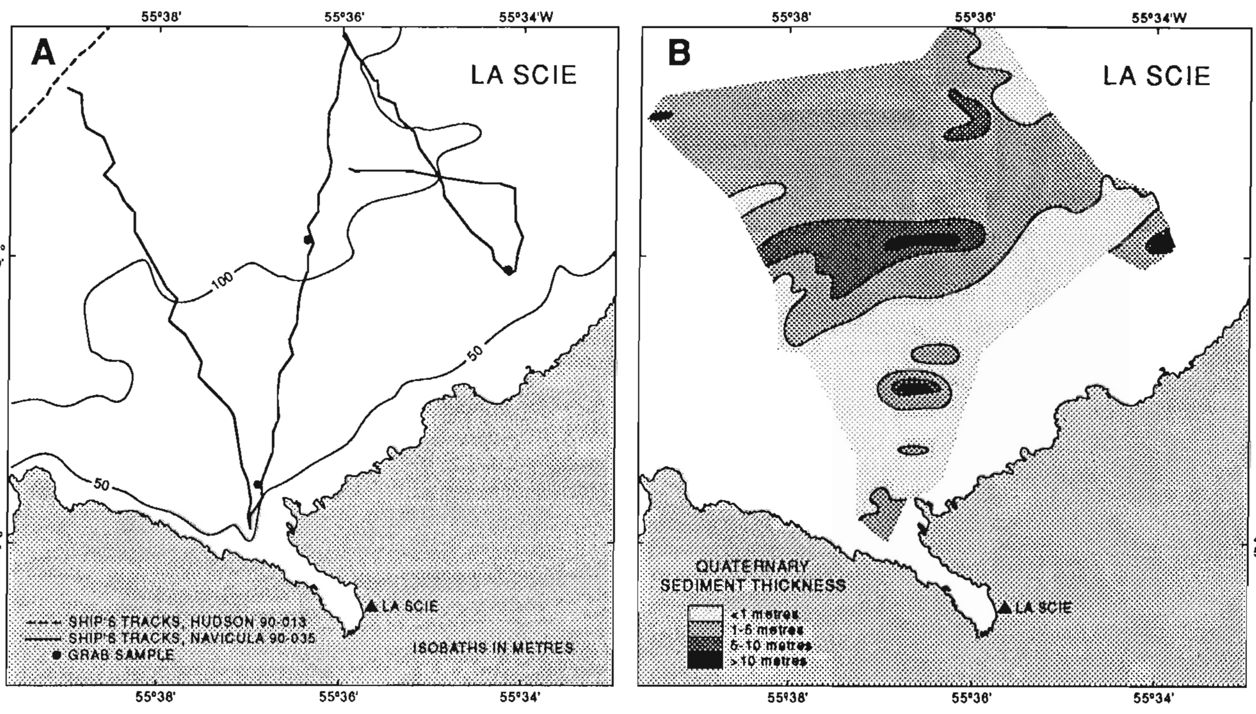


Figure 3. La Scie: A) tracks, sample locations and bathymetry; B) Quaternary sediment thickness.

These sediment bodies are small and discontinuous and their potential requires further study. Although glacial diamict is exposed on the seabed of the inner bay, modern processes in that area appear too weak to winnow and concentrate particulate gold.

LA SCIE

The settlement of La Scie is situated 30 km east of Baie Verte on the north coast of the Baie Verte Peninsula (Figure 1). In contrast to Baie Verte, seabed relief adjacent to La Scie is subdued, forming a moderate gradient out to 120 m, 1 km northwest of the coastline, and a more gentle gradient out to 65 m, 1 km to the northeast (Figure 3A). The distribution of inner-shelf Quaternary sediments north of La Scie, mapped by Shaw (1991), is presented in Figure 3B. Quaternary sediment cover out to the 90 m bathymetric contour is thin and discontinuous. The thickest deposits, reaching a maximum of 15 m, are confined to isolated depressions within bedrock. The sediment is acoustically unstratified with an irregular upper surface commonly overlain by 1 to 3 m of sand and gravel. Large patches of sand and gravel ripples occur in water depths between 45 and 55 m. Seaward, small patches of gravel ripples, developed in poorly sorted mixtures of medium sand and fine gravel, were observed in water depths of 67 to 82 m.

Quaternary sediment cover between water depths of 90 and 130 m forms a continuous veneer, 1 to 5 m thick, with a maximum thickness of 10 m in isolated bedrock depressions. This sediment is acoustically incoherent with an irregular surface that coincides with long, linear, seabed features on sidescan records. It is interpreted by Shaw (1991) as an iceberg turbate. A patchy distribution of fine to medium sand with scattered pebbles overlies the furrowed unit, reaching a thickness of 1 m in shallow depressions.

Although the data show that the outer zone off La Scie is impacted by icebergs, our observations suggest that the inner zone is also impacted. However, the absence of iceberg furrows in the inner zone (shallower than 90 m) probably results from the mobilization of seabed sediments by wave action. This zone of mobile, nearshore sediments may extend along the coast towards Ming's Bight to include the coarse clastic sediment bodies observed at the mouth of Baie Verte. Gold grains within a sample from 80 m water depth, in the approaches to La Scie, had a fresh appearance (uncorroded, unabraded). This may be interpreted as evidence that the travel distance for these grains is limited, and that a mineralized zone may be present offshore in this area.

HALLS BAY

Halls Bay is a 25 km long, 3 km wide fiord on the south coast of Notre Dame Bay (Figure 1). Tracks and sample locations are shown in Figure 4A. The head of the bay is relatively shallow (Figure 4B), dipping gently out to the 100 m isobath; trough sidewalls dip steeply toward the centre of the bay, which steepens from 100 m near its head to 450 m at its mouth.

The bay contains a ponded sequence of Quaternary sediments up to 170 m thick (Figure 4C), confined between steep (15°) sidewalls. These sediments are acoustically well-stratified, characterized by irregularly-spaced, variable intensity, continuous, coherent reflections. Intervals of acoustic transparency, with irregular upper bounding surfaces, are interpreted as evidence of sediment gravity flows which have destroyed acoustic stratification. The fiord sidewalls have a thin (~1 m) sediment cover, with bedrock outcropping in places. Southwest of Springdale the sediment package thins and there is evidence of deformation and slumping, often of sufficient intensity as to destroy acoustic stratification. A 7 m core (90-013-063) in fiord sediments several kilometres southwest of Springdale consisted mainly of stiff, gritty muds with a high content of gravel (dropstones).

Sidescan sonar surveys of the fiord sidewalls between Springdale and the head of the bay show a botryoidal pattern on the seabed. This is interpreted as small-scale terracettes developed in thin (<2 m) sediments by slow creep down the sidewalls. These features occur in depths of 60 to 200 m and on slopes up to 5°.

The head of Halls Bay is the site of well-developed raised glaciomarine deltas (Tucker, 1974; Liverman et al., 1991; Scott and Liverman, 1991). A date of 12 000 ± 220 yrs BP for shells in prodelta mud was cited by Grant (1989) as providing a minimum age for coastal end moraines in the area. Lowden and Blake (1975) cite a date of 11 000 ± 190 for shells in silt/clay delta bottomset beds, and shells in stony mud at Pilleys Island are dated at 11 900 ± 200 (Blake, 1983).

These published dates relating to the period of raised delta formation correlate well with an AMS date of 11 600 ± 80 years BP (TO-2395; adjusted for a 410 yr reservoir effect) on a *Macoma calcarea* specimen from the top of glaciomarine sediments in core 90-035-183, recovered east of Sunday Cove Island (Figure 4A). Nearby, several AMS dates have recently been obtained on shells contained in acoustically transparent deposits overlying the glaciomarine sediments. These are 8250 ± 80 yrs BP (TO-2398) on a specimen of *Nuculana pernula*, and 8870 ± 80 yrs BP (TO-2397) on a specimen of *Macoma calcarea* (both samples from core 90-035-180), and 6050 ± 70 yrs BP (TO-2396) on *Macoma calcarea* fragments from core 90-035-181. These new dates suggest that the deposition of the thick deposits within Halls Bay was coeval with the formation of raised glaciomarine deltas in the region.

The muddy sediment samples of Halls Bay were devoid of particulate gold. This may be a function of low sample density and a dominantly massive sulphide terrain with minor occurrences of gold in bedrock (Emory-Moore, 1991), but is equally likely to reflect the deep, low-energy depositional environment of the bay. This setting makes the bay a poor prospect for placer development.

HAMILTON SOUND

Hamilton Sound is situated 75 km east of Halls Bay, on the eastern edge of the study area (Figure 1). Extensive areas within the sound are shallower than 20 m (Figure 5A), although the channel west of Dog Bay Islands (Figure 5A)

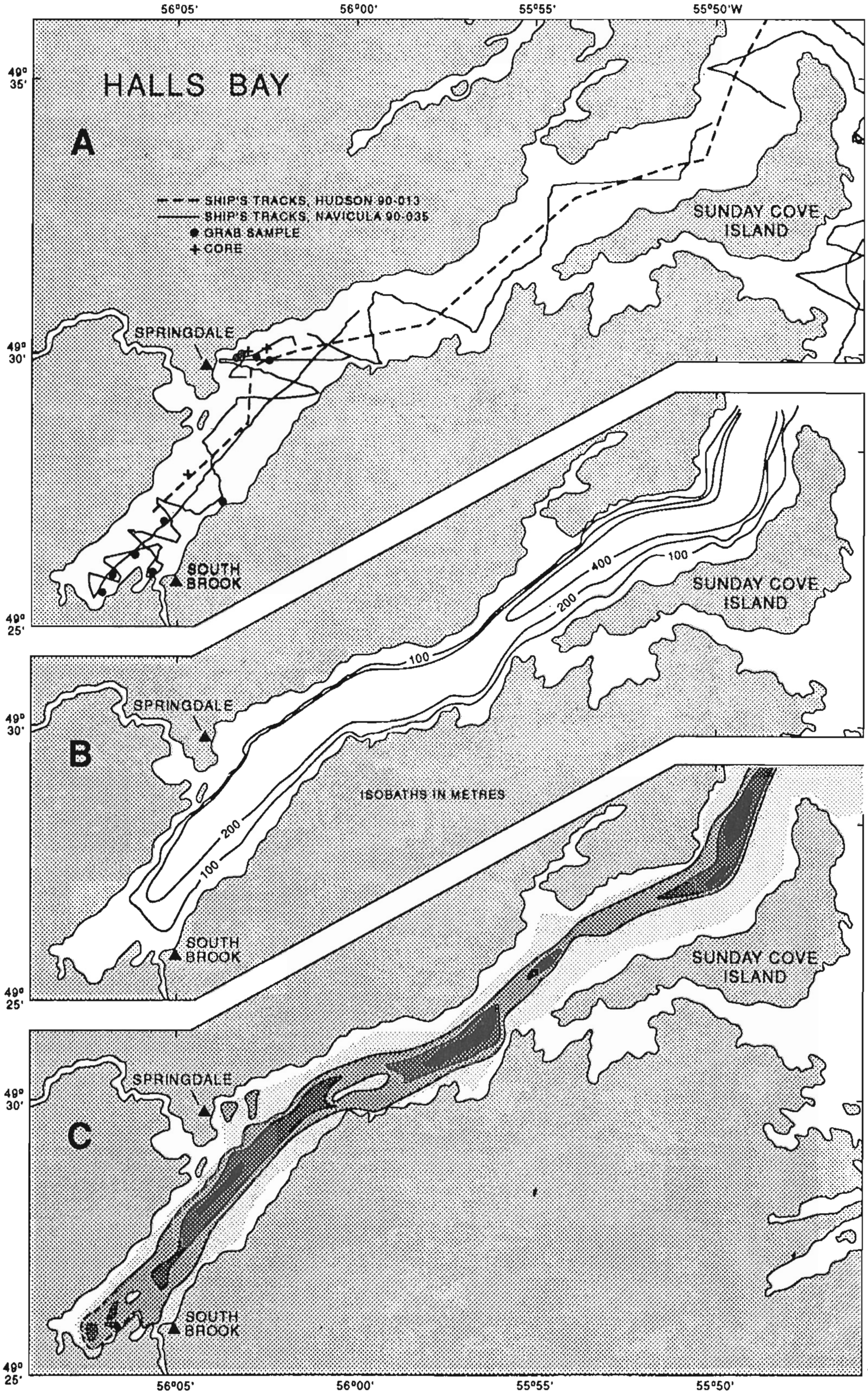
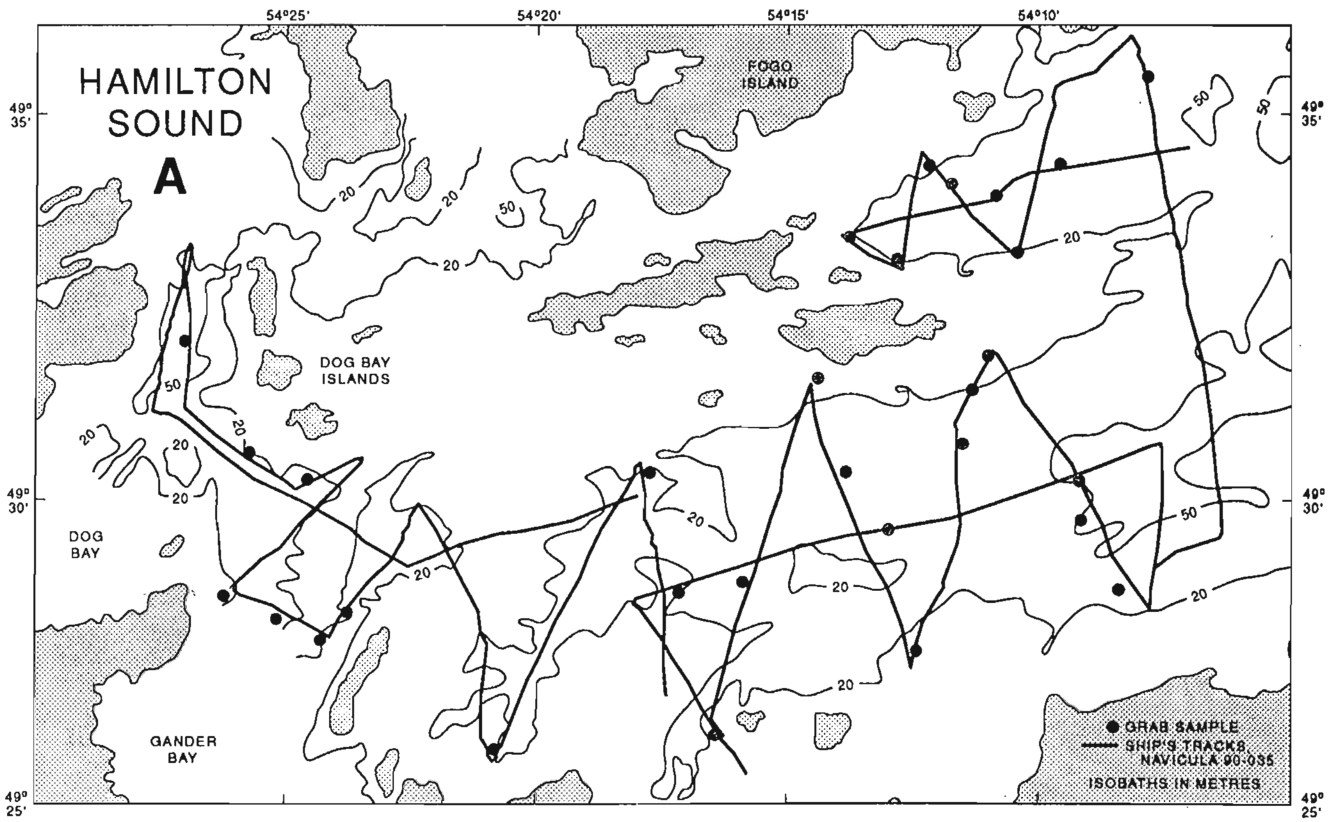


Figure 4. Halls Bay: **A**) tracks and sample locations; **B**) bathymetry; **C**) Quaternary sediment thickness.



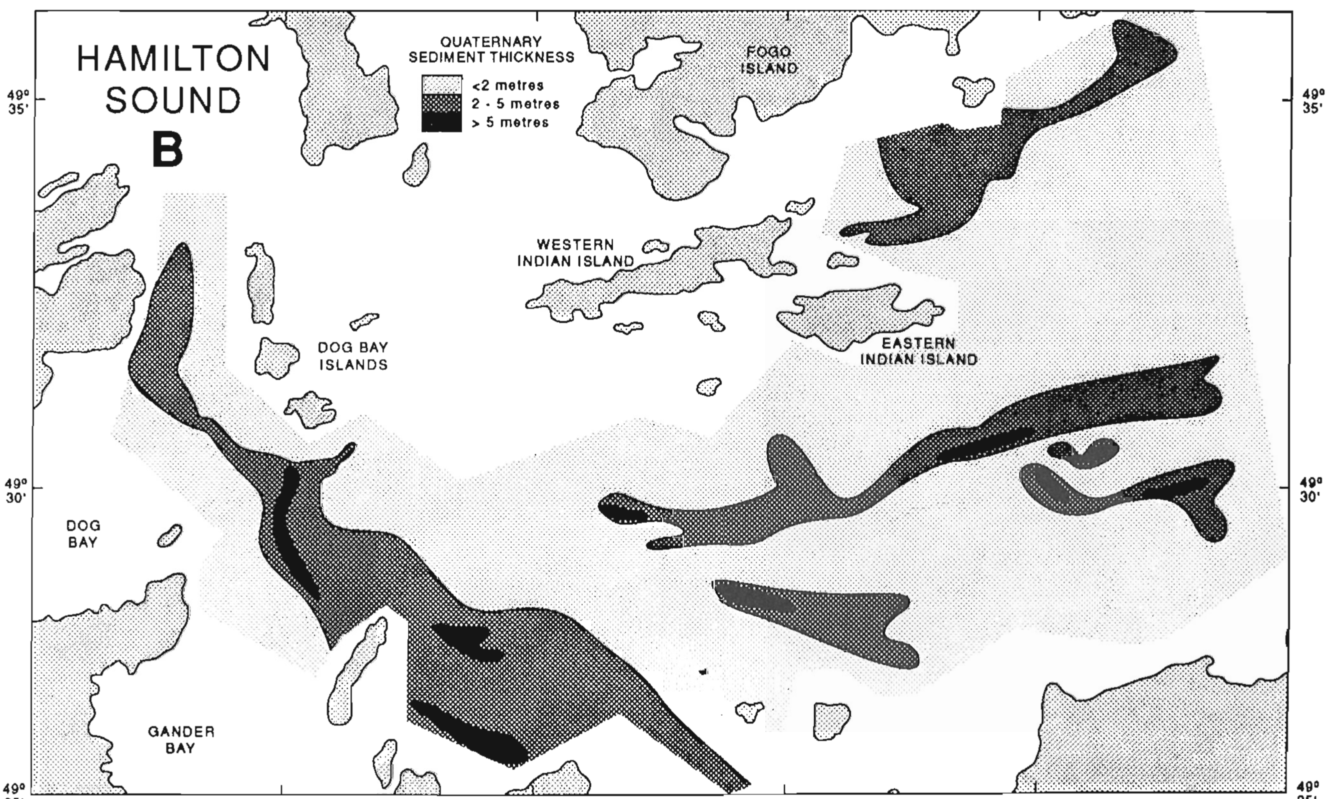


Figure 5. Hamilton Sound: **A**) tracks, sample locations and bathymetry; **B**) Quaternary sediment thickness.

has a maximum depth of 70 m and several deep channels exist southeast of Fogo Island. Most of the coastline is low, irregular, and rocky with scattered gravel and sand beaches, often strewn with large boulders.

Quaternary sediment thickness in the sound (Shaw, Russell and Jenner; unpublished data) is presented in Figure 5B. The seafloor in the western region of Hamilton Sound is primarily comprised of a coarse gravel veneer above glacial diamict, with rare occurrences of gravel ripples. Several rounded, elongate mounds of glacial diamict with positive relief of up to 14 m occur at the mouth of Gander Bay and these features are tentatively interpreted as drumlins. The central region of Hamilton Sound is characterized by a thin sandy-gravel cover over glacial diamict, and scattered bedrock exposures. In the eastern part of the sound a zone of bedrock with minor gravel extends east from Eastern Indian Island. To the south of this zone is an extensive area of sand that infills depressions between mounds and ridges of glacial diamict. The surface of the diamict is capped by a thin gravel veneer. To the north of the bedrock zone, close to the south coast of Fogo Island, is a second area of predominantly sandy seabed. Iceberg furrows and pits occur in this area in water depths between 22 and 44 m; sandy bedforms with underlying gravel exposed in the troughs have an average wavelength of 12 m and occur in water depths of 27 to 37 m.

The depositional environment of Hamilton Sound appears to be favourable for gold accumulation. However, anomalous gold was identified at only one site, southeast of Eastern Indian Island, within a fine silty sand. The absence of gold may be attributable in part to the sampling technique, since gold, if present, may be concentrated within the upper 1 m beneath the seabed (Emory-Moore, 1991). It may also be due to a paucity of local gold sources onshore.

DISCUSSION

The present study addresses the sedimentary character of the inner shelf off northeast Newfoundland, and its influence on the potential for marine placer gold in the area (Solomon and Emory-Moore, 1989). Preliminary results show that this inner shelf has a large diversity of seabed environments.

Baie Verte is a glacially over-deepened embayment that has acted as a sediment trap, accumulating greater thicknesses of glacial marine and marine muds than the nearby shelf. The contrasts between the inner and outer bays are important. The sheltered inner bay accumulates muddy sediments with minimal reworking of glacial deposits. In parts of the outer bay, coarse clastic deposits in shallow coastal compartments (Deer Cove, Devil's Cove, Marble Cove and Coachman's Harbour) are mobilized both by wave action and by grounding icebergs. They provide a favourable environment for the concentration of particulate gold.

Halls Bay is a larger embayment, similar to Green Bay and Bay of Exploits in that it hosts great thicknesses of glaciomarine deposits. The prospects for placer concentration in these deep, quiescent settings are remote. Indeed, the only anomalous gold level in this region was in sandy sediment northeast of Sunday Cove Island (Emory-Moore, 1991).

The pockets of coarse sediments at the mouth of Baie Verte appear to be outliers of a wave-mobilized facies that is developed in the region wherever the inner shelf is sufficiently shallow and wave-exposed. Six samples from these pockets contained gold in the sand fraction (Emory-Moore, 1991). This zone probably extends eastward along the coast of the Baie Verte Peninsula towards La Scie, where one of two samples analyzed also contained gold in the sand fraction.

Hamilton Sound is part of the largest area of shallow, wave-exposed inner shelf in the region. Unfortunately, most of the sound lies east of a favourable zone of mineralization onshore. The areas closest to mineralized zones (Gander Bay and Dog Bay) are least exposed to wave action and, hence, to winnowing of the glacial diamicts exposed on the seabed. The largest area of gravel ripples hitherto observed off eastern Canada is located nearby, between Hamilton Sound and Cape Freels (Figure 1). This area may have a lower potential to host gold placers in terms of its position relative to onshore gold occurrences. Nevertheless, samples collected during 1991 field season are presently being analyzed.

In summary, ongoing seabed mapping on the northeast Newfoundland inner shelf is revealing a great diversity of Quaternary deposits and seabed characteristics. This work is also delineating the extent of those coarse grained deposits which may have an enhanced potential to host placer gold. While conclusions regarding the potential for economic placer gold on the inner shelf would be premature, it should be noted that out of 88 samples collected on Navicula cruise 90-035, 14 contained anomalous levels of gold (>50 ppb) in the mud fraction and 8 contained particulate gold within the sand fraction (Emory-Moore, 1991).

Numerous samples, including large volume IKU grabs, were collected in the 1991 field season, many from areas previously unsurveyed. These areas are the west side of White Bay, Ming's Bight (near Baie Verte), off Tilt Cove (on the east side of the Baie Verte Peninsula), the Moreton's Harbour area, and the region between Hamilton Sound and Cape Freels. The results of gold assays and geochemical analyses of these new samples, and of analyses of the geophysical data which were also collected, should further advance our knowledge of the inner shelf and facilitate a proper evaluation of its gold potential.

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Structural geology of a crustal scale fault zone: the Cape Ray Fault coastal section, southwestern Newfoundland[1]

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Abstract

The coastal section of the Cape Ray Fault (CRF) exposes a crustal scale fault zone. Kinematic indicators in non-coaxial fabrics indicate that high grade amphibolitic deeper crustal-level rocks of the Port Aux Basques gneiss were thrust upon greenschist-grade supracrustal rocks of the volcano-sedimentary Windsor Point Group (WPG). The kinematics of the deformation is complex, with evidence of both reverse-sinistral overthrusting of the amphibolitic rocks over greenschist rocks, and reverse-dextral thrusting recorded in the WPG rocks as suggested by the variation in the stretching lineation trend and timing relationship. The structures recorded within WPG present complex internal geometry, typical of mylonitized rocks and could be attributed to a single progressive deformational event.

The discovery of a wide zone of CO_2 metasomatism within the mylonite zone may have important gold metallogenic implications since it could represent a fossil hydrothermal discharge zone across the CRF.

Résumé

La zone côtière de la faille de Cape Ray (FCR) représente une section à travers une faille crustale. Les indicateurs cinématiques développés dans les fabriques non-coaxiales indiquent que les amphibolites de niveau crustal profond, appartenant au gneiss de Port-aux-Basques, ont été chevauchés sur des roches supracrustales du faciès des schistes verts. La cinématique de la déformation est complexe comme lsemble d'indiquer la variation d'orientation des linéations d'étirement et les relations chronologiques. On constate des indices de chevauchement inverse et senestre des roches amphibolitiques sur les roches du faciès des schistes verts et, subséquemment, un chevauchement à mouvement inverse et dextre relevé à l'intérieur des roches du Groupe de Windsor Point (GWP) . Les structures enregistrées dans le GWP présentent une géométrie interne complexe, typique des zones de mylonites et peuvent être attribuées à une seule phase de déformation progressive.

La découverte au sein de la FCR d'une large zone ayant subie un métasomatisme en CO_2 peut avoir des implications importantes du point de vue de la métallogénie aurifère, étant donné qu'elle peut marquer l'emplacement la localisation d'une zone fossile de décharge hydrothermale.

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INTRODUCTION

The Cape Ray fault (CRF) (Gillis, 1972) is one of the best exposed fault zone of the Canadian Appalachian to which significant gold mineralization is spatially and genetically related (Dubé, 1990; Dubé et al., 1991). It represents a key locality to study the cause-effect relationships between gold and major faults. In order to address this question, it is critical to define the timing of the gold mineralizing event within the structural history and geometrical evolution of the fault zone. Detailed structural mapping has already been conducted in the area surrounding the Cape Ray gold deposit (Dubé et al., 1991). A well-exposed section of the CRF outcrops along the coast of southwestern Newfoundland (Fig. 1). The area has been studied by Gillis (1972), Brown (1973), Wilton (1983), and Piasecki (1989). We studied it as a typical cross-section of the CRF in order to define the deformational history and the lithotectonic context of the CRF away from the mineralized areas. We also wanted to determine the similarities and differences between the barren coastal section of the fault and the mineralized sites found along it (see Dubé et al., 1991). The coastal section of the CRF has been briefly described in Dubé et al., (1991); this paper emphasizes the results of fieldwork done during the summers of 1990-1991.

Brown (1973, 1977) interpreted the CRF as a Taconian suture representing the trace of the Iapetus ocean. Wilton (1983) considered it as a major, sinistral, ductile shear zone of Acadian age. Chorlton (1983) recognized two faulting stages within the CRF: 1) a pre- to syn-Late Devonian sinistral wrench faulting followed by 2) a post-Late Devonian reverse-oblique dextral shearing. Piasecki (1989) emphasized that the CRF has the character of a major suture and he also recognized two movements which were essentially the same as in Chorlton's (1983) interpretation except that Piasecki identified the dominance of subvertical movements. Following work in the area around the Cape Ray gold deposit, Dubé et al., (1991) suggested that the post-Late Devonian movement of the fault was compatible with a transpressive dextral deformation regime. Two increments of ductile deformation were recorded, the first one characterized by reverse-oblique, dextral shearing, and the second by strike slip movement (Dubé et al., 1991).

GEOLOGICAL SETTING

The CRF 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, including 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, inset). The "Late Devonian" Windsor Point Group (WPG) (Wilton, 1983), located between these two terranes, consists of strongly deformed, volcano-sedimentary rocks intruded by the synvolcanic Windowglass Hill granite (WGH) (Wilton, 1983). The northwestern and southeastern terranes have been respectively intruded by the late- to post-tectonic Strawberry and Isle aux Morts Brook granites.

GEOLOGY OF THE CAPE RAY FAULT ZONE IN THE COASTAL SECTION

The southeastern limit of the study area (Fig. 1) is made up of strongly foliated dark grey to black biotite and hornblende-rich psammites probably belonging to the Grand Bay Division of the Port aux Basques Complex (PaBC) (van Staal et al., 1992) (Fig. 1, 2), whereas the northwestern limit is a tonalitic and a megacrystic granitic phase of the Cape Ray Igneous Complex (CRIC). The megacrystic granite is characterized by abundant microcline phenocrysts (Fig. 3A) and by epidotized mafic xenoliths.

Along the coastal section of the CRF, we have mapped three distinct sequences (Fig. 1, 2): 1) a bimodal volcanic sequence to the northwest, 2) a central sedimentary sequence, both located within the WPG, and 3) a mylonitic zone developed within the WPG and probably within retrograded "PaBC" to the southeast. The latter is referred to as the "transition zone".

1) Volcanic sequence

As recognized by Wilton (1984), volcanic rocks of this sequence are bimodal in composition. Felsic volcanics predominate to the Northwest, whereas mafic volcanics are abundant in the southern part. The felsic volcanics include ignimbrites, rhyolitic flows, and associated volcaniclastic rocks. The ignimbrites are dark grey to black, aphanitic siliceous rocks containing 1-2 millimetre-scale quartz and feldspar porphyroclasts, and millimetre- to centimetre-scale fragments of jasper, pink rhyolite, and ignimbrite (Fig. 3B). This unit is adjacent to a mylonitized, megacrystic granitic phase of the CRIC. At the contact, the ignimbrite is strongly sheared for up to 1 metre and contains strained fragments of the Cape Ray granite. Pink rhyolite that outcrops southeast of the ignimbritic unit is characterized by millimetre-scale feldspar phenocrysts set in an aphanitic matrix; flow banding and brecciation are common. Associated felsic volcaniclastics occur farther southeast, and comprise a bedded sequence of lapilli and fine grained tuffs. The coarser volcaniclastics are essentially monolithic and are composed of up to 60% angular to subangular clasts of pink rhyolite.

The southeastern part of the volcanic domain is dominated by mafic to intermediate volcanics occurring over a thickness of at least 300 metres. The mafic volcanic subdomain is well exposed at Jerret Point. It includes metre-thick pillowed and brecciated flows (Fig. 3C), gabbroic sills, and/or thick massive flows, and plagioclase-rich crystal tuffs or porphyritic flows. The abundance of leucoxene within the gabbroic sills suggests a tholeiitic affinity to at least part of the volcanic domain. Sedimentary rocks such as graywackes, siltstones, shales, and cherts are locally interbedded with the mafic volcanics.

2) Sedimentary Sequence

The clastic sedimentary sequence is exposed at Windsor Point, it is made up of folded and transposed sequences of conglomerate, interbedded with centimetre- to metre-scale

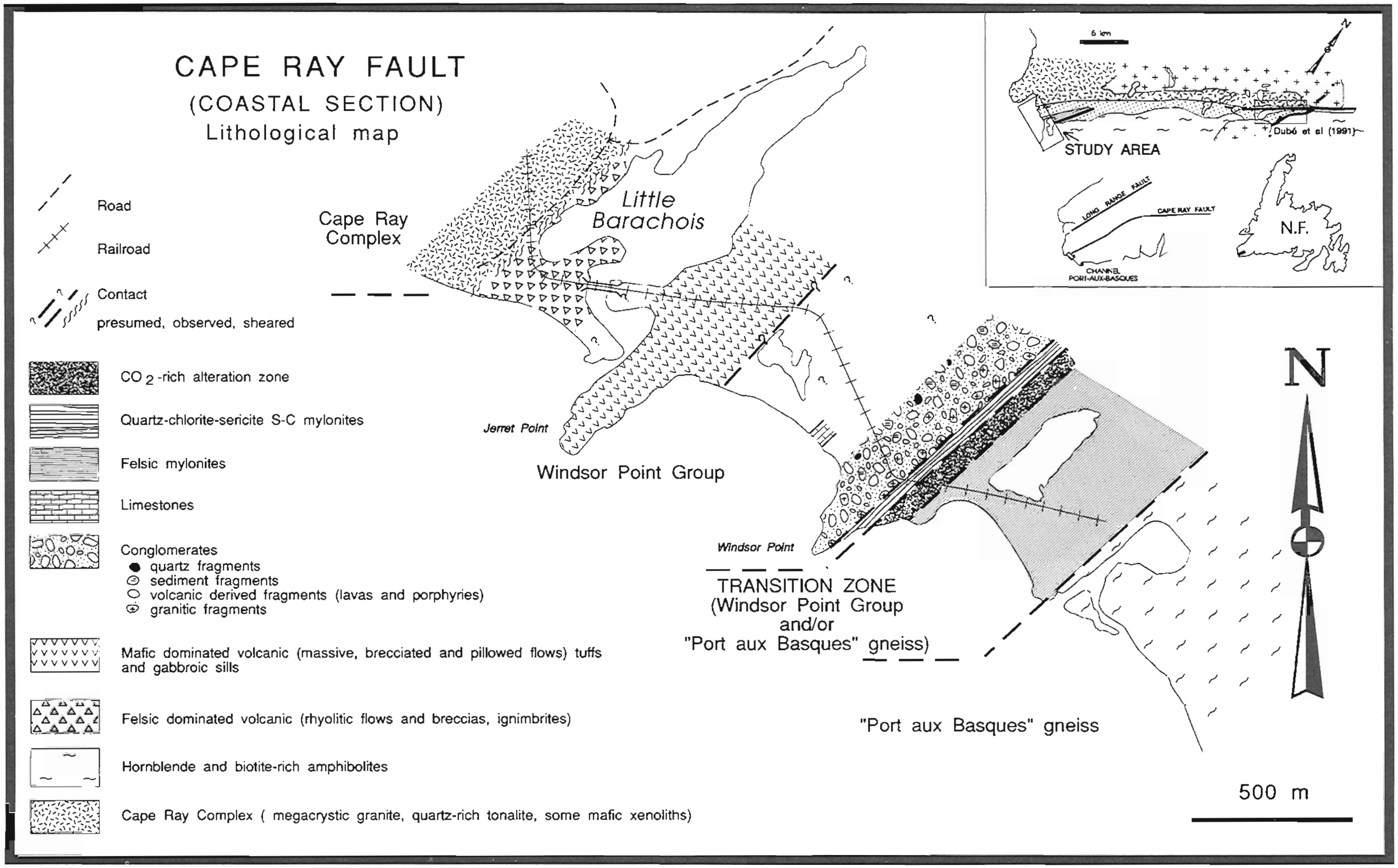


Figure 1. Geological map of the coastal section of Cape Ray fault. Inset shows a simplified map of the Cape Ray fault (Modified from Wilton, 1983).

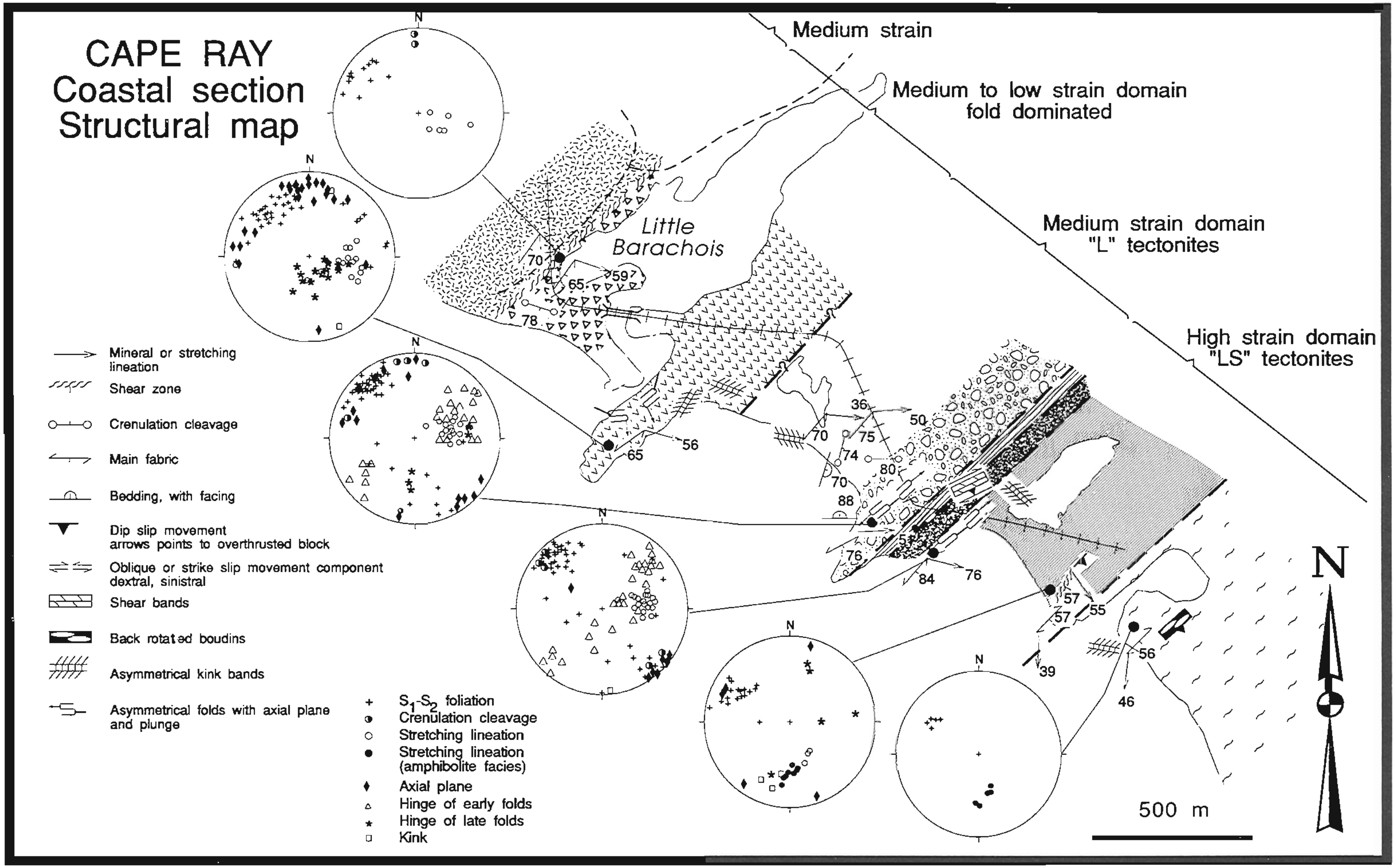


Figure 2. Structural map illustrating the major elements, kinematic indicators, and a compilation on equal area projections (lower hemisphere).

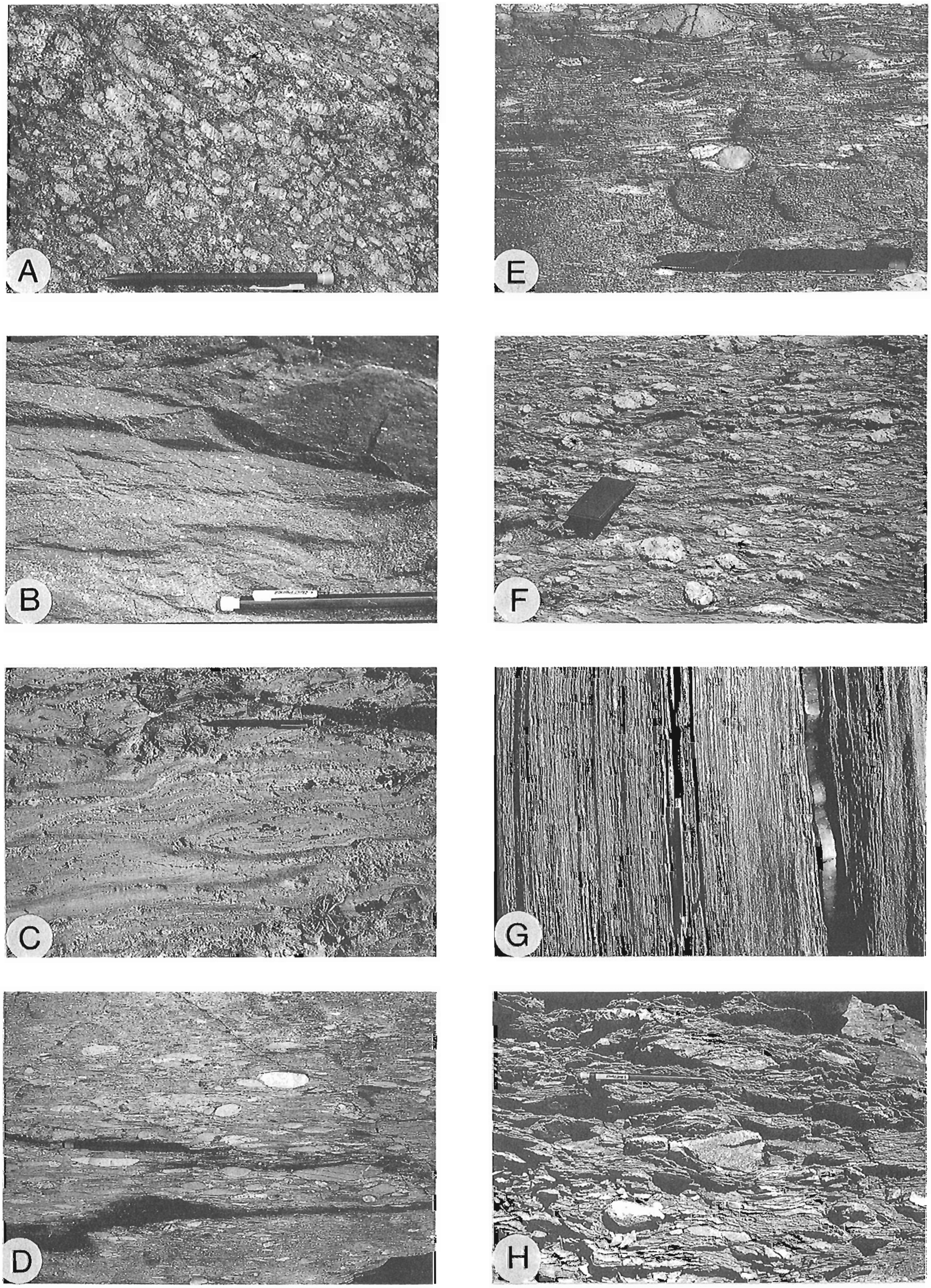


Figure 3. **A**) Megacrystic granitic phase of the Cape Ray Complex showing abundant microcline phenocrysts. **B**) Ignimbrite showing millimetre-scale quartz and feldspar crystals. **C**) Pillow basalt. **D**) Polymictic conglomerate characterized by rounded clasts. **E**) Rounded quartz pebble within the polymictic conglomerate. **F**) Volcanic conglomerate mostly made up of felsic volcanics and porphyries. **G**) Typical example of pink mylonite. **H**) Deformed granitic clasts preserved within the mylonite. All photographs are in plan view.

levels of tuff, pebbly sandstone, black shale, and siltstone. In the conglomerate, the relative proportion of clasts of different composition varies throughout the section. Northwest of Windsor Point, the conglomerate is polymictic and characterized by rounded quartz pebbles, but felsic volcanic and porphyry clasts are dominant and tuff, silstone, black shale, and ignimbrite clasts are also present (Fig. 3D,E).

At Windsor Point, rounded quartz clasts are absent in the conglomerate. Towards the northwest end of the Point, the conglomerate is mostly made up of felsic volcanic and porphyry clasts (Fig. 3F) like those described above, and is therefore better described as a polymictic volcanic conglomerate. It also contains tuff and siltstone clasts, small epidotized clasts and fine grained undeformed granitic clasts which are relatively rare on the northwest side of Windsor Point, but which become more abundant towards the southeast where they nearly constitute all of the clasts in the conglomerate. Interbedded epiclastic tuffs and sediments also decrease in thickness in this direction. Toward the southeast, the deformation also gradually increases in intensity and the volcanic conglomerate is transformed into a pinkish mylonitic rock in which some granitic clasts are locally preserved.

A strongly deformed limestone unit is present between the volcanic domain and the polymictic conglomerate. The unit is made up of millimetre-scale beige-yellow calcareous bands alternating with grey-green sedimentary or tuffaceous material. Contacts between the limestone and the adjacent units are not exposed.

3) Mylonite zone

A wide zone of banded, pinkish to beige and green mylonitic rocks is exposed between the volcanic conglomerate and the biotite-hornblende rocks of the "PaBC". These mylonites are characterized by a strong planar fabric and by millimetre-scale compositional layering defined by alternating fine grained, pinkish silica-rich layers and sericitic beige to chloritic green layers (Fig. 3G). Elongate plagioclase porphyroclasts (1-2 mm) are found locally. Petrographic studies indicate that this zone consists of mylonites and ultramylonites. They contain up to 80% small recrystallized quartz grains. Muscovite, calcite, traces of pyrite, and local feldpar porphyroclasts are also present. The pink mylonite is the dominant lithology on the northwest part of the mylonitic zone although it contains local metre-scale, dark green chloritic schist layers. Towards the southeast, near the biotite and hornblende-rich rocks, the beige mylonite is dominant, it is enriched in elongate feldspar porphyroclasts and contains biotite, suggesting an upper greenschist metamorphic grade.

The mylonites are injected by layer-parallel, or slightly oblique, discontinuous quartz veins with traces of K-feldpar, chlorite, and locally pyrite and/or iron carbonate. These veins are boudinaged. Sericite and less commonly pyrite are developed in adjacent wall rock.

A zone of CO_2 metasomatism of approximately 40 metres wide is developed within the mylonite zone (Fig. 1). This zone is characterized by the presence of iron carbonate within the rocks, and by the occurrence of ankerite and traces of pyrite within syn-shearing quartz veins. Chemical analysis of this carbonatized mylonite indicating up to 11.1% CO_2, and high values of Fe_2O_3 (up to 5.34%) indicate a strong iron carbonate related hydrothermal alteration (Table 1). There is no significant gold enrichment in the carbonatized mylonite or the associated quartz veins (Table 1). However, such a CO_2 alteration zone combined with its high K_2O content (up to 2.83%), and its location within a crustal scale fault may have important gold metallogenic implications since it could represent a fossil hydrothermal discharge zone across the CRF.

The protolith of these mylonites is not easily recognized because of the intensity of deformation and the probable diversity of units involved. However, local granitic clasts preserved within the mylonite (Fig. 3H) and the gradual transition between the volcanic conglomerate and the mylonite zone suggests that at least part of the pinkish mylonite is derived from the volcanic conglomerate. The composition of the pink mylonites suggests that the rhyolite of the WPG could be a likely protolith as suggested by Wilton (1984). Preliminary petrographic studies indicate the presence of polymineralic quartzofeldspathic porphyroclasts and suggest that a granitoid protolith is also a strong candidate.

NATURE OF CONTACTS BETWEEN LITHOLOGICAL UNITS

According to Wilton (1983) and Piasecki (1989), the contact between the Cape Ray granite and the ignimbrite of the WPG is an unconformity. But, the contact, as observed in the coastal section is sheared and best described as a sheared unconformity (see above).

The most problematic contact is the one between the WPG and rocks attributed to the PaBC. Both the overburden and well developed mylonite zone of the transition zone (Fig. 1), obscure the initial contact relationship. As discussed by Wilton (1983), prograde metamorphism within the WPG and retrograde metamorphism of the PaBC as well as the structural complexity of the area make it difficult to precisely

Table 1. Chemical analysis of carbonatized mylonites

Sample	CO_2%	K_2O%	Fe_2O_3%	Au ppb
BD-352-91	5.85	0.96	3.26	6
BD-353-91	4.95	2.27	3.54	<1
BD-354-91	1.04	0.73	3.34	4
BD-355-91	3.64	0.73	3.34	2
BD-356-91	10.1	2.83	5.25	8
BD-357-91	11.1	2.06	5.34	<1
BD-358-91	10.1	2.68	5.24	2

locate this contact. Along the coast, the biotite hornblende-rich rocks are characterized by intense planar fabrics and mineral lineations, and are not complexly folded as in the PaBC away from the CRF (see van Staal et al., 1992). If they are part of the PaBC they were strongly transposed along a northeast-trending fabric associated with an early, reverse-sinistral shearing event parallel to the CRF. It is not clear if the beige and pink quartzofeldspathic mylonites adjacent to the biotite- hornblende-rich rocks are part of retrograded PaBC, or if they are prograded WPG. We believe this zone is more probably composed of mixed slices of WPG and possible "PaBC" derived mylonites.

STRUCTURAL GEOLOGY

Three strain domains were recognized in the coastal section of the CRF. From southeast to northwest, they are: 1) a high strain domain, characterized by "mylonitic" "LS" tectonites, 2) a medium strain domain with "L>S" tectonites, and 3) a medium to low strain, "fold dominated" domain (Fig. 2). As mentioned earlier, the fold-dominated domain is bounded by a discrete, reverse-oblique ductile shear zone along the contact between the WPG and the CRIC.

1) High strain domain

The high strain domain is characterized by mylonitic "LS" tectonites. It has been subdivided in three subdomains on the basis of the structures developed. The first is located in the southwestern part of the section, and it comprises the biotite and hornblende-rich psammites and the beige quartzofeldspathic mylonites of the "transition" zone. It is characterized by a strong planar fabric oriented at N037/56° (Fig. 2) and well developed stretching and mineral lineations plunging moderately to the south-southwest (Fig. 4A). It is interesting to note that not all the hornblende crystals are oriented along the stretching lineation. Complex folding patterns typical of the PaBC away from the CRF are not observed. However, chevron-type, asymmetrical folds with a dextral vergence locally affect the mylonitic fabric. Their axial planes are slightly oblique to the fabric (N025/54°) and fold axes plunge moderately to the south-southwest (193/39°) as do the mineral lineations. Abundant extensional, subvertical quartz veins trending N030/65°, oblique to the planar fabric are inconsistent with the attitude of the stretching lineation. Locally developed kinematic indicators, such as back-rotated boudins (Fig. 4B) and C-S fabrics indicate a reverse-sinistral sense of shear (southeast-side-up). As shown by Piasecki (1989), the stretching lineations in the psammites and the beige quartzofeldspathic mylonites are at a relatively high angle to those produced during the post-Late Devonian retrograde shearing within the WPG. Indeed, the stretching lineation within the retrograded greenschist facies shear and within the WPG plunge steeply to moderately east (Fig. 2). The first signs of retrograde greenschist shearing are observed at the southern limit of the study area (Fig. 2) where a centimetre-scale retrograde shear zone is developed within beige quartzofeldspathic mylonites close to the biotite and hornblende-rich psammite. Elongate chloritic lineations within the shear plunge moderately to the east-southeast and associated shear bands indicate a reverse-dextral movement. Striations on the walls of extensional quartz veins are subparallel (086/52°) to the retrograde stretching lineation, and could be related to the same retrograde shearing.

The second subdomain is located in the central part of the high strain domain and it is separated from the first subdomain by a very strong topographic lineament which probably marks the contact between partly retrograded "PaBC" and mylonitized WPG rocks. It is characterized by a strong planar fabric associated with steeply to a moderately easterly-plunging stretching lineation. This well-exposed felsic mylonite strikes on average 054° and dips 72° to the south (Fig. 3G). It is at least 300 m wide. Non-coaxial deformation is indicated by "C-S" fabrics showing reverse-dextral movement. Over a few tens of metres, the trend of the stretching lineation is gradually rotated. From the southern limit of the second subdomain and moving north, the lineation trend varies from steeply plunging to the north-northeast (017/83°) to steeply plunging to the east-northeast (080/70°) to moderately plunging to the east-northeast (076/58°), the last being the average trend of the medium strain domains (Fig. 2).

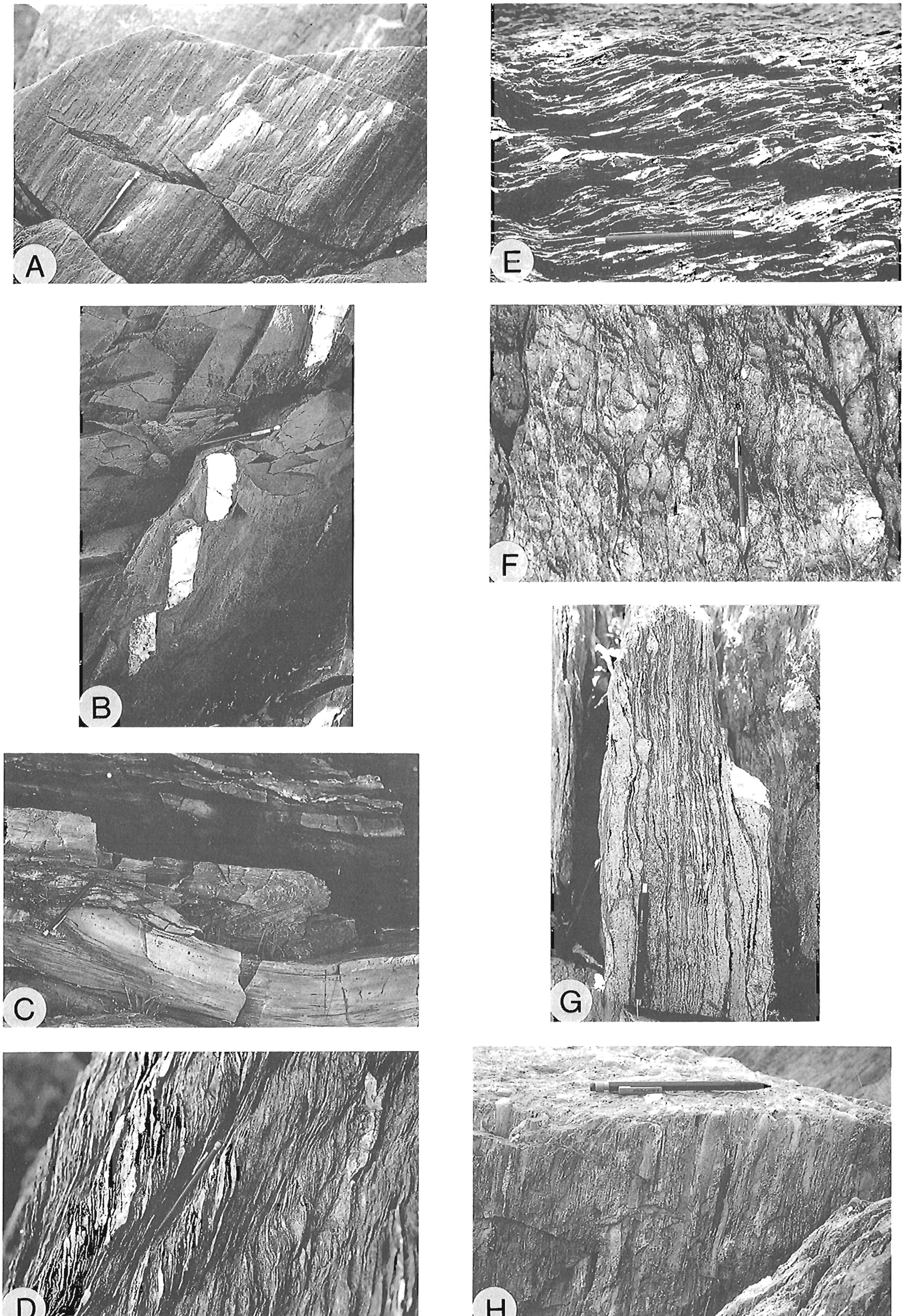
Centimetre-scale, tight, asymmetrical folds were observed in the mylonite (Fig. 4C). Their axial surfaces are subparallel to the trend of the mylonitic fabric, and to a locally well developed crenulation cleavage. Plunges of the fold hinges are variable (Fig. 2 stereonets). Some plunge steeply, either east (subparallel to the stretching lineation) or south and show a dextral vergence. Other hinges are shallow plunging (either northeast or southwest) and indicate a reversal sinistral or dextral vergence. Many subhorizontal joints (165/25°) are perpendicular to the stretching lineation.

To the northwest, the pink felsic mylonites grade into a zone of sericite-quartz-chlorite S-C mylonites, which make up the third subdomain. These mylonites are characterized by abundant and well-developed asymmetrical extensional shear bands (Hanmer and Passchier, 1991) (Fig. 4D,E). Shear bands are oriented on average at N062/60°, slightly oblique to the composite "C-S" foliation (045/75°). They are mainly observed in vertical section view but are also locally present in plan view (Fig. 4D,E). The intersection between the shear bands and the mylonitic foliation is almost perpendicular to the stretching lineation (062/52°) indicating that they can be used as evidence for a reverse-dextral movement with a southeast-side-up sense of shear.

These shear bands are developed within a volcanic conglomerate protolith where the "granitic" clasts have aspect ratios ranging from 3/1 to 10/1 in the oblique plane perpendicular to the stretching lineation to 10/1 to 15/1 in the subvertical plane corresponding with the stretching of the long axes of the fragments in the vertical plane (Fig. 4F,G).

2) Medium strain domain

To the northwest, the mylonitic "LS" high strain domain grades into a predominantly "L>S" tectonite, medium strain domain, which in turn is followed to the northwest by a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fold-dominated medium to low strain domain (Fig. 2). Between these two domains there is a small exposure of highly strained limestone.

"L>S" tectonite domain

The medium strain domain is characterized by tectonites with L>S. The lineations plunge on average 050° east-southeast, as defined by stretching of clasts and pebbles in the conglomerate (Fig. 4H). The granitoid clasts have aspect ratios ranging from 2/1 to 4/1 in the horizontal plane to 5/1 to 20/1 in the vertical plane, suggesting a strain regime in or near the constrictional field (Flinn, 1962), although, a northeast-trending fabric is defined by flattening of clasts and pebbles and by a schistosity.

Locally preserved cleavage-bedding relationships, as well as opposing younging directions, indicated by graded bedding and scour structures, suggest the presence of a large, overturned syncline plunging subparallel to the stretching lineation in this domain.

Many small-scale folds affect the fabric. The earlier ones (F_2) are relatively tight and coplanar to the general trend of the fabric. Some plunge subparallel to the stretching lineation, but most F_2 folds diverge from it. F_2 axes vary from shallow, southwesterly-plunging (Fig. 5A) to moderately steep to the northeast. Some F_2 folds are even curvilinear and clearly reorient the stretching lineation (Fig. 5B).

As recognized by Wilton (1983), two sets of "late" axial plane cleavages and local kinks and folds, are developed at a high angle to the main fabric. Centimetre- to metre-scale east-west F_3 sinistral and reverse verging folds are related to a well developed S_3 axial plane cleavage trending approximately 085/80° with shallow plunging fold axes (090/38°), whereas north-northeast dextral verging F_3' folds plunge moderately to the southwest (190/38°) and have an S_3' axial plane cleavage oriented at 025/80 (Fig. 5C). Systematic crosscutting relationships, between the two sets of cleavage are not observed but the east-west S_3 cleavage is better developed. Locally, the north-northeast-trending S_3' cleavage was seen to crosscut the east-trending cleavage.

Figure 4. A) Longitudinal view of the stretching lineation developed within the beige quartzo feldspathic mylonite of the "transition zone." **B**) Vertical section view showing back rotated quartz boudins. **C**) Plan view of F_2 folds with axial plane subparallel to the mylonitic fabric and hinges subparallel to the stretching lineation. **D**) Vertical section view of shear bands indicating a reverse movement. **E**) Plan view of shear bands indicating a dextral component of movement. **F**) Oblique view, perpendicular to the stretching lineation showing the relatively weak flattening of the clasts in the volcanic conglomerates **G**) Section view subparallel to the stretching lineation showing highly stretched clasts within the volcanic conglomerate. **H**) Vertical section view of the L>S tectonites showing the strong vertical stretching of clasts in the conglomerate.

3) Medium to low strain domain

A fold-dominated, medium to low strain domain was developed within the WPG volcanics (Fig. 2). It is characterized by centimetre- to metre-scale, steeply to moderately inclined, close to tight sinistral F_3 folds. Their axial surfaces trend east-west and fold axes plunge either subparallel to the dominant easterly-trending stretching lineation (100/50°) or to the south (180/70°). North-northeast-trending dextral F_3 folds plunging to the south are locally found. F_3 folds have similar orientations to that of the S_3 cleavage. At one site, east-trending sinistral F_3' folds refold an older centimetre-scale F_2 fold subparallel to the northeast fabric and coaxial with the stretching lineation. As in other domains, the dominant stretching lineation trends to the east and plunges moderately although south-plunging stretching lineations were locally observed. Subvertical en echelon extensional veins trending N300/54° locally suggest a dextral motion.

Kink bands

A conjugate set of steeply-dipping, east-west sinistral and north-northeast-trending local dextral kink bands, respectively, subparallel to F_3 and F_3 ' folds, is developed in all the structural domains described above (Fig. 5D). All kink bands have a reverse-kink geometry (Dewey, 1965) and record layer-parallel shortening. As deduced from the conjugate bisector method (Ramsay, 1962), the principal strain axes are oriented as follow: Z = 046/75°, Y = 169/72° and X = 141/80°. The anisotropy exploited during kinking is generally the subvertical S_1-S_0, although S_2 is also locally implicated. As for the relationship between the east-west and north-northeast S_3-S_3' crenulation cleavages, in one area the north-northeast dextral kink bands refold the sinistral east-west ones. Striations that plunge shallowly to the southwest with local sinistral steps are locally present on the main fabric and are probably related to the development of these kink bands.

DISCUSSION

Two shearing stages are recorded in the biotite-hornblende-rich rocks and the beige quartzo-feldspathic mylonites of the transition zone, but do they both affect the WPG? The variable attitude of stretching lineations in the pink mylonites (which are thought to have been derived from WPG volcanic conglomerate and/or rhyolite and granitoid), suggest that rocks of the WPG are affected. Trend variations of the stretching lineation could result from the reorientation of a moderately south-plunging lineation (related to a reverse-sinistral movement) by a later reverse-dextral movement producing an easterly-plunging lineation. This interpretation is consistent with locally observed superimposed folding relationships of east-west sinistral F_3 folds affected by north-northeast dextral F_3' folds. As well, east-west S_3 cleavage is locally crosscut by the north-northeast-trending S_3' cleavage. The local presence of southerly plunging stretching lineations and fold axes within the folded domain are also consistant with such an interpretation. On a regional point of view, the east-west F_3 and north-northeast F_3' folds could be respectively correlated with the F_3 and F_4 folds found by van Staal et al. (1992) in the PaBC outside the CRF. They propose that these folds record a change from a sinistral to a dextral transpressive regime. However, within the CRF, such crosscutting relationships of folds and cleavages were only locally observed and more work has to be done to verify if they are systematic and significant.

Variations in trend of the stretching lineation need not necessarily be the result of two deformation stages. Variations of up to 40° in lineation trend have been documented elsewhere in mylonite zones (Bell and Hammond, 1984; Hudleston et al., 1988) and are interpreted to result from a single progressive phase of deformation. A strong analogy can be drawn between structures of the CRF and those recorded in an Archean dextral transpression zone studied by Hudleston et al., (1988). As in the CRF, Hudleston et al., (1988) have recorded significant variations in the strain pattern with steeply east-plunging mineral lineations in the constrictional strain regime away from the Vermilion fault to west-plunging lineations in the flattening regime developed adjacent to the Vermilion fault. As well, the hinges of asymmetrical F_2 folds and associated subparallel mineral lineations plunge moderately to steeply to the east or west. Hudleston et al. (1988) have interpreted these variations as the result of a process of continuous shear, with pertubations of flow.

It has already been proposed by Piasecki (1989) that the southeast-plunging stretching lineations, developed in what we have called the biotite- and hornblende-rich psammites and the beige quartzo feldspathic mylonite of the "transition zone", were the result of a pre-late Devonian (pre-WPG), reverse-sinistral faulting between the "PaBC" and the Cape Ray granite, whereas a post-Late Devonian, reverse-dextral shearing event affects both the PaBC and the WPG. This model is rather similar to that of Chorlton's (1983), in that it recognized two distinct faulting events within the CRF system. The main problem with such an interpretation is the lack of clear evidence of an early reverse-sinistral shearing event within the CRIC, and within the PaBC zone in the IAMR area (Dubé et al., 1991).

Neither model completely explains all the data collected along the CRF. Given the intensity of the reverse-oblique ductile deformation observed within the WPG rocks, the grade of metamorphism, and its volcano-sedimentary nature including presence of pillow basalt, the Devonian age of the WPG, obtained by the imprecise Rb-Sr whole-rock isochron dating method (Wilton, 1983) and by the correlation with plant-fossil-bearing strata in the Billiards Brook Formation

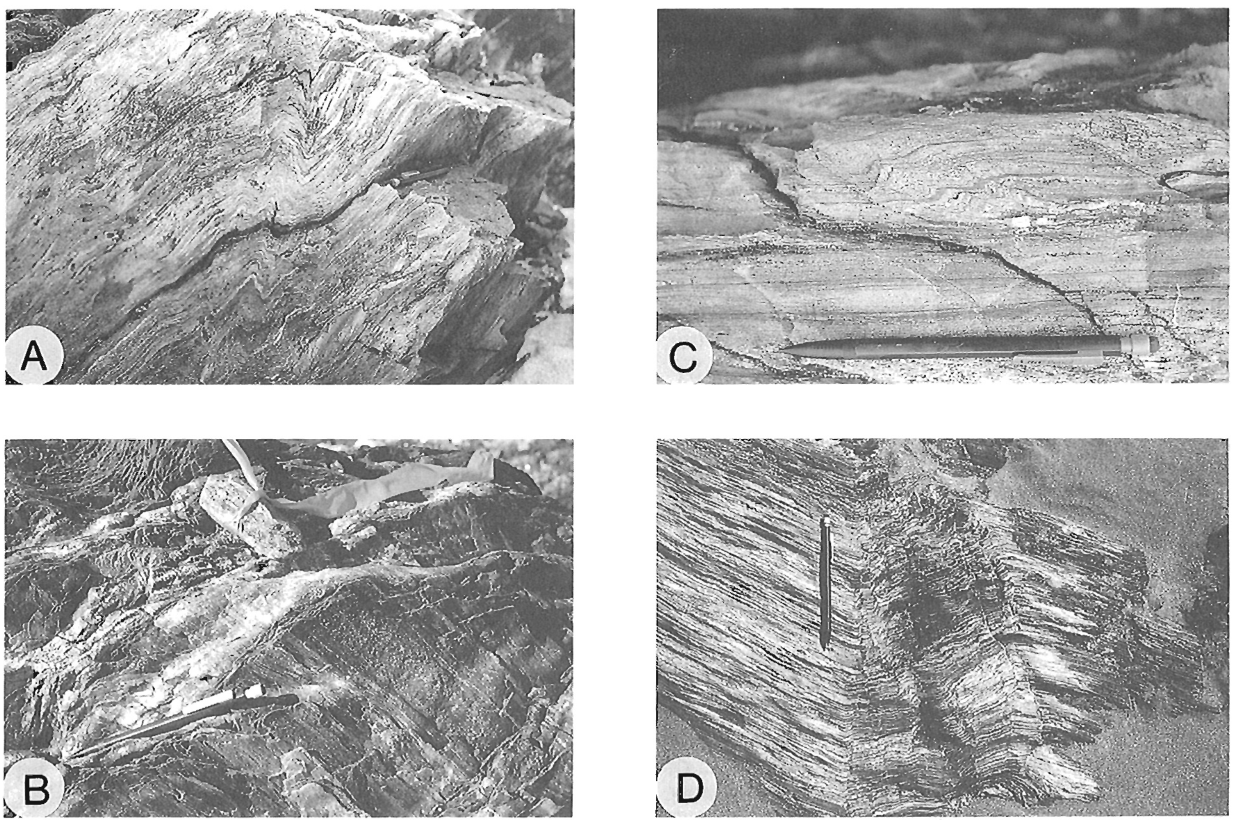

Figure 5. A) Vertical section view showing an asymmetrical fold with a shallow plunge. **B**) Oblique view showing a fold with a curvilinear hinge, deforming the stretching lineation. **C**) Plan view showing an S_3 axial plane cleavage of a centimeter scale sinistral F_3 fold. **D**) Steeply-dipping east-west sinistral kink bands in the limestone.

(Chorlton, 1983), is suspect. Ongoing field work in the area and geochronological U-Pb zircon dating of the WPG ignimbrite and the different granitic intrusions as well as examination for conodonts in the WPG limestone will be of regional tectonic importance and will hopefully resolve the uncertainties that this study has raised.

CONCLUSIONS

The coastal section of the CRF exposes a crustal-scale fault zone profile. Kinematic indicators in non-coaxial fabrics indicate that high grade (amphibolite) deeper crustal-level rocks of the "PaBC" were thrust over medium to low grade (greenschist) supracrustal rocks of the WPG. But the kinematics of the deformation is complex with evidence of localized reverse-sinistral overthrusting of amphibolitic rocks over greenschist rocks and a predominant reverse-dextral thrusting recorded within the WPG rocks as suggested by the variation in the stretching lineation trend and by the timing relationship (Fig. 2).

The internal structures recorded within WPG rocks are typical of geometrical complexities found in mylonite zones (Bell and Hammond; 1984), and could result from a single mylonitization stage. Complex relationships of folds, lineations, and foliations can be explained in terms of an extreme heterogeneity of strain and the anastomosing character of the shear-related deformation (Bell and Hammond, 1984).

Compared to the IAMR area, there is no suggestion of strike-slip movement in the CRF along the coastal section (Dubé et al., 1991). However, as it is a reactivated fault zone, different deformation regimes may result from the previous geometry of the fault zone. The flexure of the CRF to the northeast of the studied area (see inset Fig. 1), may be responsible for the variation of the deformation regime and the consequent development of strike slip structures in the IAMR area (Dubé et al., 1991). As well, heterogeneity of strain produced, for example, by stiff layers such as the WGH granite present in the IAMR area, could induce local variations in the strain regime along the length of the CRF.

From a gold metallogenic point of view, the discovery of a wide zone of CO_2 metasomatism within the mylonites of a crustal scale fault is important and could represent a major fossil hydrothermal discharge zone across the CRF. Ongoing work will address this particular aspect.

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Geological Survey of Canada Project 890026

Structural control of sill-hosted gold mineralization: the Stog'er Tight gold deposit, Baie Verte Peninsula, northwestern Newfoundland[1,2]

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Abstract

Structural elements within the Stog'er Tight Gold deposit have been studied in order to determine the structural history of the area, the controls on gold mineralization, and its relationship to the major faults. The Stog'er Tight deposit, a mesothermal altered wallrock gold deposit, is hosted by differentiated gabbroic sills within the Lower Ordovician Point Rousse Complex cover sequence. Three phases of deformation affected the rocks. The first phase (D_1) consisted of an intense ductile shearing which produced the regional foliation (S_1). The development of quartz-carbonate veins and associated alteration and mineralization is ascribed to a late (postductile) increment of D_1. The second event (D_2) produced south-verging asymmetric folds (F_2), ductile-brittle, high-angle faults and a fracture cleavage. Late north-northeast- to northeast-trending, broad open folds (F_3) refold the older structures. This study demonstrates the structural and lithologic influence of the hosting gabbroic sill on the mineralization.

Résumé

Les éléments structuraux présents dans le dépôt aurifère de Stog'er Tight ont été étudiés dans le but de déterminer l'histoire structurale de la région, les contrôles de la minéralisation aurifère et leur relation avec les zones de failles. Le dépôt de Stog'er Tight, un dépôt mésothermal compris dans les épontes altérées, est encaissé par des filon-couches gabbroïques différenciés compris dans une séquence de couverture datant de l'Ordovicien inférieur, à savoir le complexe de Point Rousse. Trois phases de déformation ont affecté les roches de la région. La première phase (D_1) consiste en un cisaillement ductile intense qui a produit la foliation régionale (S_1), le développement de veines de quartz et carbonates et l'altération hydrothermale associée. La minéralisation est reliée à une phase tardive (tardi-ductile) de la déformation D_1. Le deuxième événement (D_2) a produit des plis (P_2) asymmétriques à vergence vers le sud, des failles ductiles fragiles fortement inclinées et un clivage de fracture. Des failles tardives nord-nord-est à nord-est et des plis ouverts (P_3) replissent les structures plus anciennes. Cette étude démontre la forte influence structurale et lithologique du filln-couche encaissant sur la minéralisation.

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INTRODUCTION

This report summarizes the results of detailed mapping at a scale of 1: 1250 by the senior author, on the Noranda/International Impala Ressources Stog'er Tight gold deposit. This project, mainly concerned with the structural setting of the deposit and its relationship to major fault zones, is part of a large-scale study of gold deposits in the Canadian Appalachians (Dubé, 1990a). The Baie Verte Peninsula has historically been an active mining district, its reputation being built on volcanogenic massive sulphide deposits. However, significant discoveries during the 1980s shifted the interest to gold exploration. Major gold-only deposits in the peninsula have been morphologically divided into two main types: 1) disseminated stratabound sulphide gold deposits and 2) mesothermal vein-type (Dubé, 1990a). The Stog'er Tight prospect is included in the latter type. Good exposure, the result of systematic trenching, permitted detailed mapping of the property. The objectives of the project were to determine the structural history of the area, the controls on gold mineralization, and its relationship to the major fault zones of the Baie Verte Peninsula. Previous work done in the area include regional studies by Hibbard (1983), property-scale work by Huard (1990) and metallogenic studies by Wilton and Evans (1991).

REGIONAL GEOLOGICAL SETTING

The Stog'er Tight gold deposit is located on the Baie Verte Peninsula in northwestern Newfoundland, 10 km east-northeast of the town of Baie Verte (Fig. 1), within volcanic rocks of oceanic affinity included in the Dunnage Zone (Hibbard, 1983; Huard, 1990). In this area, the Dunnage Zone consists of the Point Rousse Complex and the Pacquet Harbour Group of Lower Ordovician age.

The Point Rousse Complex is composed of a complete, though dismembered ophiolite conformably overlain by a dominantly-mafic volcanic-volcaniclastic cover sequence (Hibbard, 1983). These rocks form a broad east-trending syncline and have been metamorphosed to the greenschist facies. Rocks of the cover sequence occupy the central part of the syncline and are bordered to the north and to the south by the ophiolitic intrusive components. The complex is tectonically bounded (Hibbard, 1983). The contacts vary from thrust faults to high angle faults. To the south, the Point Rousse Complex was thrust over rocks of the Pacquet Harbour Group along the Scrape Thrust, a major southeasterly-trending, north-dipping thrust fault (Fig. 1).

GEOLOGY OF THE STOG'ER TIGHT GOLD DEPOSIT

The Stog'er Tight deposit is hosted by differentiated gabbroic sills intruding the mafic volcanic-volcaniclastic cover sequence of the Point Rousse Complex. Rocks of the northern part of the study area consist of mafic volcanics which overlie a sequence of volcaniclastic rocks. The gabbroic units intrude this sequence in the form of conformable sills. Intense deformation and hydrothermal alteration in the central part of the study area have obliterated original textures and mineralogy to such an extent that original lithologies are relatively difficult to determine.

Volcanic-volcaniclastic cover sequence

In the study area, pyroclastic rocks dominate the cover sequence. These rocks include fine ash tuff to pyroclastic breccia with fragments ranging in size from 0.1 mm to 10 cm. The larger, centimetre-sized clasts consist mainly of porphyritic volcanic rocks. In thin section, most of the ash tuff is composed of rounded fragments of quartz and feldspar in a fine grained, microlithic groundmass suggesting a weathered pyroclastic or epiclastic origin. In the northern part of the area the pyroclastics are overlain by mafic volcanic rocks which consist mainly of massive and pillowed flows as well as iron-formation (Fig. 2).

Gabbro sills

The gabbroic rocks which hosts the gold mineralization have been mapped as three distinct sills within the pyroclastic rocks (Fig. 2). The upper northern contact between the gabbro sills and tuffaceous rocks usually displays a chilled margin whereas the lower or southern contacts are usually slightly to moderately sheared. In places, the sills show a distinct magmatic layering. The gabbro sills contain four megascopically distinct units: a basal melagabbro, a leucogabbro, a ferro-leucogabbro, and a quartz-ferrogabbro. These units have been distinguished in the eastern part of the study area where the gabbroic rocks are the least altered. Field criteria used in identifying the gabbro units are: more than 50% ferro-magnesian minerals (melagabbro), less than 50% ferromagnesians and between 3-8% leucoxene (leucogabbro), very coarse grained plagioclase crystals and abundant titano-magnetite in the leucogabbro (ferro-leucogabbro), and more than 1% of visible quartz (quartz ferrogabbro).

In the southernmost sill, the basal gabbro is strongly magnetic and contains two distinct units, a fine grained melagabbro and an overlying medium grained leucogabbro with 3-5% leucoxene as an alteration product of primary titano-magnetite. The minimum thickness of the southernmost sill is approximately 25 to 30 m, the lower contact being either unexposed or sheared. Two units have also been recognized within the middle gabbro; a basal fine- to medium-grained melagabbro overlain by a medium- to coarse-grained leucogabbro containing 3-7% leucoxene. The thickness of the middle gabbro is approximately 40 m. The northernmost or upper gabbro seems to thicken eastward where it is separated from the middle gabbro by two very narrow (less than 7 m thick) pyroclastic units. These tuffaceous units do not extend very far towards the northwest. This fact coupled with the presence of many sheared contacts suggest that the upper gabbro is probably repeated by faulting in the eastern part of the study area. Nonetheless, two distinct units are recognized within the upper gabbro. A fine- to coarse-grained leucogabbro containing 3-8% leucoxene, and a coarse grained ferro-leucogabbro containing up to 10%

titano-magnetite. The ferro-leucogabbro unit is contained within the leucogabbro unit which becomes progressively finer grained near the upper contact with the pyroclastic rocks. A fourth unit has been recognized in the central part of the study area, on the Gabbro Zone outcrop, a 3 m thick, coarse grained quartz-ferrogabbro is in fault contact with a leucogabbroic unit of the middle gabbro. This quartz-rich unit is probably part of the upper gabbro, thrust upon the middle gabbro. The thickness of the upper gabbro, which is increased by faulting in the east varies from 45 to 100 m.

In thin section, the unaltered gabbro units, metamorphosed to greenschist facies, are characterized by varying amounts of plagioclase, amphibole (actinolite), epidote, chlorite, titanomagnetite, leucoxene, carbonate, quartz, and sericite. The strongly foliated gabbro consists of varying amounts of undeformed plagioclase and epidote porphyroclasts set in a matrix of chlorite, amphibole, sericite, and recrystallized quartz and plagioclase.

Mafic dykes

In the northeastern part of the study area, the pyroclastics and gabbro units are cut by mafic dykes up to 2 m wide. They have a porphyritic texture defined by plagioclase phenocrysts in a fine grained groundmass. Many of the dykes that intrude the sills contain angular inclusions of gabbro up to 15 cm in size.

STRUCTURAL GEOLOGY

The Stog'er Tight gold deposit is located on the south limb of the regional easterly-trending syncline of the Point Rousse Complex (Fig. 1). Rocks in the study area define a series of north-northeast-trending anticlines and synclines that postdate the major regional syncline. The structures within the study area are dominated by a well developed regional foliation (S_1) that contains a down-dip stretching lineation. The lithologic contacts and main tectonic foliation are

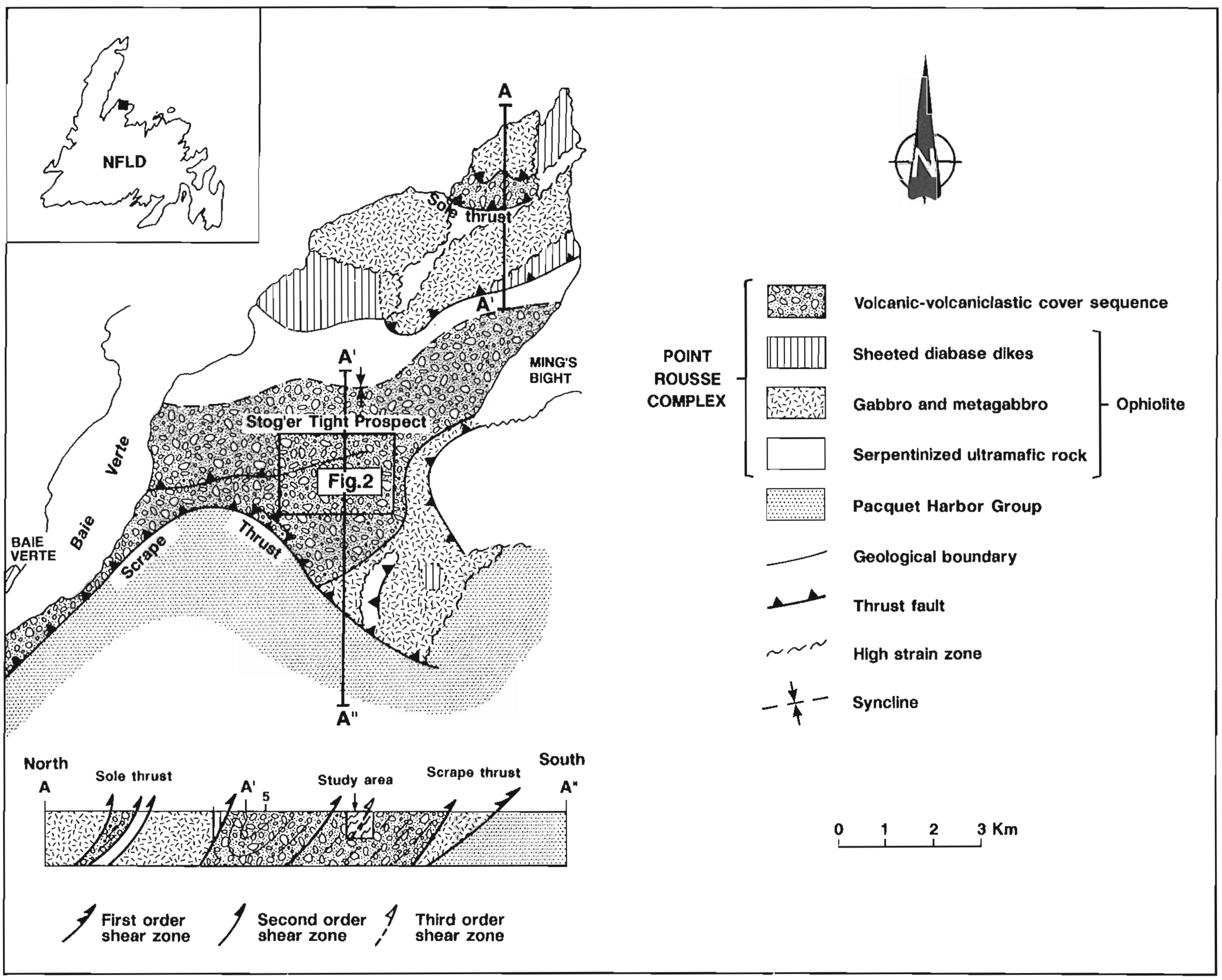

Figure 1. Localization of the study area and geology of the Point Rousse Complex. Regional north-south cross-section of the Point Rousse Complex, showing the structural setting of the Stog'er Tight gold deposit (adapted from Hibbard, 1983).

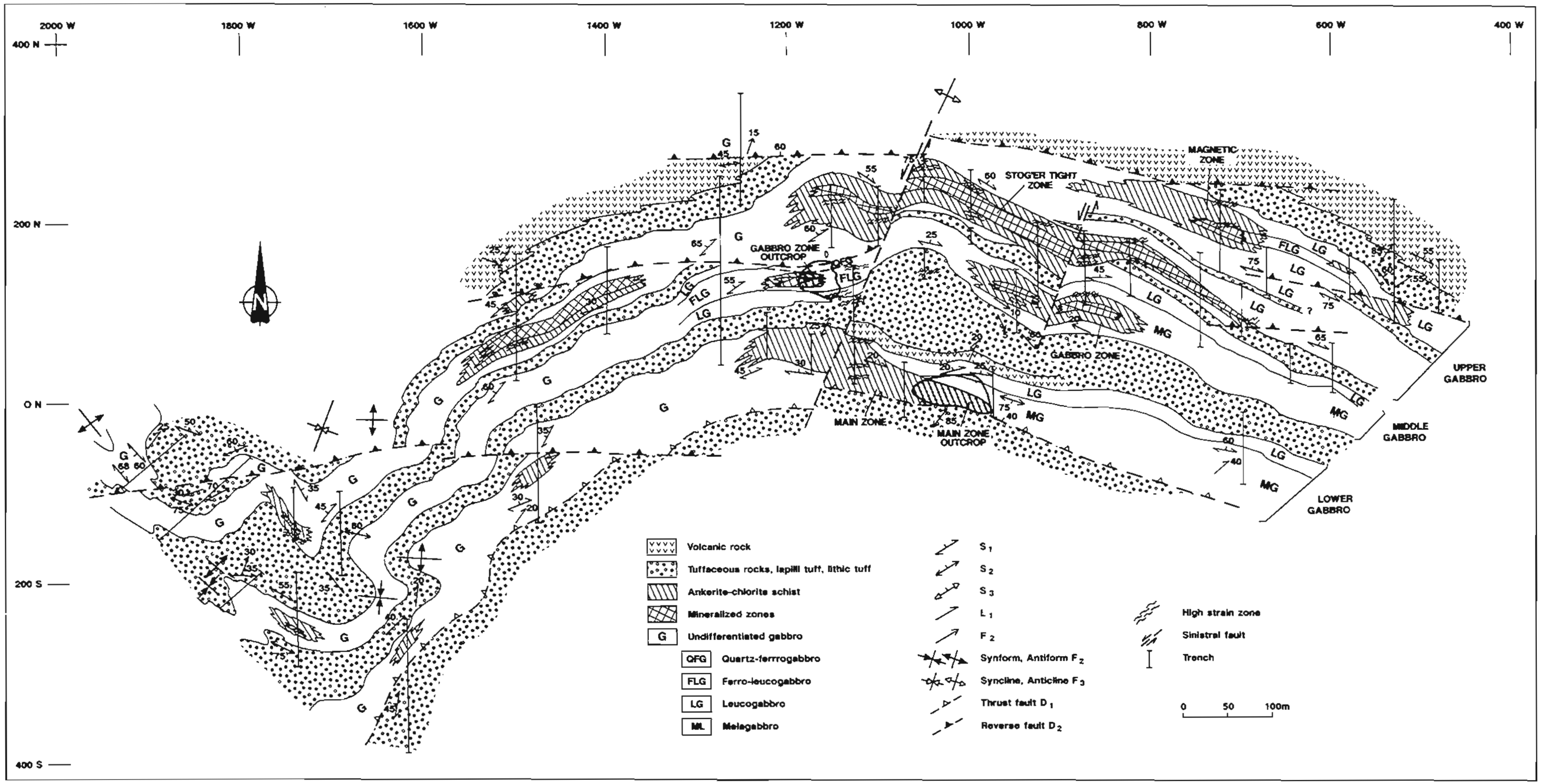

Figure 2. Geological map of the study area. (Modified from compilation maps of Noranda Exploration)

subparallel and generally dip to the north. Overprinting relationships at the outcrop scale indicate that two distinct generations of fold and cleavage followed the development of the regional foliation (Fig. 2). The regional foliation and the earlier structures have been assigned to three deformational generations termed D_1, D_2, and D_3.

The first phase of deformation (D_1) consisted of an intense ductile shearing, with no major folds observed in the study area, which produced the regional foliation (S_1) and downdip stretching lineation (L_1). This event is interpreted to be associated with the regional easterly-trending syncline affecting the Point Rousse Complex (Hibbard, 1983). The second event (D_2) produced south-verging asymmetric folds (F_2) and brittle-ductile, high-angle faults oriented roughly east-west, parallel to the F_2 fold axial surfaces and the S_2 axial planar cleavage. Late north-northeast- to northeast-trending, broad open folds (F_3) refold the F_2 folds. Late-stage brittle faults characterized by horizontal striae on slickensides slightly displace D_1 and D_2 structures as well as the mineralized zones. The main alteration and mineralization are associated with the D_1 structures and fabrics.

D_1 Event

The major D_1 deformation was responsible for the presently northerly dip of the succession due to large-scale regional folding. The study area is situated on the right way up limb of an east-trending, close to tight, F_1 syncline slightly overturned to the south-east (Hibbard and Gagnon, 1980; Fig. 1). The regional S_1 fabric is a well developed foliation that dips to the north between 25° and 40° and contains a typical downdip stretching lineation. The foliation is generally bedding-parallel, but is locally slightly oblique to bedding. Small scale, isoclinal folds with an axial-planar S_1 cleavage have been locally observed, although no large-scale F_1 fold has been recognized in the study area.

Throughout the area, the S_1 foliation is best developed in the volcaniclastic rocks. Near major ductile shear zones, the S_1 fabric intensifies and within a few metres, grades into a mylonitic foliation. Between the shear zones, strain is usually low in the more competent gabbro sills and the S_1 foliation is only weakly developed. Near these shear zones, the usually undeformed gabbroic sills become intensely foliated and develop a strong downdip stretching lineation, producing a fabric characterized by grain-shape lineation or L-tectonites. Within most of the high strain zones, the rocks are generally fine grained, laminated, and contain small fragments of undeformed gabbro displaying primary textures, thus indicating a gabbroic protolith (Fig. 3a). D_1 shear zones are generally metre scale in width (1 to 3 m), and trend subparallel to the S_1 foliation with a moderate dip (40° to 60°). In thin section, the sheared gabbro varies from a protomylonite to an ultramylonite. In places, narrow shear zones less than 0.5 m in width developed at the contact between the gabbro and the pyroclastic rocks. These shear zones are parallel to the major D_1 shear zones and are characterized by a protomylonitic fabric. The downdip stretching lineation (Fig. 3b) as well as megascopic and microscopic fabrics strongly suggest that north over south thrusting was responsible for the development of the first generation of structures. The noncoaxiality of the strain is indicated by shear bands and asymmetrical tails on rotated plagioclase porphyroclasts.

The development of quartz-carbonate veins and associated alteration and mineralization is ascribed to a late (postductile) increment of the D_1 event. Although D_1 shear zones do not show clear signs of hydrothermal alteration, subvertical extensional quartz-carbonate veins within the mineralized zones are perpendicular to the L_1 lineation (Fig. 4). Moreover, alteration zones with intense veining are spatially associated to D_1 ductile shear zones, being adjacent to them. Microscopically, carbonate minerals occur as idioblastic grains aligned more or less subparallel to L_1, and have been partially dissolved within S_2 cleavage planes. Locally, carbonate porphyroclasts show an internal schistosity which is continuous with the matrix schistosity (S_1) implying growth of carbonate after the formation of S_1 and confirming their postkinematic nature. These relationships suggest that the precipitation of hydrothermal carbonates associated to mineralization is late-D_1 and pre-D_2.

D_2 event

The D_2 deformation is characterized by mesoscopic asymmetric kink folds with 15 to 70 cm amplitudes which folded the S_1-L_1 fabric, D_1 shear zones, and related alteration zones (Fig. 3c). These southward-verging F_2 folds plunge from between 5° and 40° to the west to west-northwest, and trend east to east-southeast. S_2 axial planar cleavage dips steeply towards the north and cuts the S_1 foliation. Numerous, steeply dipping, east- to east-southeast-trending kink bands, subparallel to the F_2 axial planes, occur and record north-over-south movement. A less common set of near-horizontal, south-dipping kink bands also occur in the strongly foliated rocks. Small-scale chevron folds are developed where both sets of kink bands are present. Within the strongly foliated rocks, a crenulation cleavage subparallel to S_2, folds the S_1 surface and is affected by F_2 folds and kink bands.

Shear zones developed during D_2 record evidence of brittle-ductile behavior. Shear zones are defined by narrow, subvertical chloritic zones between 10 and 40 cm wide, characterized by a millimetre-spaced schistosity which drags and displaces the S_1 foliation and cuts across the alteration zones. Their orientation is generally consistent, striking east-southeast and dipping 70° to 80° to the north. The major D_2-related faults were identified from unpublished magnetic anomaly company maps as well as in drill core sections where displacements up to 20 m of the sills and alteration zones have been noted.

D_3 event

The D_3 deformation consisted of the development of broad open folds and a related S_3 fracture cleavage trending north-northeast to northeast. These F_3 folds affect D_1 and D_2-related structures and are expressed as a series of broad, regional north-northeast-trending anticlines and synclines.

GOLD MINERALIZATION

Based on structural and lithologic characteristics of the ore zone, Dubé (1990a) classified the Stog'er Tight prospect as a mesothermal altered wallrock gold deposit. Quartz veins in the Stog'er Tight deposit are a minor component and the ore zones consist mostly of secondary albite, pyrite, ankerite, sericite, and gold-rich altered gabbro.

Nature and geometry of alteration zones

As a result of alteration with wallrock and the time and space evolution of the hydrothermal fluid, at least three distinct alteration zones can be megascopically recognized. The dominant mineralogical association zonation in each zones was used to define three types of alteration zones: I) ankerite-chlorite, II) albite-chlorite, and

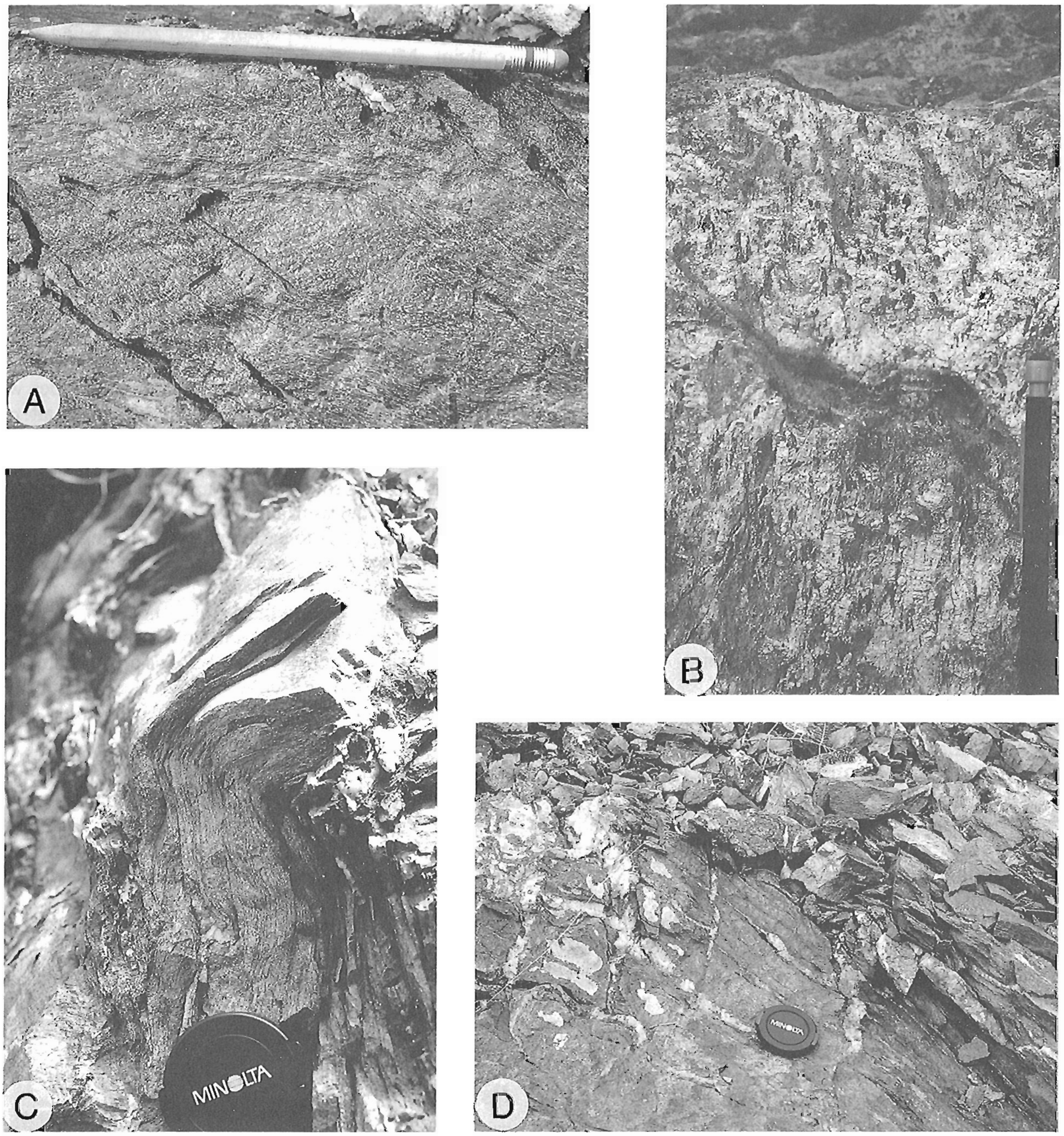

Figure 3. A) Fine grained, laminated gabbro within the high strain zone containing small fragments of less deformed gabbro in the central part of the photograph (arrow points to less deformed gabbro). **B**) Longitudinal view along foliation plane showing the downdip stretching lineation in a ferro-leucogabbro. Note the stretched magnetite crystals. **C**) Mesoscopic asymmetric F_2 kink folds overprinting the S_1-L_1 fabric. **D**) Type Va and Vb mineralized veins affected by asymmetric F_2 folds.

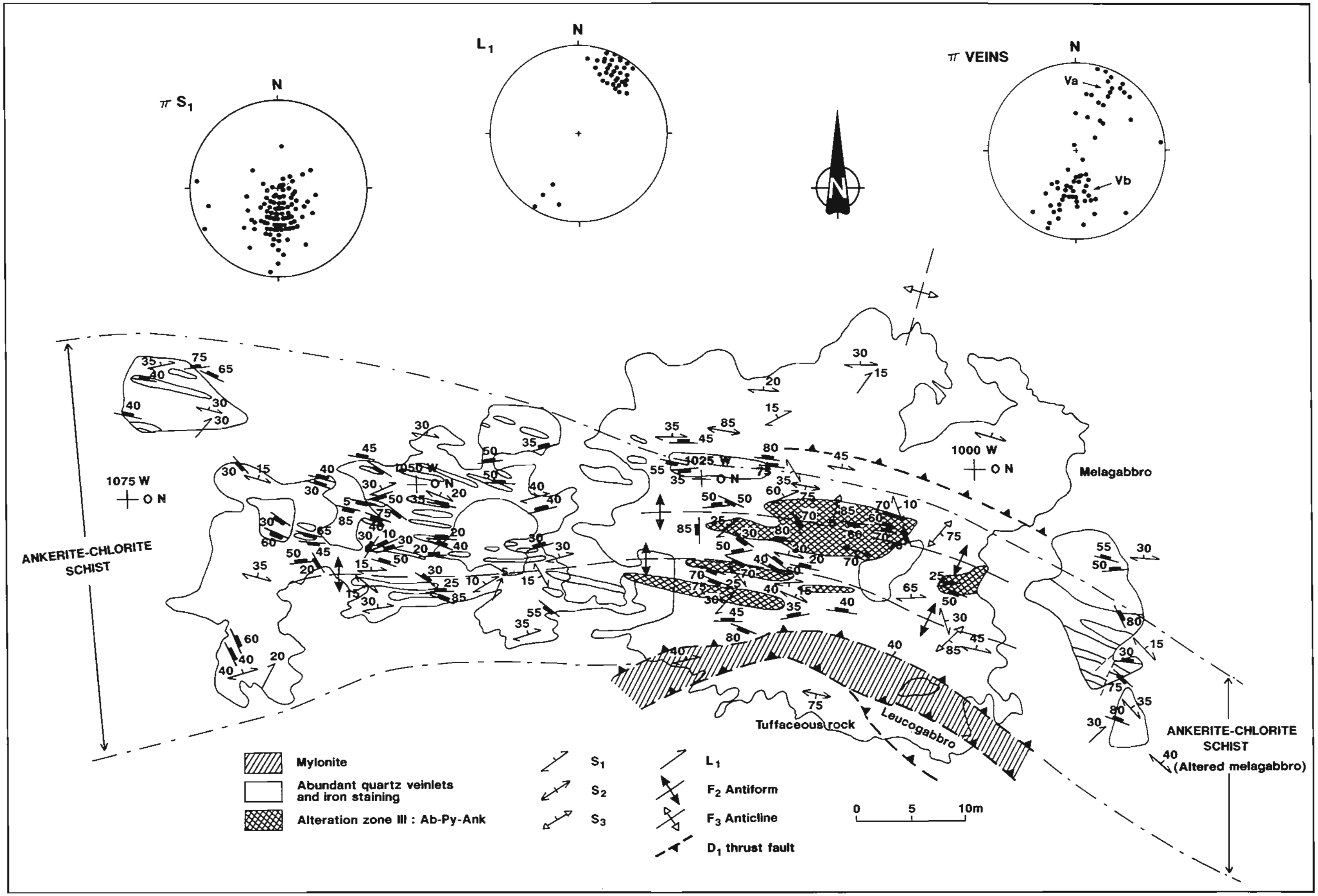

Figure 4. Detailed sketch of the Main Zone outcrop and compilation of the main structural elements on equal area projection (lower hemisphere). See Figure 2 for location.

III) albite-pyrite-ankerite-gold. Zone boundaries are relatively sharp and reflect the disappearance of one alteration mineral and/or the appearance of new alteration minerals. Visible hydrothermal alteration also occurs around the borders of mineralized veins which, in some cases, exhibits the same mineral zonation. The various megascopic alteration zones will be briefly described. A Masters thesis by J. Ramezani of Memorial University is under way and focuses on hydrothermal alteration and radiogenic isotopic characteristics of the deposit.

Alteration zone I: ankerite-chlorite schist

The first alteration zone is characterized by the abundance of ankerite as well as the prominent replacement of ferromagnesian minerals by chlorite. The gabbros are highly carbonatized and display a distinct rusty color. Rocks within this zone are intensely deformed and contain a very well-developed S_1 foliation and stretching lineation. This zone is usually found immediately adjacent to D_1 shear zones.

Alteration zone II: albite-chlorite

Alteration zone II is characterized by the appearance of white albite and the abundance of chlorite and is commonly adjacent to the mineralized zones. In the field, gabbros in this zone are usually slightly schistose and display a typical dark green coloration. Gabbros contain at least 5% white albite phenocrysts and less carbonate minerals than in alteration zone I. Secondary magnetite may be locally present.

Alteration zone III: albite-pyrite-ankerite-gold

The most significant characteristics of this zone are the formation of coarse grained, pink albite, and the appearance of pyrite and gold (Huard, 1990; Wilton and Evans, 1991). The most intensely altered gabbro can contain up to 60% albite, 20% ankerite, 10% sericite, and 10% pyrite, completely replacing the host rock. Albite occurs as coarse euhedral crystals up to 5 mm in diameter and pyrite is disseminated or forms coarse euhedral crystals up to 1 cm long. Pyrite overgrows leucoxene in varying degrees and gold occurs as microveinlets and disseminated blebs within the pyrite (Huard, 1990).

In the highly mineralized zones within the least deformed gabbro, alteration zones II and III are both present. Increasing alteration intensity is shown by the following events: (1) alteration of the host rock (alteration zone II) accompanied by crystallization of ankerite within open space fractures created by stretching and brittle fracturing of the plagioclase phenocrysts, (2) injection of quartz-carbonate veins in host rocks with localized alteration zones, and (3) complete replacement of host rock to produce advanced alteration zone III. The richest gold mineralization is always correlated with advanced alteration zones (zone III). Within alteration zone I, narrow lenses of less than 50 cm, recording alteration events 2 and 3, have been identified. These lenses of relatively lower strain consist of a more massive looking, fine grained gabbro occurring within strongly foliated, finer grained, laminated gabbro. They can be recognized in outcrop by their distinctive oxidized iron staining, and the injection of numerous quartz veinlets (Fig. 4).

Distribution of the mineralized zones

The gold mineralization occurs within four distinct zones. These are from south to north, the Main Zone, the Gabbro Zone, the Stog'er Tight Zone and the Magnetic Zone (Fig. 2). The Main Zone and the Gabbro Zone are found within the lower and middle gabbroic sills respectively, whereas the Stog'er Tight Zone and the Magnetic Zone are hosted by the upper gabbro. The altered zones are continuous and stratabound, and can be followed for several hundred metres. The most important zone, the Stog'er Tight Zone extends approximately 500 m and thins progressively towards the east-southeast. Channel samples grading up to 23 g/t Au over 7.0 metres were reported (Huard, 1990). In unpublished drill core sections, the four alteration zones can be easily identified. They are subparallel to S_1 and extend downdip to a maximun depth of 200 m. Highly altered and mineralized lenses are contained within the altered zones. However, due to lack of exposure in the third dimension, their exact geometry and distribution within the altered zones is unknown. Within the Stog'er Tight Zone, the mineralized zone can be followed along several hundreds of metres towards the east with no apparent plunge. On the Main Zone outcrop the trace of the mineralized zones are subparallel to S_2 (N100°-110°) and occur within F_2 antiformal structures (Fig. 4).

Description of vein types

Four dominant types of veins were recognized in the area. These will be referred as Va-, Vb-, Vc-, and Vd-types. Types Va and Vb represent the main mineralized veins, whereas Vc and Vd types are not associated with significant gold content. Va veins consist of quartz-ankerite+chlorite+pyrite. They strike 120° and dip 65° to 85° to the south. These are extensional veins perpendicular to the L_1 stretching lineation (Fig. 4). Vb veins are slightly discordant with S_1, strike east-west and dip between 30° and 60°. They consist of quartz and ankerite+pyrite. Type Vc quartz-carbonate veins are emplaced along the F_2 axial fracture cleavage. Type Vd quartz veins trending southeast to south-southeast are not associated with alteration and are generally barren.

The mineralized veins (Va and Vb) show no signs of stretching, boudinage, or isoclinal folding typical of D_1 ductile shearing although they are affected by asymmetric F_2 folds (Fig. 3d). Stereographic projections of Figures 4 and 5 show that the poles to mineralized veins are distributed along a great circle defined by the F_2 fold axis.

Structural control of the mineralized zones

The highly mineralized zones are located within the least deformed ferro-leucogabbro units of the sills. Deformation within these more competent rocks is characterized by foliation-parallel flattening and extension along the L_1

stretching lineation (Fig. 3b) creating fractures which cut across plagioclase crystals, forming porphyroclast aggregates that retain their primary textures. The highly foliated gabbro units, although strongly foliated, recrystallized, and pervasively carbonatized, show low gold content. Consequently, it is proposed that the gold was preferentially deposited within the highly fractured, coarser grained gabbroic units that contained initially higher iron content (ferro-leucogabbros) rather than strictly controlled by the proximity of D_1-related shear zones (ankerite-chlorite schists).

The timing of the gold mineralization can be restrained with respect to deformation by the following relationships:

1. Subvertical Va veins are perpendicular to the L_1 stretching lineation.
2. The mineralized veins (Va and Vb) show no signs of stretching, boudinage, or isoclinal folding typical of D_1 ductile shearing.
3. Most Va and Vb veins as well as the alteration zones are folded by F_2 folds.
4. D_1 shear zones show absolutely no signs of hydrothermal alteration.
5. Alteration zones with intense veining are spatially associated to D_1 ductile shear zones.
6. Carbonate minerals are aligned subparallel to L_1, have been partially dissolved along S_2 cleavage surfaces, and display postkinematic relationships with respect to the S_1 foliation.
7. D_2 shear zones cut across the altered and mineralized zones.

These relationships strongly suggest that the hydrothermal alteration and associated mineralization of the host gabbro is late-D_1 and pre-D_2. Fracturation, veining, and gold mineralization can be assigned to a late-stage ductile-brittle event within the D_1 phase.

An alternative interpretation of the deformation history is possible. Considering the coaxiality of the D_1 and D_2 structures, one could assign their development to a single and progressive deformational event, D_I. In such a case, D_1 structures would reflect an earlier more ductile part of the deformation (D_E) and D_2 structures would correspond to a late stage, more brittle phase (D_L). Accordingly, D_3 structures would be termed D_{II}. Such a model was proposed by Dirks and Wilson (1991) for the Arltunga Gold Field in Australia where mineralized quartz vein systems occur within a structural setting similar to the one observed in the Stog'er Tight area. Although large-scale north vergent kink zones and steeply south-dipping shear zones overprint the main low-angle thrust, they are suggested to have formed as a

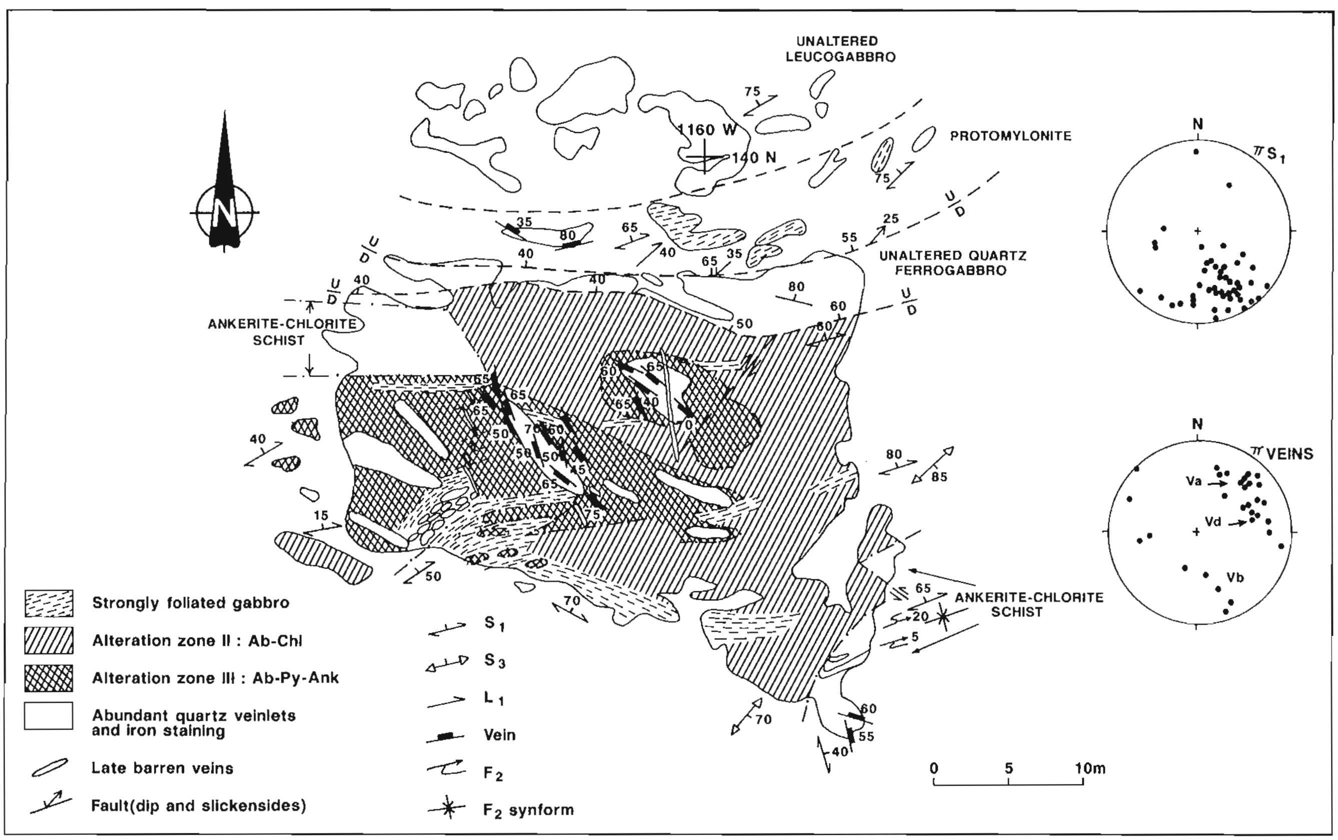

Figure 5. Detailed sketch of the Gabbro Zone outcrop and compilation of the main structural elements on equal area projection (lower hemisphere). See Figure 2 for location.

conjugate set in a north-south-directed compressional regime (Dirks and Wilson, 1991). Veins formed early during the development of kink folds as dilational openings and were consequently folded, partly fractured, sheared, and mineralized during ongoing deformation and localization of strain within steep reverse faults. Although we favour a structural evolution divided into three distinct deformational events (D_1, D_2, and D_3), a model involving progressive deformation for the first two events (D_1 and D_2) is not in disagreement with the observed relationships described above and should be considered for the Stog'er Tight deposit.

DISCUSSION AND CONCLUSION

Mineralized zones within the Stog'er Tight gold deposit were controlled by D_1 ductile shear zones that affected differentiated gabbro sills. The preferential development of the shear zones along lithologic contacts is related to competency contrast and layer anisotropy. The alteration, veining, and gold mineralization postdate ductile shearing and are localized within the hanging wall or, more rarely, within the footwall of D_1 shear zones. Another important control on gold deposition was the iron-rich ferro-leucogabbro units, which display much higher gold levels than adjacent intensely veined units. Both the layer anisotropy and the lithological controls favor the stratabound form of the orebodies. As suggested in other mesothermal altered wallrock gold deposits (Dubé, 1990b), the sulphur contained in the hydrothermal fluid reacted with the primary iron contained within the ferro-leucogabbro during pyritization, thus destabilizing the gold bisulfide complex and allowing precipitation of gold.

The D_1 ductile shear zones are related to major faults of the Baie Verte Peninsula area. Major and minor shear zones and faults of the Point Rousse area, which are ascribed to the same major Acadian deformational event can be divided into distinct categories. As previously proposed by Robert (1990) for shear zones of the Val d'Or area in the Abitibi Subprovince, three different orders of shear zones are defined on the basis of scale and regional importance. The Point Rousse Complex is tectonically bounded to the south by the Scrape Thrust. This major southeasterly-trending, north-dipping thrust fault juxtaposes rocks of the Point Rousse Complex over rocks of the Pacquet Harbour Group. The Scrape Thrust is a major interblock fault and marks the contact between Hibbard's (1983) Paratectonic and Eastern Orthotectonic blocks. Other faults within the Point Rousse Complex (Fig. 1, cross-section) which dip steeply to the northwest, and were interpreted as high-angle thrust faults by Hibbard (1983) are coeval with the formation of the regional foliation (S_1) and are interpreted as secondary structures related to the emplacement of the Point Rousse Complex over the Pacquet Harbour Group (Hibbard, 1983). In the study area, D_1 shear zones which are subparallel to major faults affecting the Point Rousse Complex, are interpreted as third order structures related to the more important, first order Scrape Thrust. All three orders of shear zones show variable intensity of hydrothermal alteration. Although no significant mineralization has been observed within the Scrape Thrust, major gold occurences have been identified along second and third order shear zones in the Point Rousse area (Deer Cove and Pine Cove).

The Stog'er Tight gold deposit is a Paleozoic example of a subclass of mesothermal gold deposits, those hosted in layered gabbroic sills. This type of gold deposit is common worldwide and some of the best examples are the Archean gold deposits enclosed in the Golden Mile dolerite sill at Kalgoorlie in Australia (Phillips, 1986), gold deposits enclosed within the Bourbeau sill in Abitibi (Dubé et al., 1987; Dubé et al., 1990; Dubé, 1990b) and the San Antonio deposit in Manitoba (Poulsen et al., 1986; Lau, 1988; Ames et al., 1991). The analogy, both on the alteration type and geometrical point of view is particularly strong with the San Antonio deposit.

The genesis of this type of mesothermal gold deposit is clearly controlled by the strong mechanical and chemical influence of the hosting gabbroic sill on the development of the structural permeability which allowed the circulation of the hydrothermal fluid, the development of hydrothermal alteration, and the destabilization of the gold-bearing bisulphide complex. The structural history and timing of the mineralization of the Stog'er Tight deposit thus illustrate important relationships between major fault zones and gold mineralization.

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Geological Survey of Canada Project 890026

The Middle Ordovician (Caradocian) Deschambault Formation, St. Lawrence Lowlands, southern Quebec: a shallow water carbonate ramp on a drowning platform

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Abstract

The Deschambault Formation succession in the St. Lawrence Lowlands belongs to a drowning carbonate platform sequence laid down during the Middle Ordovician in southern Quebec. The formation is characterized by a number of carbonate lithofacies typical of shallow water depositional environments ranging from nearshore to more or less deep subtidal settings. Regional lithofacies distribution locates the deepest setting in the central segment of the studied area. Location of the Middle Ordovician shoreline and/or synsedimentary extensional tectonism are proposed as possible controls for the variation in regional lithofacies distribution. The lithofacies architecture is best explained in term of deposition on a gently sloping carbonate ramp, however, a flat-topped, shoal-rimmed carbonate platform model is not ruled out. Evidence of lowstand conditions are lacking in the Deschambault Formation, the succession is interpreted as resulting from highstand conditions in a transgressive episode.

Résumé

La Formation de Deschambault des basses-terres du Saint-Laurent appartient à la série sédimentaire résultant de l'effondrement de la plate-forme ordovicienne de roches carbonatées du sud du Québec. La formation est caractérisée par des lithofaciès carbonatés de milieux de sédimentation d'eau peu profonde dont la nature variait de littorale à plus ou moins profonde. La distribution régionale des lithofaciès localise le milieu de sédimentation le plus profond dans le secteur central de la région étudiée. La position de la ligne de rivage à l'Ordovicien moyen ou une tectonique active d'effondrement, ou les deux, sont proposées comme mécanismes possibles, responsables de la variation dans la répartition régionale des lithofaciès. L'architecture des lithofaciès est typique d'un modèle sédimentaire de rampe carbonatée faiblement inclinée, cependant un modèle de plate-forme horizontale, à marge sablonneuse, est également considéré. Les conditions marines de bas niveau sont absentes de la formation de Deschambault et la succession est interprétée comme étant le résultat d'un haut niveau marin dans un intervalle transgressif.

INTRODUCTION

The almost flat-lying 1500 to 3000 m Cambrian - Ordovician succession of the St. Lawrence Lowlands of southern Quebec records platformal sedimentation in a complete Wilson cycle, from opening of the Iapetus ocean in Late Precambrian time up to its partial closure in Ordovician time (Globensky, 1987). Following this event (Taconian orogeny), marine to subaerial siliciclastic sedimentation resumed in a successor basin with, however, almost no record of younger than latest Ordovician sedimentation.

Early mapping and study of the entire area resulted in a lithostratigraphic framework (Globensky, 1987) relying on the numerous paleontologic and resulting biostratigraphic works. However, the sedimentology and diagenesis of most carbonate and siliciclastic units is poorly constrained, apart some works either on specific units (Hofmann, 1963; Mehrtens, 1979; Belt and Buissière, 1981; Bernstein, in prep.) or oriented toward specific themes (B.E.I.C.I.P., 1975; Tassé et al., 1987). Local and regional facies distribution, sequence studies, carbonate diagenesis, and interplays between facies patterns and active tectonism still need to be fully documented and integretated before paleogeographic and paleotectonic evolutions of the Quebec's segment of the western margin of Iapetus can be completely understood.

In that long-term perspective, initial field work focused on two units (the Deschambault Formation and the Philipsburg Group) that could best help to understand the relations between the carbonate platform evolution and the nearby ongoing Taconian orogeny.

The Deschambault Formation belongs to the Middle - lowermost Upper Ordovician Trenton Group. The Trenton succession is interpreted to represent progressively deeper conditions on a drowned carbonate platform following the initiation of Taconian loading (Globensky, 1987). The Trenton carbonate platform of southern Quebec evolved in the Quebec Reentrant at a time when Taconian allochtnons

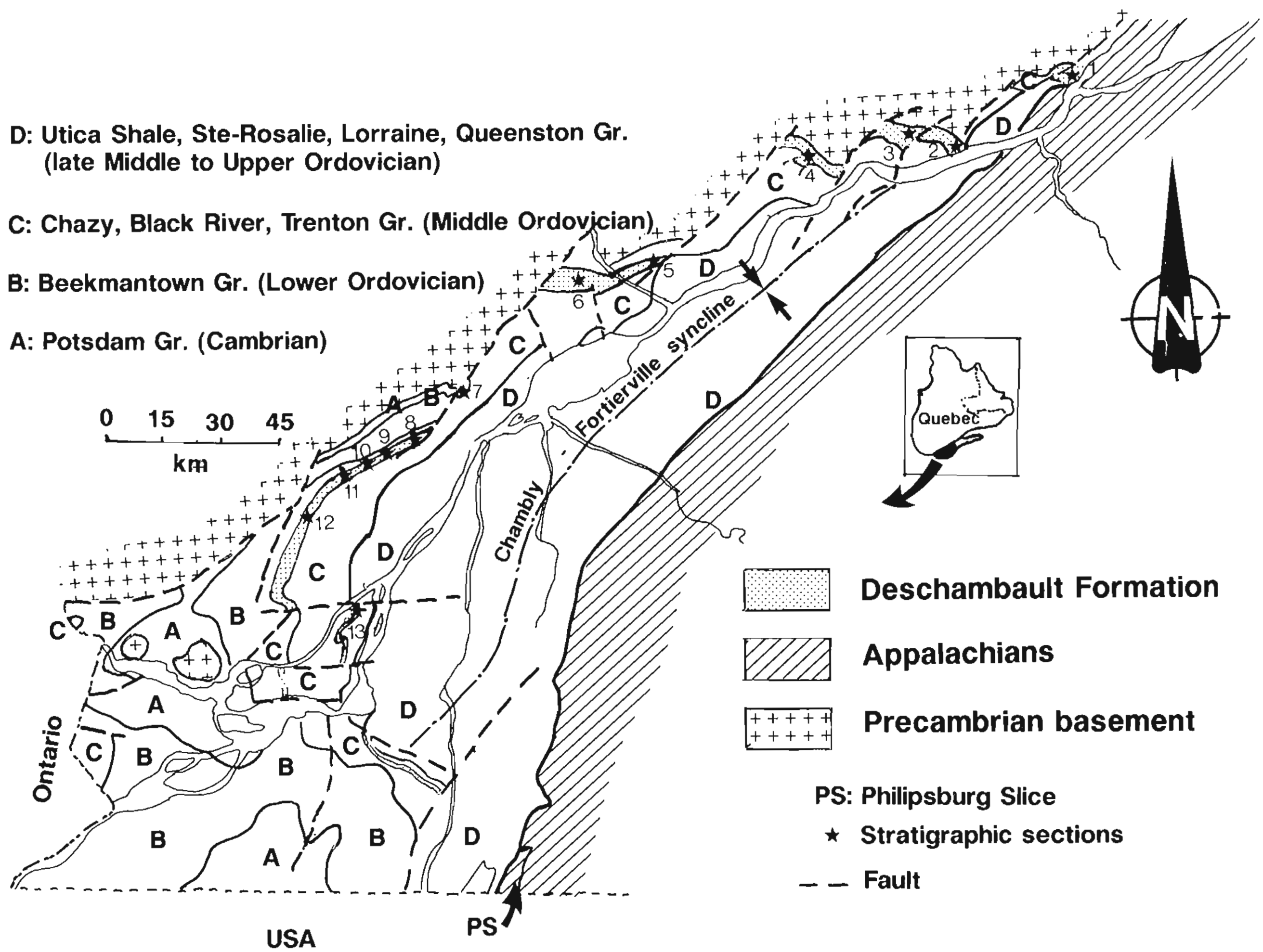

Figure 1. Simplified geological map of the St. Lawrence Lowlands with locations of the described sections of the Deschambault Formation. Sections 1: rivière Montmorency, 2: Neuville quarry, 3: rivière Jacques Cartier, 4: rivière Sainte-Anne, 5: rivière au Lard, 6: rivière Yamachiche, 7: rivière du Chicot, 8: Composite from Saint-Thomas quarries and rivière Chaloupe, 9: rivière de l'Assomption, 10: rivière Ouareau, 11: rivière Saint-Esprit, 12: rivière de l'Achigan, 13: Francon quarry (from Mehrtens, 1979).

were already emplaced at the St. Lawrence Promontory in western Newfoundland (Knight and James, 1987; Stenzel et al., 1990). This north-east to south-west diachronism in the Taconian collision is best explained by the geographic location of Quebec and Newfoundland platforms with respect to inherited Precambrian reentrants and promontories (Bradley, 1989). The Deschambault Formation was initially studied (two months of field work), mainly because of the diversity of its lithofacies when compared to other Trenton units (Globensky, 1987). The field work resulted in recognition of regional as well as local lithofacies distribution in the Deschambault Formation. The Lower Ordovician Philipsburg Group of southern Quebec Appalachians was the subject of few days of field work. On the eastern side of the North America craton, the Early Ordovician was the time of transition from passive to active continental margin. The first evidence of convergence of plates is given by the migration of a peripheral bulge on the platform (Hiscott et al., 1983; Quinlan and Beaumont, 1984) in the northeast (Jacobi, 1981; Knight et al., 1991) and southwest (Rowly and Kidd, 1981; Shanmugan and Lash, 1982) segments of the craton margin. This event resulted in large-scale erosions and unconformities. However, evidence of these regional, widespread unconformities is virtually absent in the Early Ordovician Beekmantown Group of the St. Lawrence Lowlands (Bernstein, 1991). Reconnaissance survey aimed at the coeval Philipsburg Group in the Appalachian external domain (Slivitzki and St-Julien, 1987) of southernmost Quebec. Major unconformities, various karstic and subarial surfaces were observed in the Philipsburg Group. It suggests, as reported very recently (after our field season) by Knight et al. (1991), that evidence for the migration of the peripheral bulge are present in southern Quebec. This topic will be subject of future contributions.

THE DESCHAMBAULT FORMATION

The Deschambault Formation belongs to the lower part of the Middle - lowermost Upper Ordovician Trenton Group. The Deschambault Formation is dated as Caradocian (latest

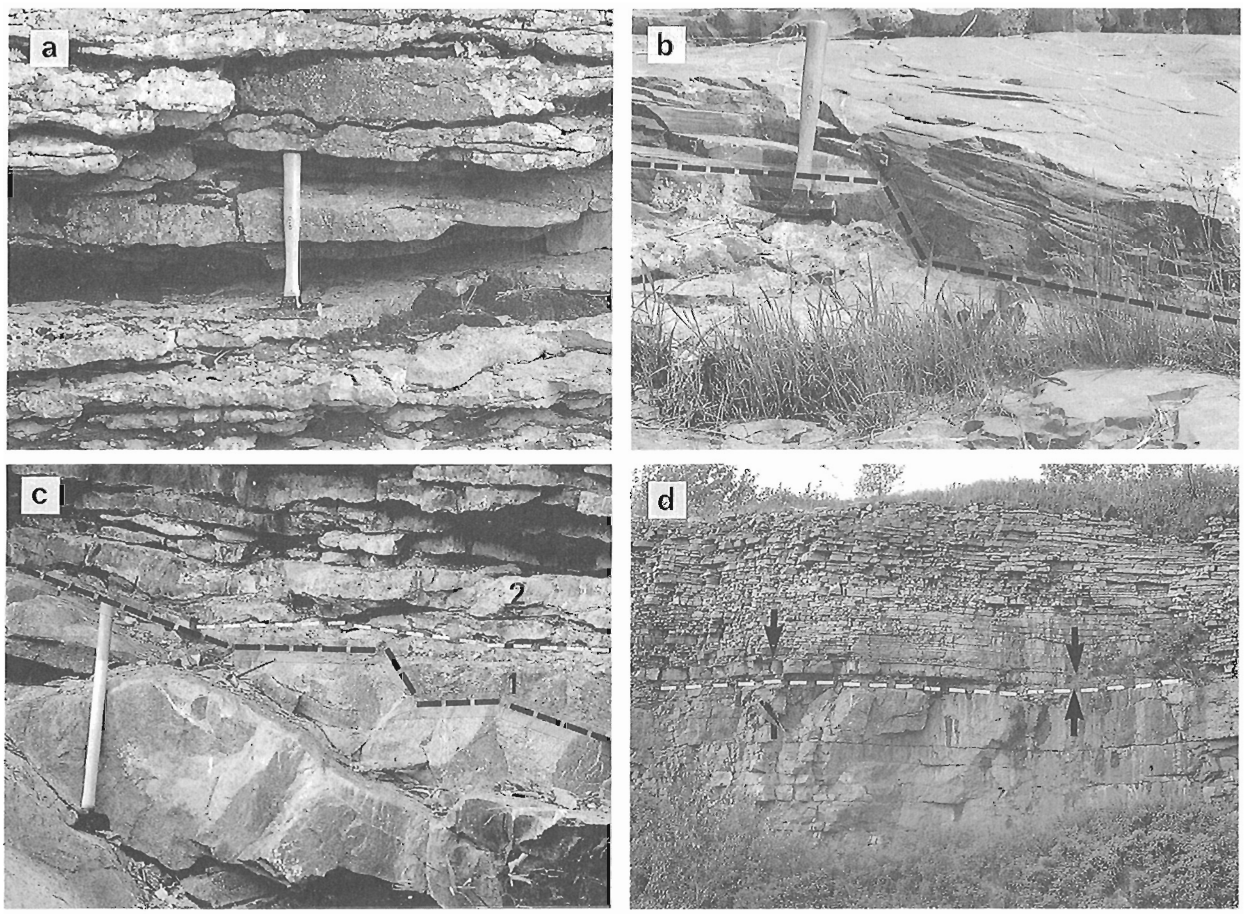

Figure 2. a) Conformable contact between the Deschambault and Ste-Anne formations. Hammer head stands on the uppermost bed of the Ste-Anne Formation. Section 4. b) Unconformable contact (outlined in black) between the Deschambault and underlying Ouareau formations. Section 10. c) Unconformable contact (outline in black) between the Deschambault and Precambrian basement. (1) *Solenopora* packstone and (2) wavy-bedded wackestone calcilutite. Section 1. d) Contact (outline in white) between the Deschambault and overlying Neuville formations. Note the progressive thinning (shown by black arrows) of Neuville beds as they onlap the top of the Deschambault. Section 1.

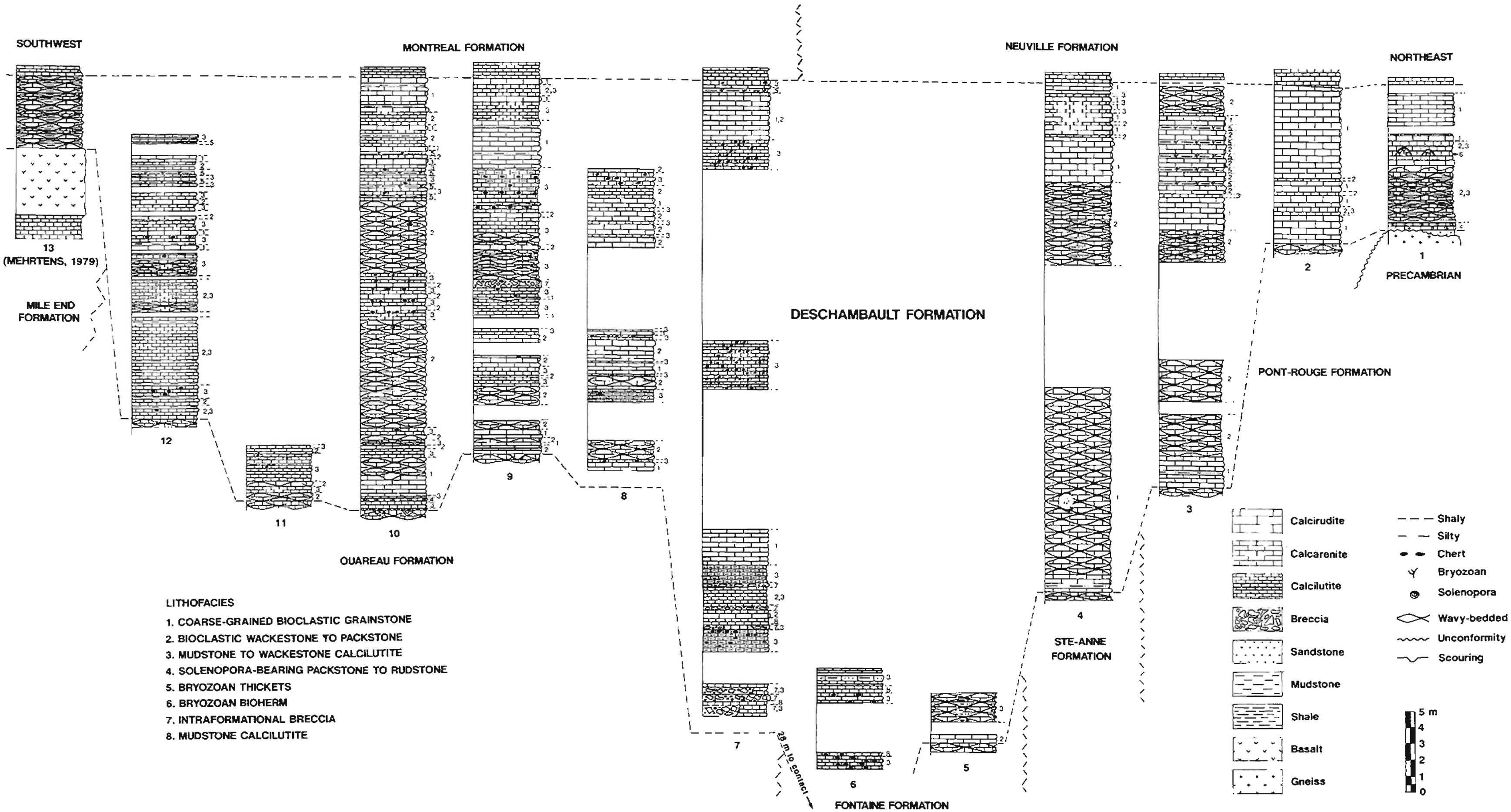

Figure 3. Lithostratigraphic correlations for the Deschambault Formation. The contact between the Deschambault and Neuville / Montréal formations is used as datum line. Location of sections on Figure 1.

Middle Ordovician) from brachiopod, bryozoan, and conodont faunas (B.E.I.C.I.P., 1975; Mehrtens, 1979; Globensky, 1987).

The Deschambault Formation is, from its coarser-grained nature, a distinctive unit in the dominantly finer-grained sediments of Trenton Group. The Deschambault did not attract much sedimentologic attention besides Mehrtens' (1979) thesis dealing with the entire Middle Ordovician succession. However, the litho- and biostratigraphy as well as the paleontology of the Deschambault is fully covered by a number of studies and thesis (Globensky and Jauffred, 1971; Clark et al., 1972; Pickerill and Forbes, 1979; Riva et al., 1977; Globensky, 1987).

The Deschambault Formation outcrops on the northwest limb of the Chambly-Fortierville syncline from Quebec City down to the Montréal area (Fig. 1). The thrust-emplaced taconian nappes at the eastern margin of the St. Lawrence Lowlands prevent the Deschambault Formation from outcropping on the southeast side of the syncline (Fig. 1). The Deschambault Formation overlies either conformably or with a slight discordance (Fig. 2a and 2b), many lower Middle Ordovician units (Mile End, Ouareau, Fontaine, Saint-Alban, Sainte-Anne, Pont-Rouge formations). In the northeasternmost sector, the Deschambault unconformably overlies the Precambrian basement (Fig. 2c). The Deschambault Formation is overlain by the upper Middle Ordovician Montréal and Neuville formations in the southwestern and northeastern segments of the St. Lawrence Lowlands, respectively. The best exposed contact shows an onlap relationship (Fig. 2d).

Interestingly, thickness of the formation varies from about 10 m in the vicinity of Quebec City, thickening up to close to 70 m in the Trois-Rivières area and progressively thinning to a minimum of 4 m in the Montréal area (Fig. 3). This variation is also coupled with some distinct lithofacies variations.

Lithofacies and paleoenvironmental interpretation

The Deschambault Formation is dominated by carbonates; siliciclastics are restricted to some shaly interbeds and clay seams in the limestones. We recognized eight lithofacies, the distribution of which are shown on Figure 3.

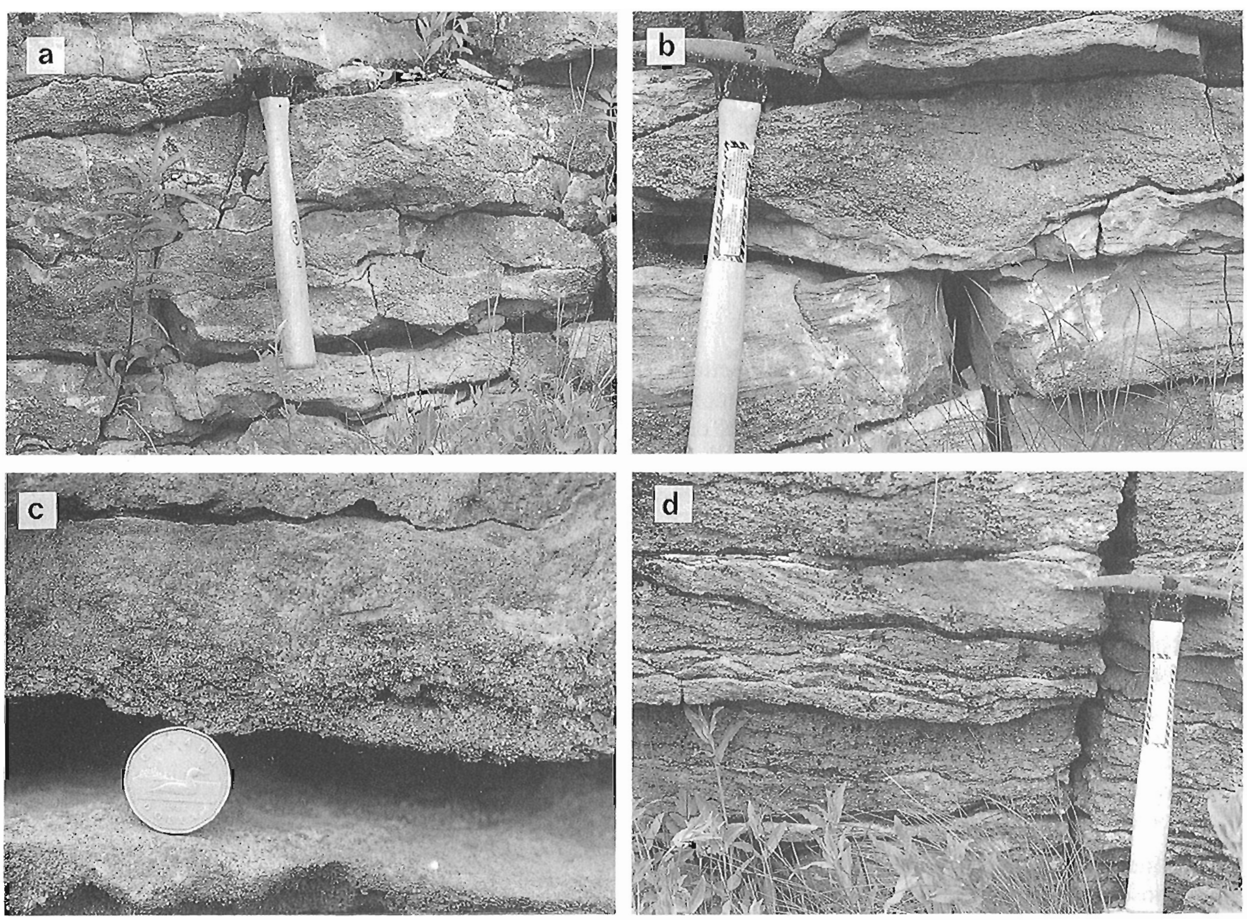

Figure 4. Lithofacies 1. a) Amalgamated, slightly wavy grainstone beds. Section 2. b) Grainstone bed of lithofacies 1 scouring into parallel laminated packstone of lithofacies 2. Section 2. c) Graded grainstone bed. Section 1. d) Abundant planar to through crosslaminations in grainstones. Section 3.

Lithofacies 1: coarse-grained, bioclastic grainstone

This is the most typical lithofacies of the Deschambault Formation. It is best developed over the northeastern and part of the southwestern segments of the studied area (Fig. 3).

The lithofacies is made up of medium- to thickly-bedded (8-60 cm, average 15 cm) grainstones. Beds commonly amalgamate (Fig. 4a) and define crudely-bedded successions. The lithofacies shows abundant sedimentary structures such as scouring (Fig. 4b), grading (Fig. 4c), various thick parallel and crosslaminations (Fig. 4d), with either high or low angle planar and trough types. The grainstones are composed of various bioclastic and less common intraclastic material in a grain-supported framework cemented by sparry calcite. Bioclasts are dominated by pelmatozoans debris and fragments of bryozoans. Brachiopods, pelecypods, gastropods, and trilobites were also found. Size of allochems range from fine sand- to pebble-sized with coarse sand-sized material being most abundant. The fauna is reworked and strongly abraded.

Interpretation

Based on the amalgamate nature of bedding, the abundance of sedimentary structures indicative of a high-energy depositional environment, the absence of any visible mud or clasts finer than sand-sized particles, and the reworked nature of bioclastic material, this lithofacies is interpreted to represent sand shoals and sheets accumulated in an agitated shallow water setting. However, sand shoals can be found either at the margin of a rimmed shelf or in the more inner part of a carbonate ramp (Halley et al., 1983; Wilson, 1986). The geometry of the carbonate platform will be discussed below.

Lithofacies 2: bioclastic wackstone to packstone

This lithofacies is observed in every measured sections (Fig. 3). The lithofacies occurs in planar to more commonly wavy beds (Fig. 5a) ranging from 6 to 25 cm (average 15 cm) in thickness. These beds are commonly separated by millimetres to centimetre-sized clayey mudstone interbeds. The rock offers few to locally abundant sedimentary structures such as parallel, low-angle crosslaminations and wave ripples (Fig. 5b). These are locally bioturbated.

Bioclastic material is locally abundant enough to give way to packstone deposit. The bioclasts are dominated by various bryozoans of both the massive colonial and delicate ramose forms. Brachiopods, gastropods, pelecypods, trilobites, pelmatozoans, and corals are also observed. They are commonly concentrated in centimetre-sized lenses and pockets. However, reworking was likely local and not intense since it did not abraded much of bioclasts.

The shaly interbeds are typified by concentration of well preserved ramose bryozoans (Fig. 5c). Locally, the amount of bryozoans becomes important enough that a peculiar lithofacies is assigned to that deposit (see lithofacies 5 below).

Interpretation

The open marine fauna of the lithofacies, the local reworking of the fauna, and sedimentary structures observed suggest a shallow marine subtidal deposit for the lithofacies (Wilson, 1986). The setting was either slightly deeper than lithofacies

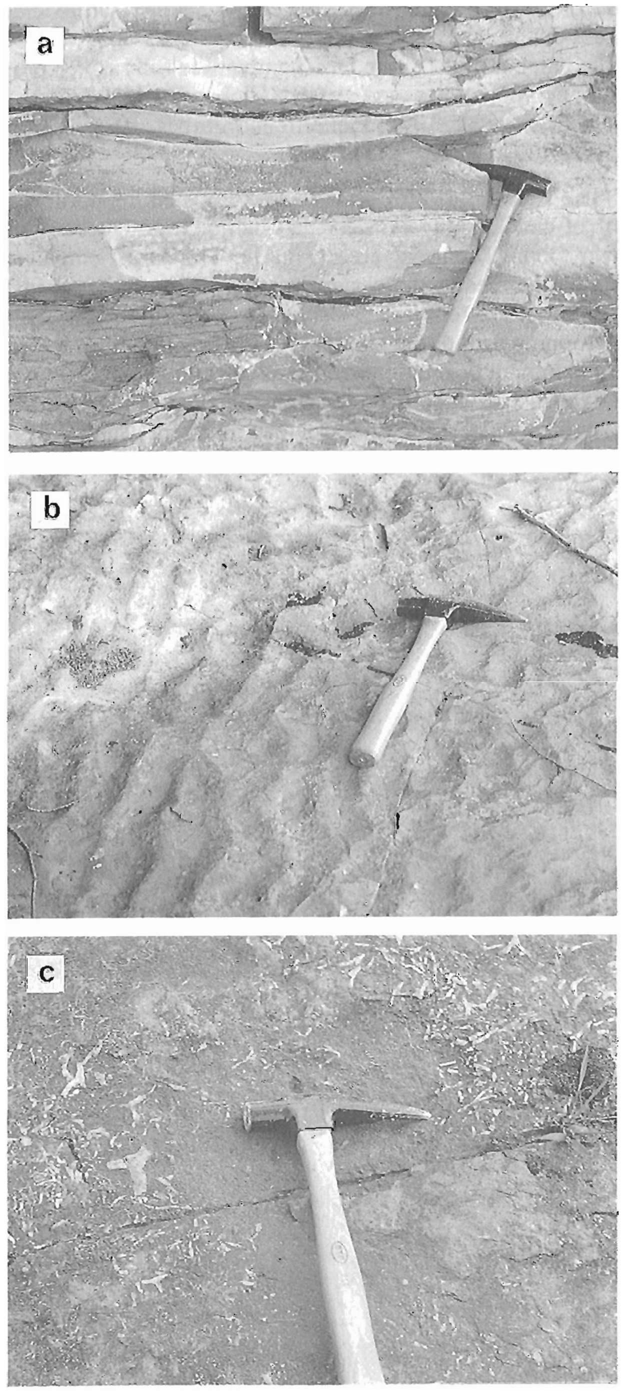

Figure 5. Lithofacies 2. a) Slightly wavy beds of wackestone to packstone. Section 10. b) Small scale wave ripples developed on top of wackestone calcilutite. Section 10. c) Plane-bedding view of shaly mudstone interbed with abundant ramose bryozoans. Section 3.

1 or protected from high energy marine swells, likely by the previously described sand shoals. As for the shaly interbeds, a brief but sudden shutdown of the carbonate factory (James, 1979) has to be invoked (see discussion under lithofacies 5).

Lithofacies 3: mudstone to wackstone calcilutite

The lithofacies is widely distributed in measured sections (Fig. 3). It is represented by planar and wavy beds ranging from 6 to 20 cm thick (average 12 cm). Limestones are separated by millimetre to centimetre-sized shaly mudstone interbeds. The lithofacies is almost devoid of sedimentary structures aside from some low-angle crosslaminations of hummocky type (Fig. 6a). In these structures, lensoid concentrations of bioclastic material are noted, whereas elsewhere bioclasts are scattered throughout the rock. Bioclasts are represented by a diversified fauna with bryozoans, brachiopods, and gastropods, among others. Some well preserved, pelmatozoan stems are observed (Fig. 6b) as well as, in the northeastern sector, a biostratigraphic distinctive bed with oriented orthocones (Fig. 6c). Bioturbation is common and is represented by *Chondrites*-like ichnofacies (Fig. 6d).

In the central and southwestern segments of the Deschambault outcropping sector, chert is observed in that lithofacies. The chert occurs as commonly rounded and sometimes coalescing nodules (Fig. 6e). These can make up to 80% of individual beds.

Interpretation

The lithofacies is interpreted as a slightly deeper subtidal deposit, below fairweather wave base but above storm wave base. Such an interpretation is based on the nature of sedimentary structures indicative of occasional bottom sediment remobilisation, and on the diversified and well preserved open marine fauna (Wilson and Jordan, 1983). The origin of the chert still has to be documented. Some uncommon field outcrops show it as being related (or enhanced?) by late, small fracturing events. However, most of the chertified occurences do not show any obvious structural control.

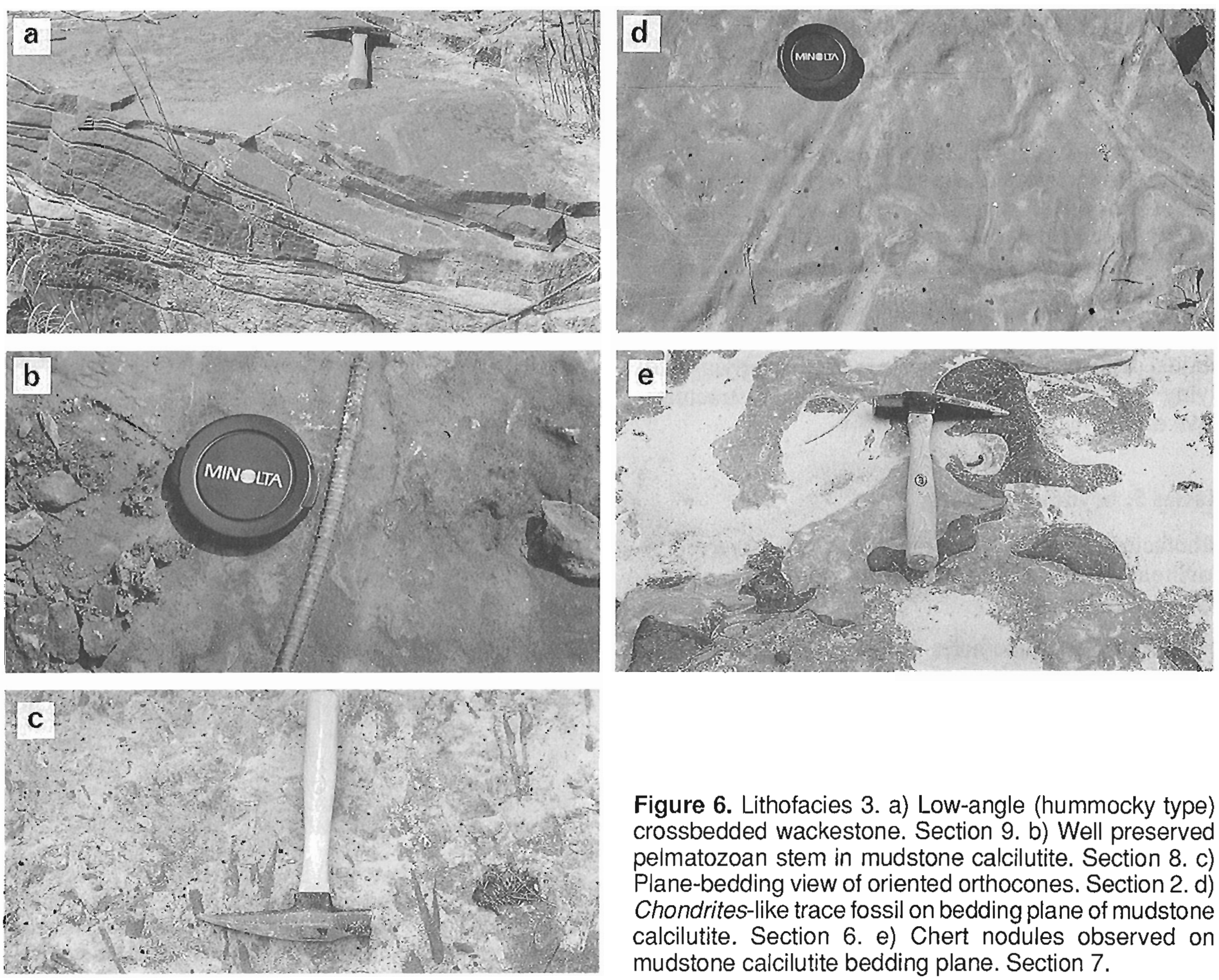

Figure 6. Lithofacies 3. a) Low-angle (hummocky type) crossbedded wackestone. Section 9. b) Well preserved pelmatozoan stem in mudstone calcilutite. Section 8. c) Plane-bedding view of oriented orthocones. Section 2. d) *Chondrites*-like trace fossil on bedding plane of mudstone calcilutite. Section 6. e) Chert nodules observed on mudstone calcilutite bedding plane. Section 7.

Lithofacies 4: Solenopora-bearing packstone to rudstone

This lithofacies was only observed at two localities, the rivière Montmorency and rivière Ouareau sections (Fig. 3, sections 1 and 10).

At the rivière Montmorency, the lithofacies occurs as a 50 cm-thick bed conformably overlying a 50 cm undated basal sandstone unit (Fig. 7a). The lithofacies is also transgressive over the Precambrian crystalline basement (Fig. 2c) even occurring into irregularly-shaped erosive structures developed in grenvillian gneiss (Fig. 7b). The rock is typified by a high concentration of red algae (*Solenopora*) resulting in a grain-supported packstone to rudstone bed. The red algae are well rounded, unsorted and commonly broken suggesting some reworking. Matrix is made up of bioclastic carbonate mud with brachiopod and pelmatozoan remains. The bed displays poorly defined thick parallel laminations, and is gently sloping 5° to the southeast, contrasting with the horizontally bedded overlying wackestones (Fig. 7a).

The *Solenopora*-bearing packstone at the rivière Ouareau occurs as a 30 cm-thick bed in the lower part of the succession, conformably overlying and overlain by subtidal wackestone deposit. The red algae are abundant, well rounded with no indication of wave reworking.

Interpretation

The *Solenopora*-bearing lithofacies at the rivière Montmorency is interpreted as a sand/gravel shoal, possibly a foreshore deposit as suggested by its relationship with Precambrian basement, the reworked nature of the bioclastic material, and seaward gently dipping bed (Inden and Moore, 1983).

The same lithofacies at the rivière Ouareau, is interpreted as a subtidal deposit, based on the nature of the overlying and underlying lithofacies and lack of sedimentary structures indicative of high-energy deposit.

Lithofacies 5: bryozoan thickets

This lithofacies, although rather rare (Fig. 3), is observed in the northeastern and southwestern segments of the outcropping sector.

As described for lithofacies 2 and 3, shaly interbeds occurring between wackestone to packstone bioclastic limestones are locally rich in bryozoans. Some of these interbeds can reach up to 15 cm in thickness and are almost entirely composed of ramose bryozoans of *Batostoma* and *Hallopora* genera (Riva et al., 1977). The amount of bryozoans can be as high as 90% and averages 65% (Fig. 8a), forming in situ growth thickets. The matrix between bryozoans is made up of calcareous (diagenetic?) shaly mudstone with few other bioclastic material besides some small brachiopods. No evidence of reworking of the bioclasts was observed and the lithofacies is devoid of sedimentary structures.

Interpretation

The bryozoan thickets likely represent in situ accumulation, below fairweather and possibly storm wave bases, in an episode of undiluted siliciclastic mud sedimentation. The siliciclastic event is possibly related to a rapid but small-scale, transgressive event that disrupted the fragile carbonate production zone (James, 1979). The "polluted" water restricted biological activity to less-ecologically altered genera (bryozoan). The origin of the sudden transgressive events could possibly be related to high order Milankovitch cycles. However, more work is needed before suggesting the presence of such cyclicity. The bryozoan thickets lithofacies represents short-lived siliciclastic events in a relatively calm subtidal environment. It ended when the carbonate production zone was able to re-establish itself and keep up with rising sea level.

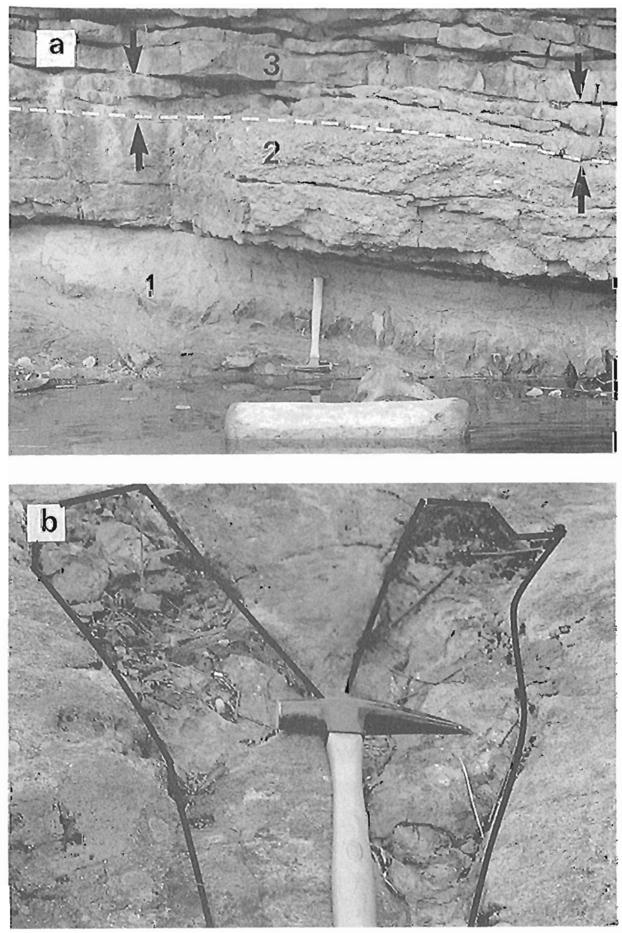

Figure 7. Lithofacies 4. a) Basal sandstone (1) conformably overlain by *Solenopora* packstone to rudstone limestone (2). Both are slightly dipping to the right of the photograph. Overlying the white dashed contact are horizontal and wavy-bedded wackestone calcilutites that progressively thin (arrows) towards the left. Section 1. b) Erosive features (outline in black) shaped into precambrian gneiss and filled by *Solenopora*-rich packstone. Section 1.

Lithofacies 6: bryozoan bioherms

This lithofacies is only known at the Rivière Montmorency (Fig. 3, section 1), where it occurs as two distinct bioherms overlying and surrounded by lithofacies 2 and 3 limestones (Fig. 8b). However, few thin beds of coarse-grained, bioclastic grainstones of the lithofacies 1 are also noted.

The bioherms occur as 1.4 m x 40 cm and 5.8 m x 1.2 m massive units. Exact geometry of these structures is unknown due to lack of third dimension view. The bioherms are composed of a ramose bryozoan bafflestone framework making up to 60% of the rock. No encrusting organism, either skeletal or non-skeletal, was observed in field exposures. The delicate framework is infiltrated by lime mud with accessory brachiopods, pelecypods, and pelmatozoans. Surrounding material on all visible flanks of the bioherms are devoid of internal sedimentary structures. These beds gently draped the bioherms suggesting small scale synoptic relief on the sea floor. No evidence of asymmetry or mechanical breakdown of the bioherms is noted.

Interpretation

The bryozoan bioherms are interpreted to have developed as small patch reefs (James, 1983). The framework were erected in relatively quiet water, possibly below fairweather wave base as indicated by: 1) the lack of sedimentary structures in the surrounding sediment as well as on the fragile framework that acted as baffle for suspended carbonate mud and accumulation of fines, 2) the lack of any visible early marine cementation and, 3) the absence of mechanical breakdown.

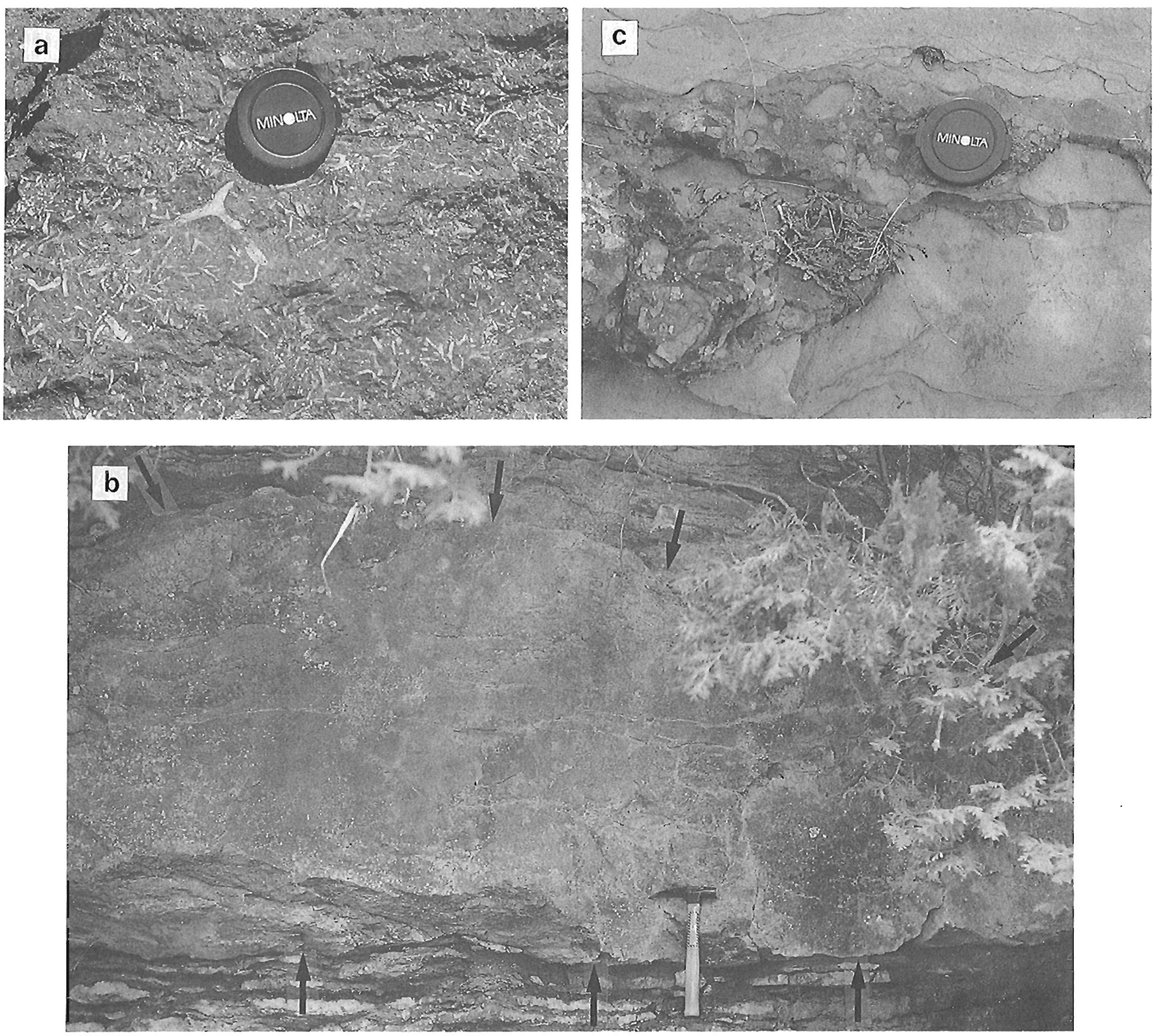

Figure 8. a) Dense network of in situ ramose bryozoans in calcareous shaly mudstone forming a growth thicket. Lithofacies 5, section 10. b) 1.4 m high bryozoan bioherm (limit outlined by arrows) surrounded by wavy wackestone calcilutite. Lithofacies 6, section 1. c) Intraformational breccia resulting from the in situ dismembering of the adjacent mudstone calcilutite. Lithofacies 7, section 7.

However, the restricted occurrence of these bioherms to one locality is an enigma since they occur above, and are surrounded by widely distributed open marine limestones.

Lithofacies 7: intraformational breccia

This lithofacies distribution is restricted to the central segment of the studied area (Fig. 3). It is closely associated with lithofacies 3 limestones, being either sandwiched between beds of the latter or laterally passing into it (Fig. 8c). The breccias generally occur as 5 to 30 cm (average 15 cm) thick units. The clast-supported and structureless breccias are made up of 50% to 90% clasts. The fragments are commonly angular and unsorted. Clast size varies from 1 mm to 15 cm (average 6 cm). They are composed of mudstone to wackestone calcilutites similar to the surrounding beds. The matrix is made up of fine crushed particles derived from the clasts with a variable amount of infiltrated siliciclastic clays. No in situ bioclastic material is present in the matrix.

Interpretation

The close spatial relationship of the breccias with lithofacies 3 beds as well as the nature of the clasts suggest an in situ formation for lithofacies 7. It would best be called "dismembered lithofacies". Its formation likely followed early seafloor lithification of the limestones, but the mechanism responsible for its generation is currently unknown. Instability of slightly sloping sediments could be envisaged.

Lithofacies 8: mudstone calcilutite

This lithofacies is also known as "sub-lithographic" limestones by other workers. Its occurrence is restricted to the central and southwestern segments of the studied area (Fig. 3).

The mudstone calcilutite occurs as thin (5 to 10 cm), even-bedded unit that are devoid of any visible bioclastic material. The lithofacies is structureless and generally sandwiched between lithofacies 2 or 3 limestones with sharp contacts to locally scoured tops.

Interpretation

From its field characteristics, the mudstone calcilutite is preliminary interpreted as possible peri-platform ooze (Cook and Mullins, 1983). It likely represent the deepest water sediment observed in the Deschambault succession.

DISCUSSION

The array of carbonate lithofacies can best be interpreted in term of depositional setting on a shallow water, gently sloping, carbonate ramp (Fig. 9a). However, conceptually, a flat-topped, rimmed carbonate platform model can not be ruled out (Fig. 9b). The latter model is not favoured since typical slope lithofacies are missing. Moreover, the common interfingering between lithofacies 1 (sand shoals/sheets) and lithofacies 2, 3, and 6 (bioclastic muddy sands, bioclastic muds, and patch reefs) argues for the ramp model.

Based on sedimentological, faunal and floral criterias, Mehrtens (1979) proposed two depositional environments for the limestones of the Deschambault Formation. These are: 1) an agitated shallow water setting for the carbonate sands and, 2) a quiet shallow water environment for the muddy sediments. Her reconstruction (Mehrtens, 1979, Fig. 22) shows a somewhat gently-dipping shelf, with a maximum depth (based on bryozoan assemblages) of less than 50 m. The carbonate sand lithofacies is located seaward as compared to the carbonate mud lithofacies. The author only partly agrees with this reconstruction since some of the muddy lithofacies (No. 3) displays sedimentary structures typical of storm remobilization without indication of fairweather wave-induced structures. Moreover, our own lithofacies 7 and 8 are clearly below storm wave base deposits. The interpretation of the Deschambault limestone package as only two distinct lithofacies is too simplistic, it does not account for various environments present in the muddy lithologies.

Lithofacies of the Deschambault Formation are fairly similar to other Middle Ordovician shallow water carbonate units (Wilson, 1986). This is especially true when dealing with the North American Middle Ordovician strata. At that time, shallow epeiric seas covered most of the continent (Cecile, 1989); carbonate production was intense in dominant ramp settings (Wilson and Jordan, 1983; Wilson, 1986; Cecile, 1989). Indeed, flat-topped, rimmed carbonate platforms are unknown in that time interval on the western margin of Iapetus ocean (Wilson, 1986).

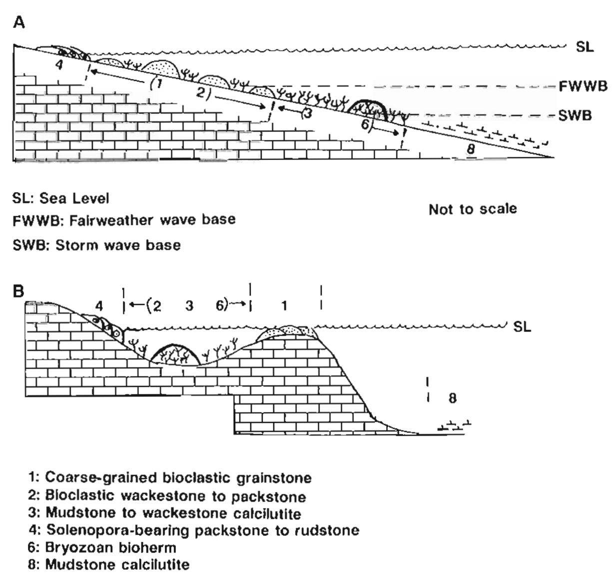

Figure 9. Two possible paleoenvironmental models for the Deschambault Formation together with proposed lithofacies architecture. a) A gently sloping carbonate ramp where lithofacies pattern is controlled by various wave bases. b) A rimmed carbonate shelf where lithofacies architecture is dependant on the nature of the rimmed margin.

Lithofacies distribution of the Deschambault Formation shows that the deeper water lithofacies are best developed in the central and part of the southwestern segments of the studied area (Fig. 3). This spatial arrangement could suggest that the shallower setting was probably located farther northwestward. An alternative model is based on active extensional tectonism as proposed by oil exploration wells study of the B.E.I.C.I.P. (1975). The study documented important thicknesses variations of Trenton units in the Lake St-Pierre (our central region) and Montréal areas. These are interpreted as resulting from more or less quickly subsiding domains related to movements along synsedimentary faults. This tectonic scenario was also proposed by Mehrtens (1979). She suggested that synsedimentary growth faults are responsible for the thickening of units in the central segment of the studied sector. Indeed, number of east-west and northeast-southwest trending faults are present along the western margin of the St. Lawrence Lowlands (Globensky, 1987). Synsedimentary movement along many of these is thought to have lasted from Precambrian up to Late Ordovician time. Extensional activity could account well for the local thickening of the Middle Ordovician Deschambault Formation as well as for the slightly deeper sedimentary settings. Further investigation in the under- and overlying successions will possibly help in elucidating this point.

Lowstand system tracks are, exposures allowing, lacking in the Deschambault Formation. The formation is preliminary interpreted as representing highstand depositional condition with internal, higher order transgressive - progradational couplets. The Deschambault highstand condition is, from general depositional settings proposed from underlying and overlying Middle Ordovician succession (Globensky, 1987), part of a larger scale transgressive system track, tectonically-driven by Taconian loading. The tectonically-induced transgressive conditions in Middle Ordovician time on the St. Lawrence Lowlands is superimposed on a global eustatic sea level rise as suggested by Vail et al. (1977) sea level curves. However, discrimination between these two end members and the importance of their signal in the combined result still has to be worked out.

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Contrasting styles of Lower Devonian sedimentation: the Upper Gaspé Limestones of the southeastern (Percé area) and western (lac Matapédia area) segments of Gaspésie, Quebec

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Abstract

The Lower Devonian Upper Gaspé Limestones succession of the Percé area offers much similitude with the succession of the adjacent northern area. The dominant facies is typical of distal outer shelf setting submitted to occasional storm processes. Regional facies distribution in the Shiphead Formation shows an east-west lateral transition to the siliciclastic Fortin Group. The threefold nature of the Upper Gaspé Limestones is recognized west of Gaspésie, in the lac Matapédia region. Stratigraphic works in on both sides of the lake show that the succession is similar to the one previously recognized in the sector west of Murdochville. It is interpreted to represent sedimentation in a deep shelf environment.

Résumé

La succession des calcaires supérieurs de Gaspé (Dévonien inférieur) de la région de Percé est similaire à celle de la région adjacente au nord avec un faciès dominant caractéristique d'un milieu de plate-forme externe distale soumis à l'action occasionnelle de tempêtes. La répartition régionale des faciès pour la formation de Shiphead montre une transition est-ouest avec le groupe de Fortin de composition silicoclastique. La nature tripartite du schéma lithostratigraphique des calcaires supérieurs de Gaspé est présente à l'ouest de la péninsule gaspésienne, dans le secteur du lac Matapédia. La lithostratigraphie reconnue à l'est et l'ouest du lac est similaire à celle de la région à l'ouest de Murdochville. Elle est interprétée comme résultant d'une sédimentation dans un milieu de plate-forme profonde.

INTRODUCTION

The Lower Devonian (Pragian) Upper Gaspé Limestones were previously studied in the eastern (Lavoie et al., 1990) and central (Lavoie et al., 1991) segments of Gaspésie. Sedimentologic study of the succession showed that the carbonate-dominated facies architecture for the eastern segment of the peninsula was controlled by active tectonism with sedimentary belts following the shape of a pull apart basin (Lavoie, in press). Environments of deposition become progressively deeper from north to south, the shallowest being typical of proximal outer shelf setting (Lavoie, in prep.). Preliminary paleoenvironmental interpretations for the siliciclastic-richer succession in the central segment of the peninsula show that synsedimentary active felsic and mafic volcanism was important in controlling facies patterns, together with nature of sediments shed in the relatively deep water basin developed in that area (Lavoie et al., 1991; Lavoie, 1991).

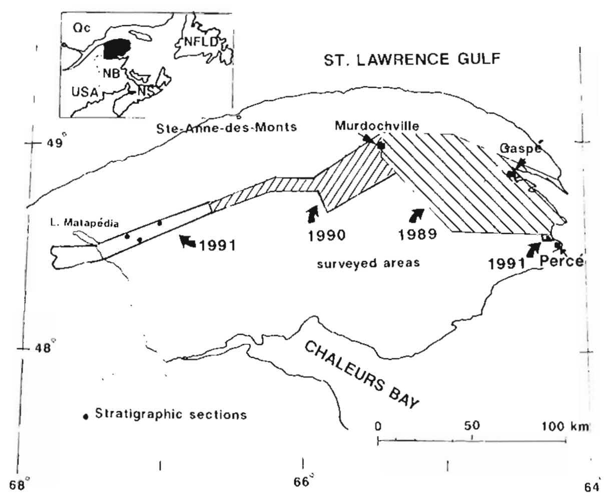

Figure 1. Location map of studied areas in Gaspésie, Québec.

Field work (2 weeks) carried during the 1991 campaign was located in the southeasternmost (Percé area) and westernmost (lac Matapédia area) outcropping sectors of the Upper Gaspé Limestones of Gaspésie (Fig. 1). The Percé area is important for the Early Devonian tectono-sedimentary history of the Gaspé basin since three major dextral strike-slip faults (Bassin du Nord-Ouest, Troisième Lac et Grande Rivière) converge in this relatively small area (Kirkwood, 1989; Fig. 2 herein). Two of these (Bassin du Nord-Ouest and Troisième Lac) are interpreted to have played an important role in controlling the facies architecture in the northern adjacent area (Lavoie et al., 1990; Lavoie, in press). The nature of Lower Devonian facies in the lac Matapédia area is needed to complete the facies picture of the Upper Gaspé

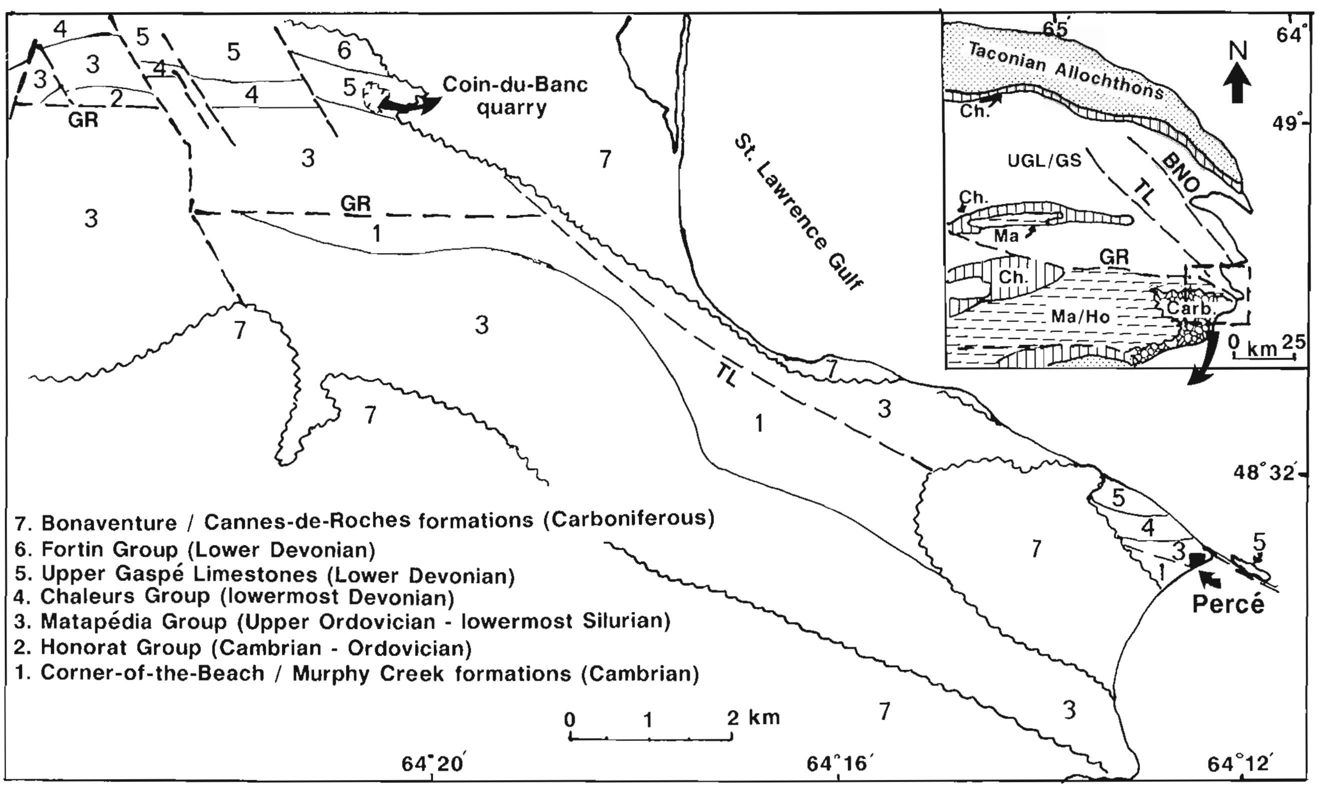

Figure 2. Simplified geological map of the Percé area with location of the Coin-du-Banc quarry, together with (inserted) broad scale geology of eastern Gaspé Peninsula and location of the Percé area. BNO: Bassin Nord-Ouest fault, TL: Troisième Lac fault, GR: Grande Rivière fault, Ho: Honorat Group, Ma: Matapédia Group, Ch.: Chaleurs Group, UGL: Upper Gaspé Limestones, GS: Gaspé Sandstones, Carb.: Carboniferous cover. Geology of Percé area is from Kirkwood (1989).

Limestones and for extending the threefold division of that unit into this sector for which lower Devonian is only known through regional mapping (Béland, 1960).

THE PERCE AREA

The Percé area is structurally complex but the spectacular nature of exposures led geologists to start its geological study in the early years of the century (Clarke, 1908). The geology of this area is summarized in Kirkwood (1989), and schematically sketched on Figure 2.

The Lower Devonian limestones outcropping in the area were previously assigned to the Cap Bon Ami and Grande Grève formations (McGerrigle, 1950). The local term of Murailles Formation (Kindle, 1936) was also used to designate lower Devonian limestones in the Percé sector. Stratigraphic works by Lespérance (1980a) in the Forillon peninsula, approximately 75 km north of the Percé area, resulted in a threefold lithostratigraphic framework (Forillon, Shiphead and Indian Cove formations) for these Lower Devonian limestones. It had proven to be applicable over all of Gaspésie (Lespérance 1980a,b; Rouillard, 1986; Lavoie et al., 1990 and 1991). Kirkwood (1989) recognized the three units in the Percé area with, as everywhere else in Gaspésie, the Forillon and Shiphead formations being synonymous to the Cap Bon Ami, and the Indian Cove Formation to the Grande Grève. The term Murailles Formation is synonymous to the Forillon Formation.

One stratigraphic locality was carefully described. It is a disused quarry located 12 km northwest of Percé in the vicinity of Coin-du-Banc (Fig. 2). That locality is located almost exactly on the Troisième Lac fault trace. There, close to 200 m of the Upper Gaspé Limestones are exposed (Fig. 3), with the Forillon and Shiphead formations being best represented.

Forillon Formation

The Forillon Formation overlies, in faulted contact, the undivided White Head Formation of Late Ordovician to earliest Silurian age (Kirkwood, 1989). The 128 m measured for the Forillon Formation are composed of two distinct rock packages (Fig. 3), correlatable with the Chemin du Roy and Cap Gaspé members previously proposed in the Forillon peninsula area (Lavoie et al., 1990; Lavoie, in press).

At the bottom, the succession exposes 56 m of the Chemin du Roy Member. However, exact thickness is unknown due to faulting. This member is made up of dolomitic siliciclastic mudstones (Fig. 4a) occurring in beds ranging from 2 to 12 cm in thickness (average 8 cm). These are thoroughly bioturbated by *Zoophycos*. The mudstones are interlayered with beds 5 to 15 cm thick of shaly calcilutite with mudstone texture that are finely parallel-laminated with laminations uncommonly convoluted. The amount of calcilutite increases up-section from a minimum of 10% for the lowermost exposures, up to 100% as siliciclastic mudstones disappear at the top of the member. No macrofauna was observed in that interval.

The overlying 72 m are correlated with the Cap Gaspé Member. Transition with the underlying unit is gradual. The Cap Gaspé Member is almost entirely composed of siliceous calcilutite with a mudstone texture. The poorly fossiliferous (pelmatozoans) calcilutite occurs in planar beds ranging from 6 to 20 cm in thickness (Fig. 4b). Few fine parallel laminations, locally intensely bioturbated by *Zoophycos*, are

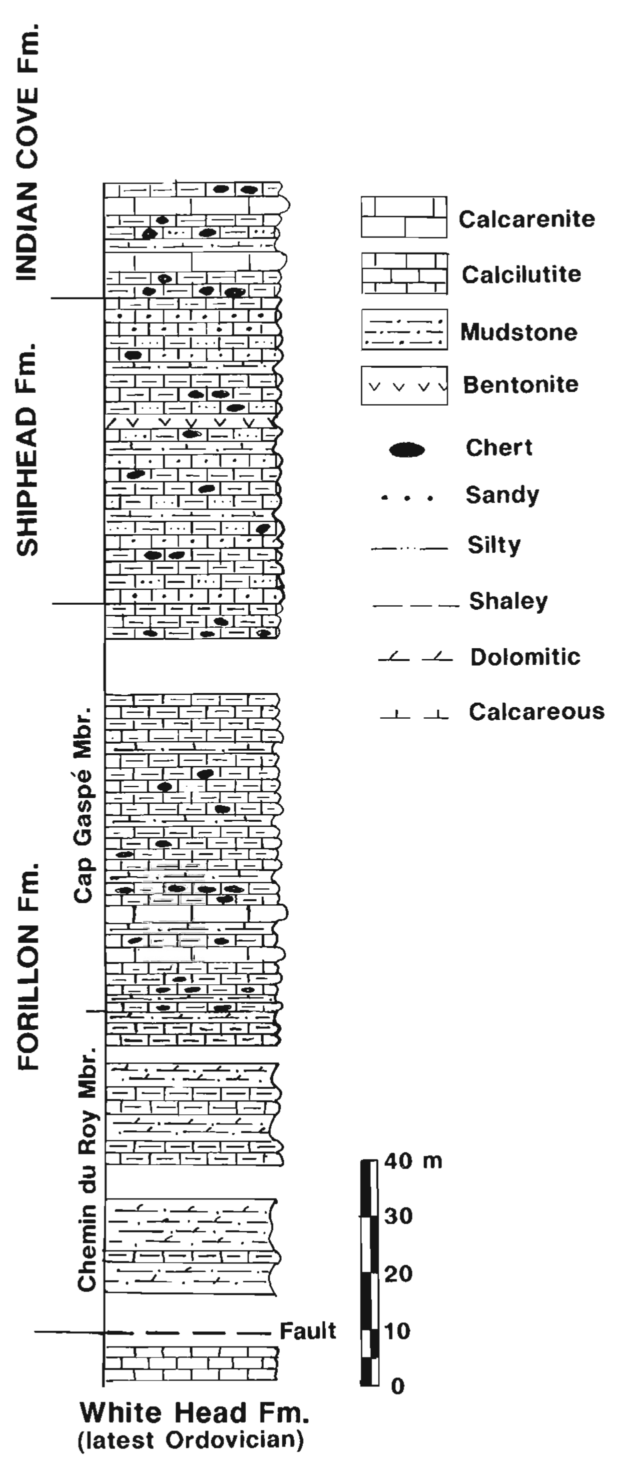

Figure 3. Lithostratigraphic section of the Upper Gaspé Limestones at the Coin-du-Banc quarry.

developed. Rare interbeds are noted. These constitute less than 5% of rock volume and are made up of millimetre- to centimetre-sized interbeds of calcareous mudstones. This unit is characteristically silicified and chertified with chert averaging less than 10% of the rock. Chert occurs as centimetre-sized stringers and nodules. Finally, scattered in the calcilutite, beds and lenses, less than 25 cm thick, of crinoid-rich, packstone to grainstone calcarenites (Fig. 4c) are found. These can locally constitute up to 40% of rock per metre of section, but their overall amount averages less than 5%. The calcarenites show common sharp bases with few erosive examples. No internal sedimentary structures were observed other than some faint grading and local imbrication of long axis of small shells.

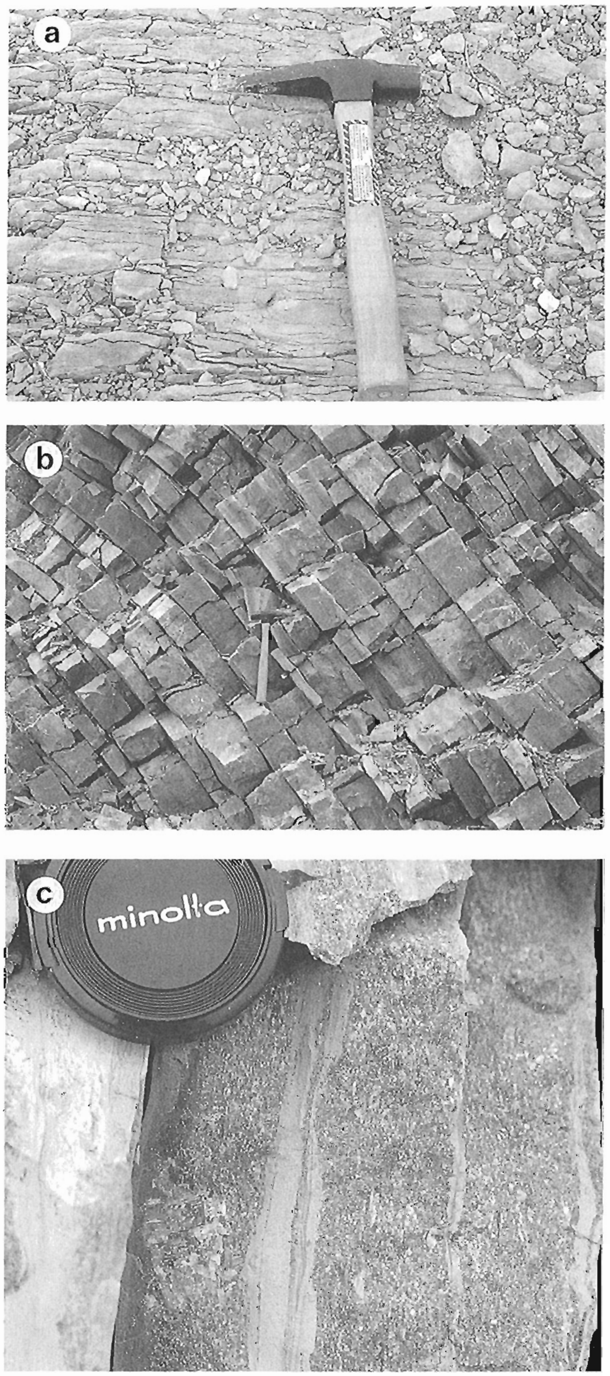

Figure 4. Forillon Formation at the Coin-du-Banc quarry. **a**) Dolomitic siliciclastic mudstone of the Chemin du Roy Member. **b**) Planar-bedded mudstone calcilutite of the Cap Gaspé Member. Hammer for scale. **c**) Faintly graded, bioclast-rich packstone calcarenite showing episodic current activities. Cap Gaspé Member.

Shiphead Formation

The Shiphead Formation conformably overlies the Cap Gaspé Member of the Forillon Formation (Fig. 3). The transition is rather diffuse and arbitrary, being marked by an increase in siliciclastic content of the carbonate. Distinction from the underlying Forillon Formation relies on the presence of silt- to sand-sized quartz particles distributed in the

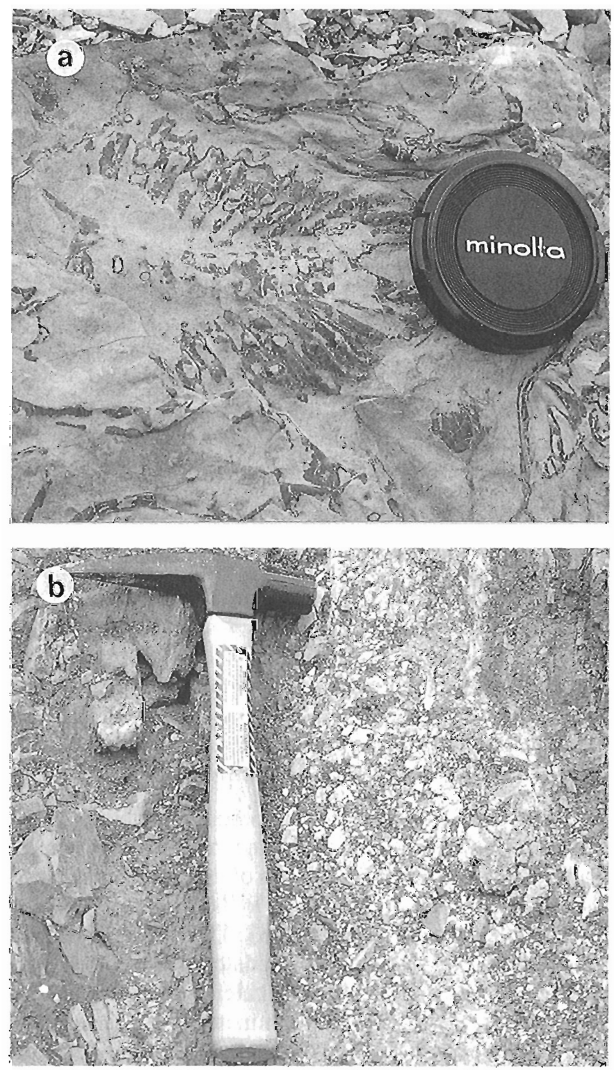

Figure 5. Shiphead Formation at the Coin-du-Banc quarry. **a**) Packstone with abundant trilobite skeletons. **b**) Bentonite layer (pale shaded) interbedded with mudstone calcilutite.

calcilutite or concentrated in burrows. The 55 m thick Shiphead Formation is mostly composed of siliceous calcilutite with a mudstone texture. These occur in planar beds ranging from 5 to 25 cm in thickness (average 12 cm) with few, fine parallel laminations. The calcilutite commonly shows many transitions from reddish to grey in color. Such changes occur not only from bed to bed but also inside an individual horizon, testifying for diagenetic control. Interbeds are locally observed, and made up of centimetre-sized beds of calcareous siliciclastic mudstones. The commonly bioturbated (*Zoophycos*) calcilutite is almost devoid of bioclast, except for some trilobites remains forming one distinctive packstone horizon (Fig. 5a). Chert, although present, is less abundant than in the Forillon Formation. Presence of some centimetre-sized bentonite beds is noted (Fig. 5b). These fine grained volcaniclastic beds are typical of the Shiphead Formation over all of Gaspésie (Lavoie et al., 1990, 1991).

Indian Cove Formation

The Indian Cove Formation constitutes the 20 uppermost metres exposed in the quarry. Regional mapping showed that the unit can reach almost 250 m in the Percé area (Brisebois, 1981). The Indian Cove Formation conformably overlies the Shiphead Formation. Transition is given by the disappearance of silt- to sand-sized siliciclastic particles in the limestones. In the quarry, the Indian Cove Formation is made up of siliceous to cherty calcilutite with mudstone texture similar to previously described Cap Gaspé Member of the Forillon Formation. The calcilutite is commonly bioturbated by *Zoophycos*, but is almost devoid of macrofauna besides rare pelmatozoans. Interbeds are few and made up of millimetre-sized beds of impure calcilutite. Some packstone calcarenite beds and lenses, less than 25 cm thick, were found in the uppermost metres of the succession. These share the same field characteristics with the previously described calcarenites of the Cap Gaspé Member.

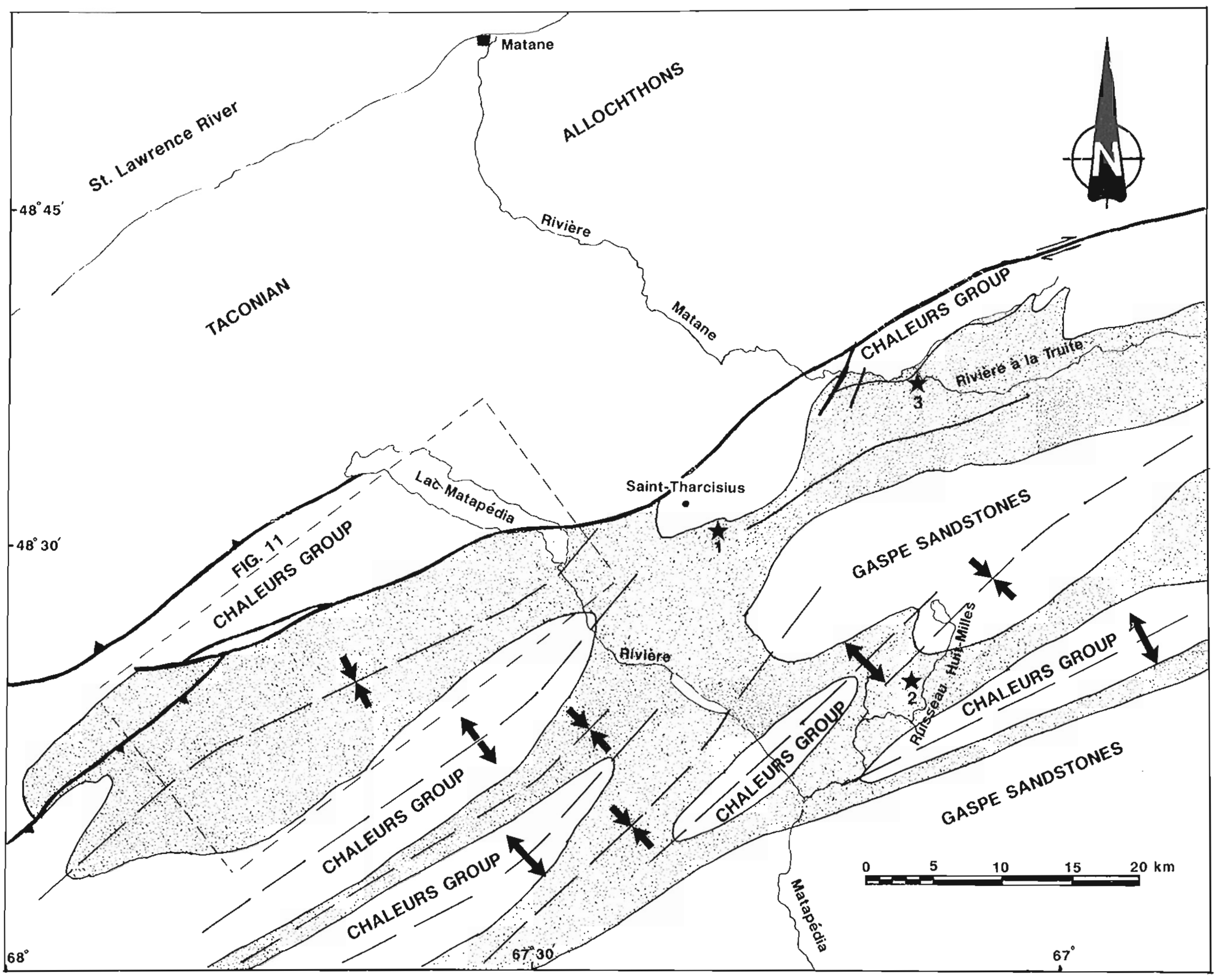

Figure 6. Simplified geological map of the western segment of Gaspésie. Upper Gaspé Limestones outcrop sectors are shaded. Stars are for sections on Figure 7. Geology modified from unpublished MERQ compilation maps.

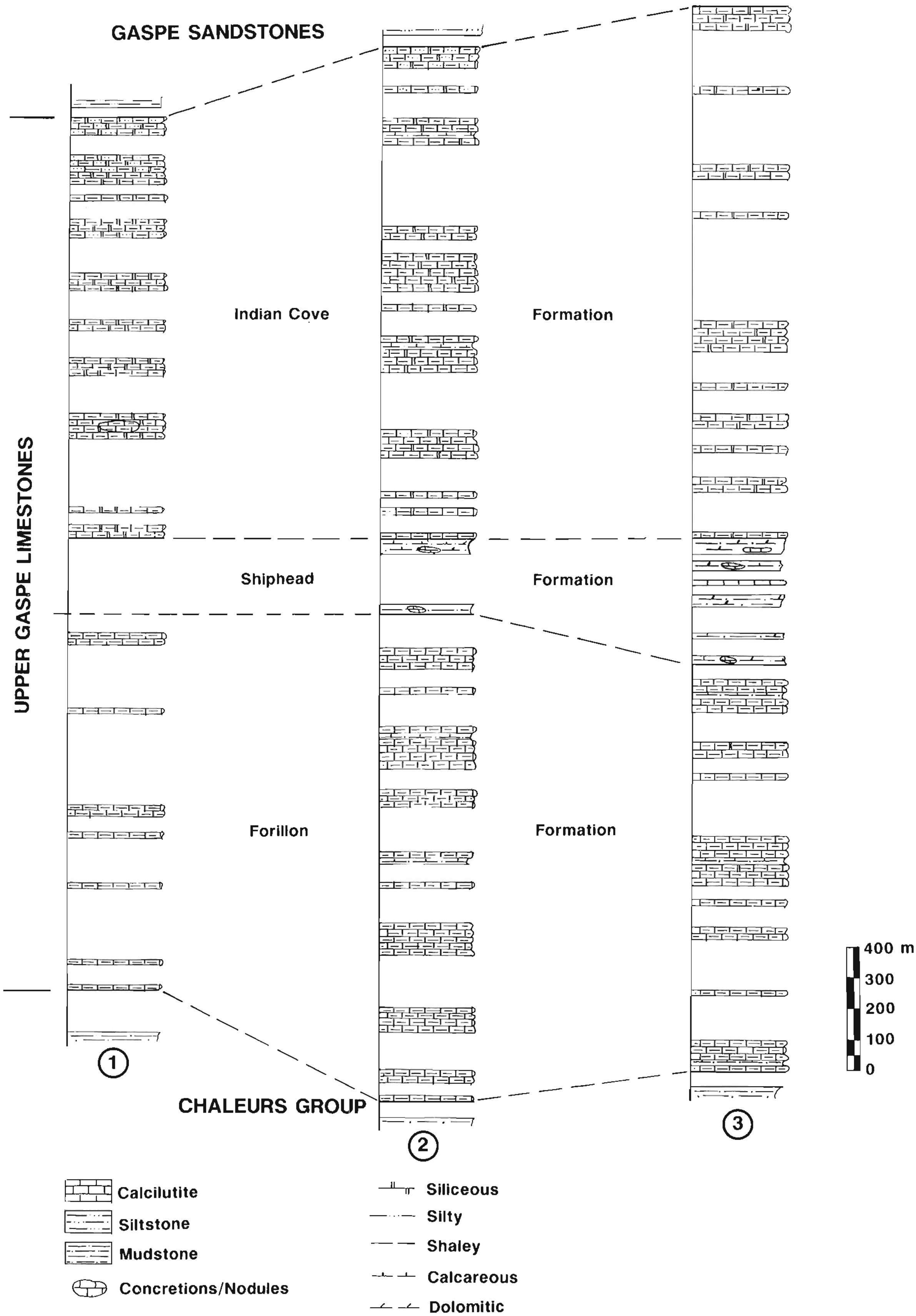

Figure 7. Lithostratigraphic correlations for the Upper Gaspé Limestones, east of lac Matapédia. The Shiphead – Indian Cove contact is used as datum line. Location of sections on Figure 6.

Summary of Percé area

The Upper Gaspé Limestones of the Percé area outcrop south of the Bassin Nord-Ouest and in the vicinity of the Troisième Lac faults (Fig. 2). In the northern adjacent area, the area delineated by these structural elements is characterized by facies typical of distal outer shelf settings (Lavoie, in press; Fig. 2 herein). In the Percé area, the Upper Gaspé Limestones are typified by poorly fossiliferous, thoroughly bioturbated calcilutite with mudstone texture. Such lithology is interpreted to likely represent outer shelf deposit below fairweather wave base, an interpretation based on the fine grained nature of the sediment, the lack of current- or wave-reworking of the sedimentary structures, and the presence of *Zoophycos*, a below wave base ichnofauna (Seilacher, 1964, 1967). Calcarenites are embodied in the fine grained succession, these are interpreted to represent storm remobilization of bottom sediment. The storm origin of these beds is deduced from sedimentary structures (some scouring, grading, and imbrication) in an otherwise succession typical of outer shelf deposition and devoid of fairweather-induced structures. From the facies model proposed by Handford (1986), the calcarenites in the Percé area likely represent distal storm deposits. As everywhere in the eastern segment of the Gaspé Peninsula, the Shiphead Formation is characterized by a higher siliciclastic content together with presence of fine grained volcaniclastic beds. Interestingly, a few kilometres west of the quarry, the base of the much thicker Shiphead Formation is given by a 50 m thick succession of thickly bedded, quartz-rich conglomerates and sandstones (Bourque et al., in press), overlain by impure limestones. Farther west, in the Grande Rivière area (Lavoie et al., 1990), the Shiphead Formation is replaced by a siliciclastic succession assigned to an informal tongue of the Fortin Group (Lavoie, in press). This east-west facies transition is coupled with thickness increase of the unit.

THE LAC MATAPÉDIA AREA

The lithostratigraphy of the Upper Gaspé Limestones in the area east of lac Matapédia was previously studied by Rouillard (1986). He was able to extend the threefold lithostratigraphic framework of Lespérance (1980a). However, the area west of that lake is currently only known through regional mapping by Béland (1960), with Lower Devonian rocks still assigned to the Cap Bon Ami and Grande Grève formations. Moreover, the two formations were not mapped as separate units. The objectives of our survey in the area was: 1) to extend (if possible) the lithostratigraphic framework of the Upper Gaspé Limestones in the area west of lac Matapédia and, 2) to identify depositional environment for the Upper Gaspé Limestones.

The area east of lac Matapédia

Mapping by Ollerenshaw (1967) allowed to recognize the presence of the Cap Bon Ami and Grande Grève formations. Rouillard (1986) described the Upper Gaspé Limestones succession outcropping in the Ruisseau Huit-Milles area. At this locality, he recognized a threefold rock package with a lower limestone unit, a middle siliciclastic mudstone unit and, an upper unit formally designed as the Indian Cove Formation.

Our work concentrated on three sectors, the Ruisseau Huit-Milles, the Rivière à la Truite, and the Saint-Tharcisius areas (Fig. 6). The resulting lithostratigraphy (Fig. 7) is rather uniform, with variations in thicknesses likely related to unexposed structural elements in this poorly outcropping sector. No localities showed stratigraphic contacts between formations.

Forillon Formation

The lower limestone unit of Rouillard (1986) is here assigned to the Forillon Formation, based on lithological grounds as well as on the lithostratigraphic framework proposed for the area adjacent to the east (Lavoie et al., 1991). Assumed thickness of the Forillon Formation ranges from about 600 m up to almost 800 m with the uncertainty previously discussed. At the three localities, the Forillon Formation is a rather monotonous succession dominated by unfossiliferous, shaly calcilutites with a mudstone texture (Fig. 8a). The calcilutite occurs in planar beds ranging from 5 to 40 cm in thickness (average 12 cm). These locally display faint parallel laminations with some evidence of burrowing. Interbeds are uncommon but when present, they are made up of siliciclastic calcareous mudstones occurring in centimetre-sized beds (Fig. 8b). Siliceous calcilutite is rarely observed and occurs as unrecessed beds in the more shaly and weathered impure limestone succession.

This succession is similar to the Forillon Formation for the entire eastern adjacent area, between the Ruisseau Burntwood and the Etang à la Truite (Lavoie et al., 1991). Local variations in the siliciclastic content of the rock result either in shaly calcilutites or highly calcareous mudstones. A similar lithologic succession was formally designed as the Lesseps Member of the Forillon Formation (Lavoie et al., 1991). Such a designation could also be applied to the Forillon Formation in this sector. As recognized by Rouillard (1986), the succession is different from the Forillon Formation succession on Forillon peninsula and east of this sector up to Murdochville (Lavoie et al., 1990).

Shiphead Formation

The middle mudstone unit of Rouillard (1986) is here assigned to the Shiphead Formation, based on lithostratigraphic and lithological grounds. The Shiphead Formation is poorly represented in the area, and it does not even outcrop in the Saint-Tharcisius sector. Assumed thickness varies from 125 m to about 200 m with the same uncertainty as previously discussed. The poorly outcropping nature of the Shiphead Formation, as in the eastern adjacent area, likely relies on its lithological nature (Lavoie et al., 1991). The most typical lithology of the Shiphead Formation is given by intensely weathered siliciclastic mudstones that are locally dolomitized (Fig. 9a). Bedding is commonly masked by fissility and cleavage. These mudstones locally bear centimetre-sized ellipsoid carbonate concretions. These

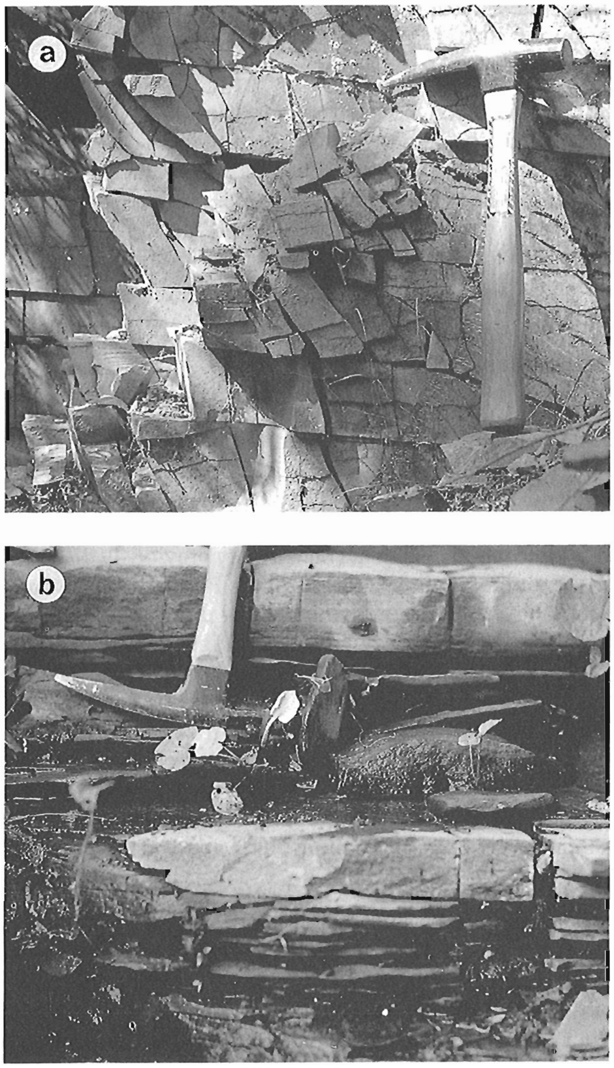

Figure 8. Forillon Formation. **a**) Massive, shaly calcilutite with mudstone texture. Section 3. **b**) Shaly calcilutite interbedded with recessed siliciclastic mudstone showing intense cleavage. Section 2.

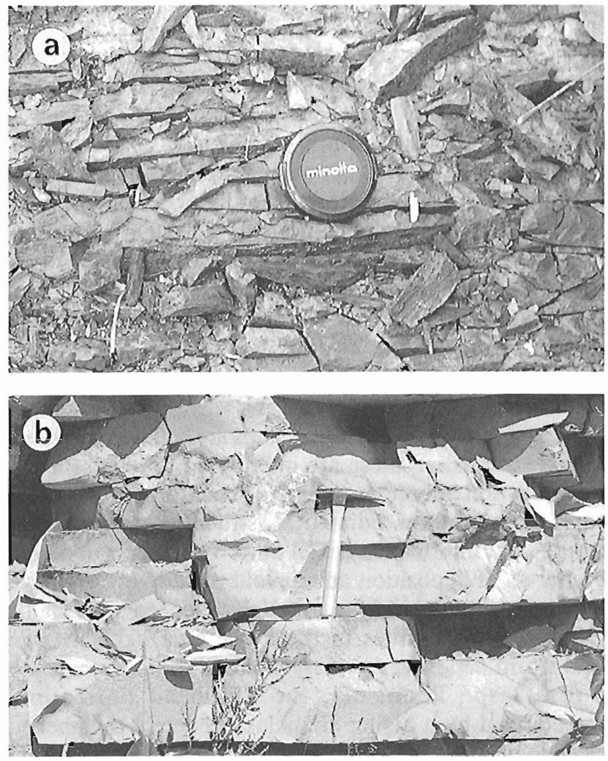

Figure 9. Shiphead Formation. **a**) Thinly-bedded, more or less dolomitized, siliciclastic mudstone. Section 3. **b**) Parallel-bedded, siliciclastic-free, dense calcilutite. Section 2.

can make up to a maximum of 5% of the rock per metre of section. An 8 m thick unit of siliciclastic-free, dense calcilutite (= sublithostratigraphic limestone of other workers) with mudstone texture (Fig. 9b) was observed in the Rivière à la Truite area. This unit is underlain and overlain by the previously described siliciclastic mudstones. Rouillard (1986) based on SAREP's (Société Acadienne de Recherches Pétrolières) work, mentioned the presence of andesitic tuffs in the upper part of the Shiphead Formation. Unfortunately, these were not observed in our stratigraphic survey.

The siliciclastic-dominated succession assigned to the Shiphead Formation is fairly similar to the Shiphead succession in the eastern adjacent area (Lavoie et al., 1991), where siliciclastic-rich facies are prominent. However, no coarse-grained siliciclastic (Brandy Brook Member) or volcanogenic facies were noted, a situation that could be explained by distance from the source areas. As noted by Rouillard (1986), the Shiphead succession is fairly dissimilar with the one of the type area, that is the Forillon Peninsula.

Indian Cove Formation

The uppermost unit of the Lower Devonian limestones in this area was assigned to the Indian Cove Formation by Rouillard (1986). Our observations agree with that assignment. Assumed thickness of the Indian Cove Formation ranges from 675 up to 850 m. The Indian Cove Formation of this area is composed of siliceous shaly calcilutite with a mudstone texture (Fig. 10a). These mostly unfossiliferous calcilutites occur in planar beds ranging from 5 to 25 cm in thickness (average 15 cm). In the upper part of the formation, they show parallel laminations and lenses of dolomitic siltstone that are locally bioturbated by *Zoophycos*. Chert that typifies the unit in eastern Gaspésie (Lavoie et al., 1990), is not observed. Some outcrops show the presence of poorly silicified, centimetre-sized nodules (Fig. 10b) in the otherwise highly silicified calcilutite.

The siliceous calcilutite is typical of the Indian Cove Formation from Murdochville up to lac Matapédia. However, volcanogenic facies observed in the more central segment of the peninsula (Lavoie et al., 1991) are not observed in the lac Matapédia sector, arguing again for spatial restricted occurrence of volcanic centres.

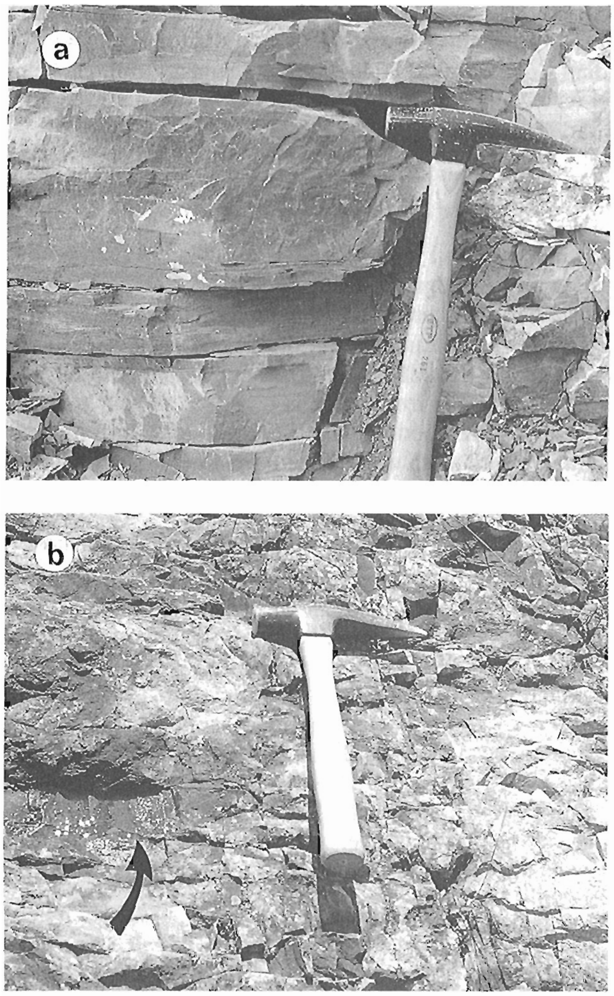

Figure 10. Indian Cove Formation. **a**) Thinly- to thickly-bedded siliceous calcilutite. Section 2. **b**) Highly silicified calcilutite with slightly recessed, less-silicified nodule (arrow). Section 1.

The area west of lac Matapédia

Lower Devonian limestones outcrop in a series of synclines west of lac Matapédia (Béland, 1960; Fig. 6 herein) up to the Témiscouata region (Bourque and Gosselin, 1987, 1988). In that area, the former designation as Cap Bon Ami and Grande Grève formations is still in use for these rocks, but no detailed stratigraphic works are available for these units.

Reconnaissance work was carried over the entire Lac Mitis Syncline (Béland, 1960), immediately adjacent to the lac Matapédia (Fig. 1, 6, and 11). In that area, Béland's map (1960) shows the distribution of Lower Devonian limestone-dominated units, but without distinguishing the Cap Bon Ami and Grande Grève formations. Our survey shows that, as in the area adjacent to the east, the Upper Gaspé Limestones offer the same threefold lithostratigraphic framework previously presented. Figure 11 sketches, for the first time, the distribution of the three units with major structural elements. However, no detailed section measuring was done.

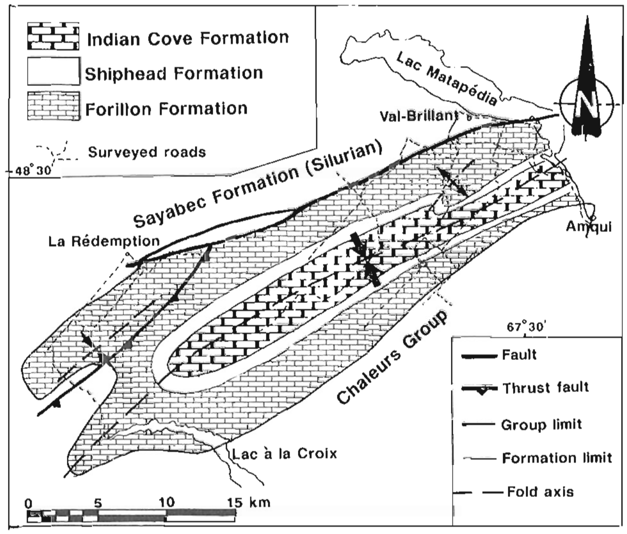

Figure 11. Schematic geological map of the Lac Mitis syncline (west of lac Matapédia) showing the spatial distribution of the Forillon, Shiphead, and Indian Cove formations together with major structural elements. See Figure 6 for general location of the area.

Forillon Formation

The Forillon Formation is made up of unfossiliferous shaly calcilutite with mudstone texture occurring in planar beds ranging from 5 to 30 cm (average 15 cm) in thickness (Fig. 12a). No interbeds were observed. Thickness of the unit is assumed to be about 750 m and compares well with the 600 to 800 m estimate for the eastern adjacent area.

Shiphead Formation

As in the area adjacent to the east, outcrops of the Shiphead Formation are sparse. Lithologies assigned to the Shiphead are only observed in the eastern segment of the Lac Mitis syncline. Elsewhere, an interval devoid of outcrops is always present between typical Forillon and Indian Cove lithologies. Lithologies assigned to the Shiphead Formation are siliciclastic-rich. The best exposed are made up of finely laminated siltstones that are locally calcareous. They occur in planar beds a few centimetres thick (Fig. 12b). Some intensely weathered siliciclastic mudstones are seldom observed, and outcrop as slightly displaced debris. These mudstones are similar to those in the area adjacent to the east. Maximum stratigraphic thickness for the Shiphead Formation is about 200 m.

Indian Cove Formation

The Indian Cove Formation is represented by siliceous shaly calcilutite with mudstone texture. It occurs in planar beds ranging from 6 to 25 cm thick (Fig. 12c). The calcilutite locally shows some internal parallel laminations that are sometimes bioturbated by *Zoophycos*. No interbeds were observed. Interestingly, the degree of silicification is slightly higher in the Lac Mitis syncline when compared to the area adjacent to the east. Diffuse chertified zones in limestones

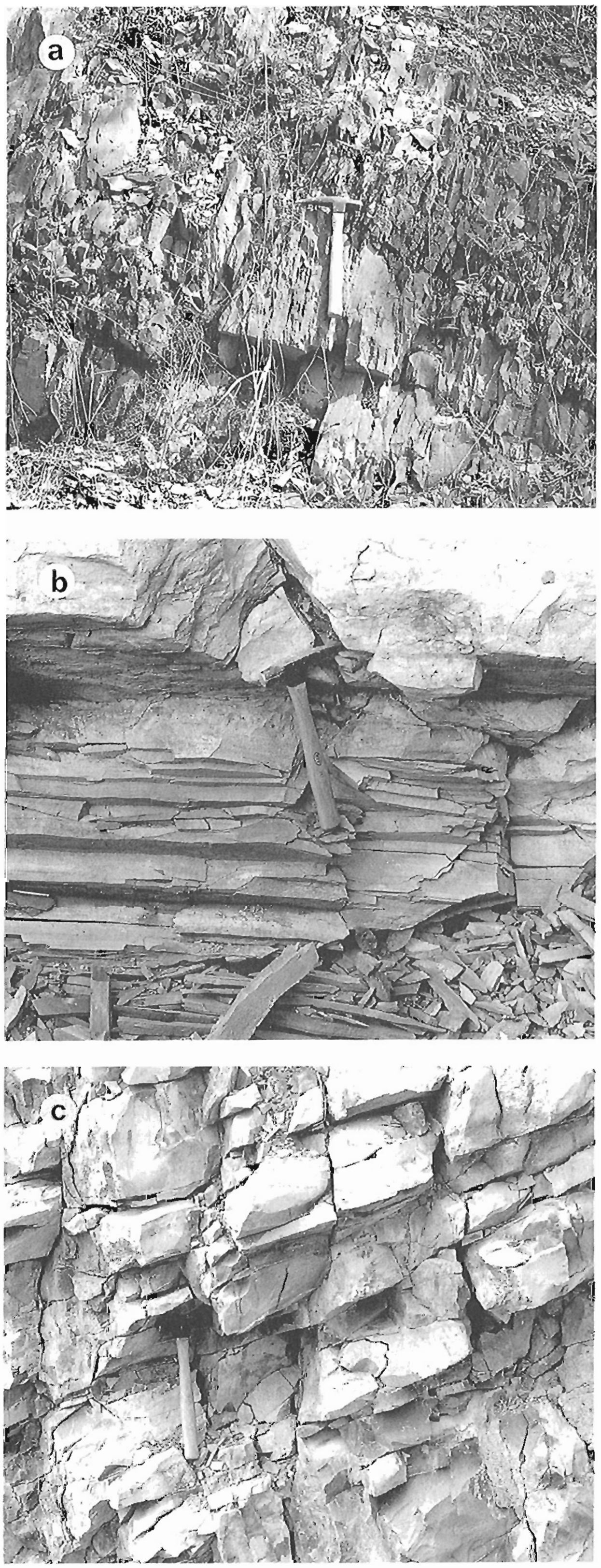

Figure 12. Upper Gaspé Limestones – Lac Mitis syncline. a) Shaly calcilutite with prominent cleavage. Forillon Formation. b) Planar-bedded calcareous siltstone (Shiphead Formation) overlain by siliceous calcilutite (Indian Cove Formation). c) Massive beds of siliceous to seldom cherty calcilutite. Indian Cove Formation.

were noted in some outcrops. However, chertification never culminates in the formation of nodules or stringers. The Indian Cove Formation outcrops in the heart of the syncline and its true thickness is unknown. However, a minimum of 415 m is present.

Summary of the Lac Matapédia area

The areas east and west of Lac Matapédia are typified by a similar lithostratigraphic succession of Lower Devonian Limestones. The threefold lithostratigraphic framework for the Upper Gaspé Limestones, previously established in the eastern sector (Rouillard, 1986), is confirmed with our own observations. This framework is proposed for the area west of the lac Matapédia. The Forillon and Indian Cove formations are both represented by fine grained, impure carbonates that are locally finely laminated and burrowed. This succession is similar to the typical lithology of these two units in the central part of the Gaspésie (Lavoie et al., 1991). This lithology likely represents deposition on a relatively deep shelf with the dominant lime muds originating from unexposed shallow water platform(s?) (Lavoie, in press, in prep.). Further works will probably show that the silicification of the Indian Cove Formation is related to the presence of an important sponge fauna. The Shiphead Formation, as in the central segment of the Gaspésie (Lavoie et al., 1991), is typified by a higher siliciclastic content resulting in siliciclastic lithologies. The poorly outcropping nature of the unit precludes any precise paleoenvironmental interpretation for this formation which could also be a deep water unit.

CONCLUSIONS

The Upper Gaspé Limestones in both the southeasternmost (Percé) and westernmost (Lac Matapédia) sectors of Gaspésie offer a threefold lithostratigraphic package made up of the Forillon, Shiphead and Indian Cove formations. However, differences between these two areas are noticeable. Facies in the Percé area are of outer shelf type with presence of distal storm deposits. Similar Lower Devonian limestone facies are observed in the northern adjacent area, south of the

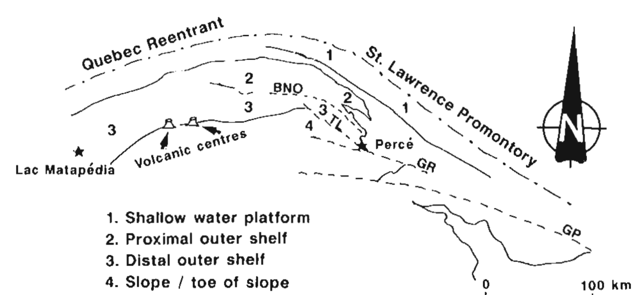

Figure 13. Preliminary interpretation of Early Devonian paleogeography on Gaspé Peninsula with location of the Percé and lac Matapédia sectors, modified from Lavoie (1991). BNO: Bassin Nord-Ouest fault, TL: Troisième Lac fault, GR: Grande Rivière fault, GP: Grand Pabos fault. Palinspastic restored map is from Bourque et al. (in press).

Bassin Nord-Ouest fault. It suggests that synsedimentary faulting also occurred in the Percé area. The northeastern unexposed shallow water shelf, recognized from petrographic works and regional facies distribution (Lavoie et al., 1990, in press, in prep.), likely extended southward (Fig. 13). Moreover, the north-south transition from the Upper Gaspé Limestones to the Fortin Group previously documented in the adjacent area (Lavoie et al., 1990), is also recognized in the Shiphead Formation, this time along an east-west transect from Percé to Grande Rivière areas.

The Upper Gaspé Limestones lithologic succession in the Lac Matapédia area, is similar to most of the previously documented facies in the central segment of the peninsula (Lavoie et al., 1991) with the noticeable absence of volcanic and coarse grained sedimentary lithologies. Preliminary paleoenvironmental interpretation is that of a distal deep shelf setting away from major shallow water environment but still in the reach of winnowed muds originating from the latter.

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Origin of ophicalcites in ophiolitic mélanges, southern Quebec Appalachians: field occurrence and preliminary geochemistry

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Abstract

The Rivière des Plantes ophiolitic Mélange in southern Quebec contains carbonatized ophiolites known as ophicalcites. The carbonatization followed serpentinization and shearing of ultramafic material. The first carbonate is given by mechanically deposited micrite in fractures. Precipitation of isopachous crusts of calcite followed, and finally blocky carbonates sealed the remaining pore space. The complex history of carbonatization is illustrated by the crosscutting relationships between fractures and the various types of fillings. Late cements, as revealed by cathodoluminescence, are composed of two distinct phases; 1) early dull luminescent granular crystals and, 2) late bright luminescent, large xenomorphic crystals. Both cements yielded fairly similar $\delta^{13}C$ values, near the marine isotopic composition. However, $\delta^{18}O$ are slightly different for both cements and are highly depleted when compared to normal seawater. These values were possibly generated by precipitation from hydrothermal marine-like waters, followed by an isotopic re-equilibration during greenschist metamorphism.

Résumé

Le Mélange ophiolitique de la Rivière des Plantes au sud du Québec, renferme des ophiolites carbonatisées ou ophicalcites. La carbonatisation suit la serpentinisation et le cisaillement des roches ultramafiques. Elle est initiée par la sédimentation de micrite dans des fractures, suivie par une précipitation de couches isopaques de calcite et complétée par des ciments équigranulaires de carbonates. La carbonatisation est caractérisée par un patron complexe de fracturation et de remplissages variés. Les ciments tardifs, tel que révélé en cathodoluminescence, sont composés d'une phase initiale de cristaux à luminescence terne, suivie d'une phase ultérieure de grandes plages à luminescence brillante. Les deux phases ont des valeurs semblables en $\delta^{13}C$, telles que celles d'origine marine. Les valeurs en $\delta^{18}O$ sont différentes pour les deux ciments et fortement appauvries en isotopes lourds comparativement au signal marin. Elles résultent possiblement d'une précipitation à partir d'un fluide hydrothermal d'origine marine, suivie d'une rééquilibration isotopique lors du métamorphisme au faciès des schistes verts.

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INTRODUCTION

Ophiolites and ophiolitic mélanges are well exposed in southern Quebec Appalachians (Fig. 1) and are important rock units since they mark a major suture zone (the Baie Verte - Brompton line) between continental rocks of the Humber zone and the thrusted over oceanic rocks of the Dunnage zone (Williams and St-Julien, 1982; Cousineau, 1991).

A volumetrically minor component of ancient and modern ultramafic successions are carbonatized ophiolites known as ophicalcites. These are ultramafic rocks that were first serpentinized and then carbonatized. They are generally a few tens of metres thick. Ophicalcites are made of brecciated serpentinite with a complex history of micrite infills, fracturation, and carbonate cementation. However, the origin of ophicalcites is poorly known, being interpreted either as product of high temperature metamorphism (Trommsdorff et al., 1980), submarine hydrothermal weathering (Spooner and Fyfe, 1973), combined tectonic and sedimentary events (Bernouilli and Weissert, 1985), and even as product of late, soil-related processes (Folk and McBride, 1976). The discovery of actively-forming ophicalcites in modern oceanic transform fault settings (Melson et al., 1972; Bonatti et al., 1974) supports the tectono-sedimentary scenario for at least some of these occurrences. The proposed scenario is that serpentinized ophiolites rose as diapirs, along faults, to be later carbonatized (Bonatti et al., 1977; Cortesogno et al., 1981). This contention is important since presence of ophicalcites could mark the position of ancient transform faults in oceanic domains (Saleeby, 1984; Bernouilli and Weissert, 1985).

In southern Quebec Appalachians, carbonatized mafic and ultramafic rocks are present along major faults and inside ophiolitic mélanges (De Broucker, 1986; Cousineau, 1991). In particular, the Rivière des Plantes ophiolitic Mélange (Fig. 1, 2), besides containing various lithologies as diversified as serpentinized ophiolitic rocks, granitoids, and metasandstones, also comprises carbonatized upper mantle rocks or ophicalcites (Cousineau, 1991). All rocks in this mélange are now metamorphosed to the regional greenschist facies (Cousineau, 1991).

The objectives of this research are: 1) to decipher the origin of southern Quebec Appalachians ophicalcites and, 2) to document the stages of transformation from fresh to serpentinized and ultimately to carbonatized ultramafic rocks. This contribution presents field observations and preliminary geochemistry for the ophicalcites in the Rivière des Plantes ophiolitic Mélange.

GEOLOGICAL SETTING

The Rivière des Plantes ophiolitic Mélange (RPOM) is located about 75 km south-southeast of Québec City (Fig. 1) and belongs to a more important unit known as the Saint-Daniel Mélange (Cousineau, 1991), interpreted as a former taconian accretionary prism (St-Julien and Hubert, 1975; Cousineau, 1988). Readers interested in the pre-taconian (Middle Ordovician) paleogeography and tectonic evolution of this segment of the Appalachians are referred to detailed studies by St-Julien and Hubert (1975) and Cousineau (1988, 1991).

The RPOM is made of compositionally very diversified slivers where fresh, serpentinized, and carbonatized ultramafic rocks are abundant (Fig. 2). Among these ultramafic units, harzburgite and dunite are prominent; wherlite, orthopyroxenite, gabbro, and basalt occur locally (Cousineau, 1991).

FIELD OCCURRENCE OF OPHICALCITES

In field outcrops, carbonatization of ultramafic rocks is, the least to say, highly variable. In general terms, ophicalcites are composed of millimetre- to metre-sized serpentinized ophiolitic material embodied in a matrix composed of micrite and/or carbonate cement. All ultramafic rocks are intensely serpentinized but deformation is variable, some showing original magmatic features whereas others are pervasively sheared. However, transition from serpentinized ultramafic rocks to more or less carbonatized rocks can be reconstructed from field (Fig. 3a-d) and petrographic works. The degree of

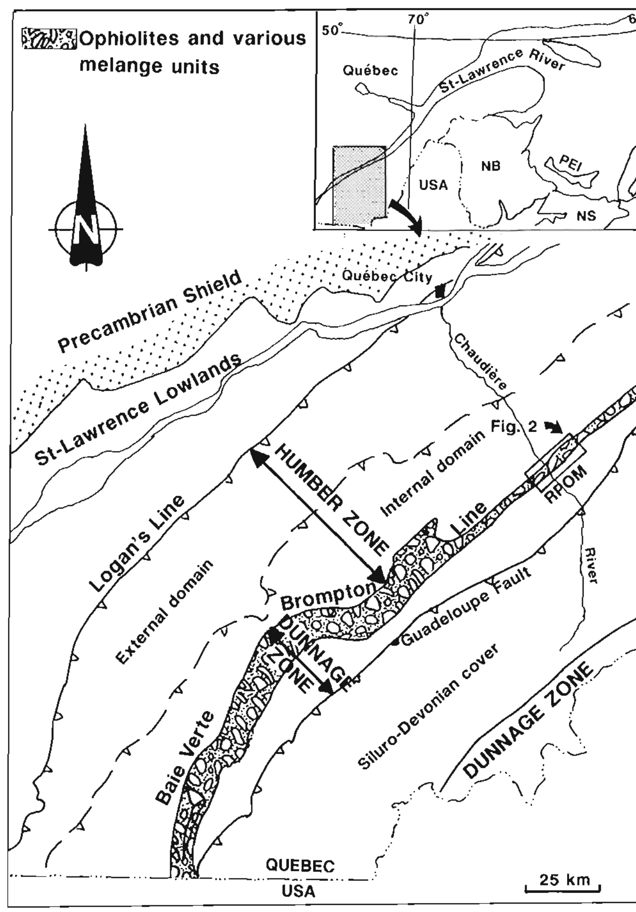


Figure 1. Geology of southern Québec Appalachians with their main tectonostratigraphic divisions together with location of ophiolites and ophiolitic mélanges. (RPOM): Rivières de Plantes ophiolitic Mélange. Modified from Slivitzky and St-Julien (1987).

carbonatization of ultramafic rocks goes from faintly fracturated serpentinite with carbonatized phenocrysts and to some carbonate matrix infills, up to intensely brecciated units showing various carbonate infills and cements with very few preserved indicators of the previous lithology. Uncommonly ophicalcite protrude into the shaly rocks of the Saint-Daniel Mélange (Fig. 4).

The complexity of carbonatization of the ultramafic rocks is obvious in the field. It is represented by three major carbonate generations. The most important generation, the oldest one, is present in all field occurrences. It is made up of grey to less common reddish micrite mechanically deposited in various fractures and voids. In the most carbonatized examples, it can make up to 50% of total carbonates. In the least altered ultramafic rocks, it is frequently the only

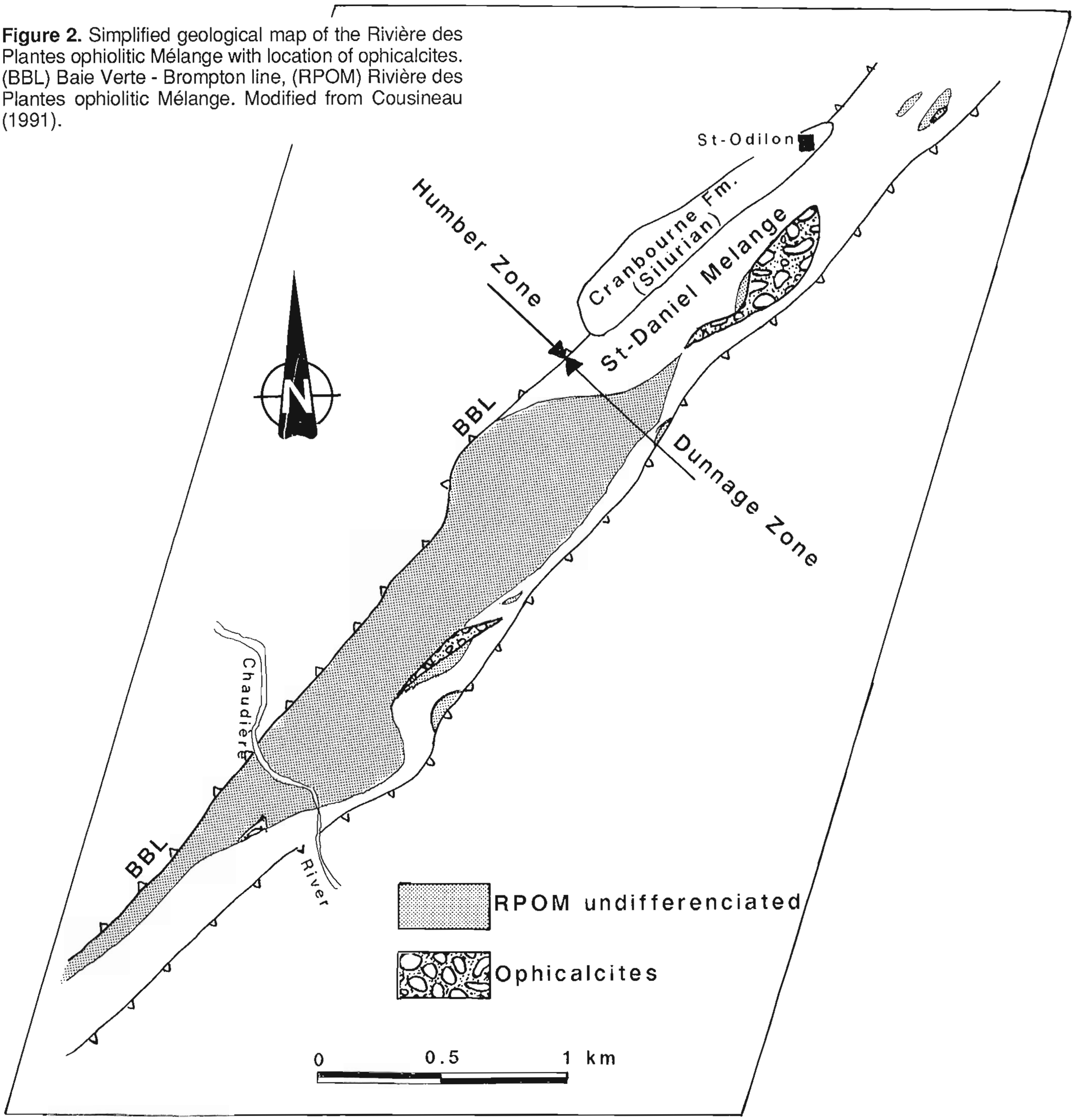


Figure 2. Simplified geological map of the Rivière des Plantes ophiolitic Mélange with location of ophicalcites. (BBL) Baie Verte - Brompton line, (RPOM) Rivière des Plantes ophiolitic Mélange. Modified from Cousineau (1991).

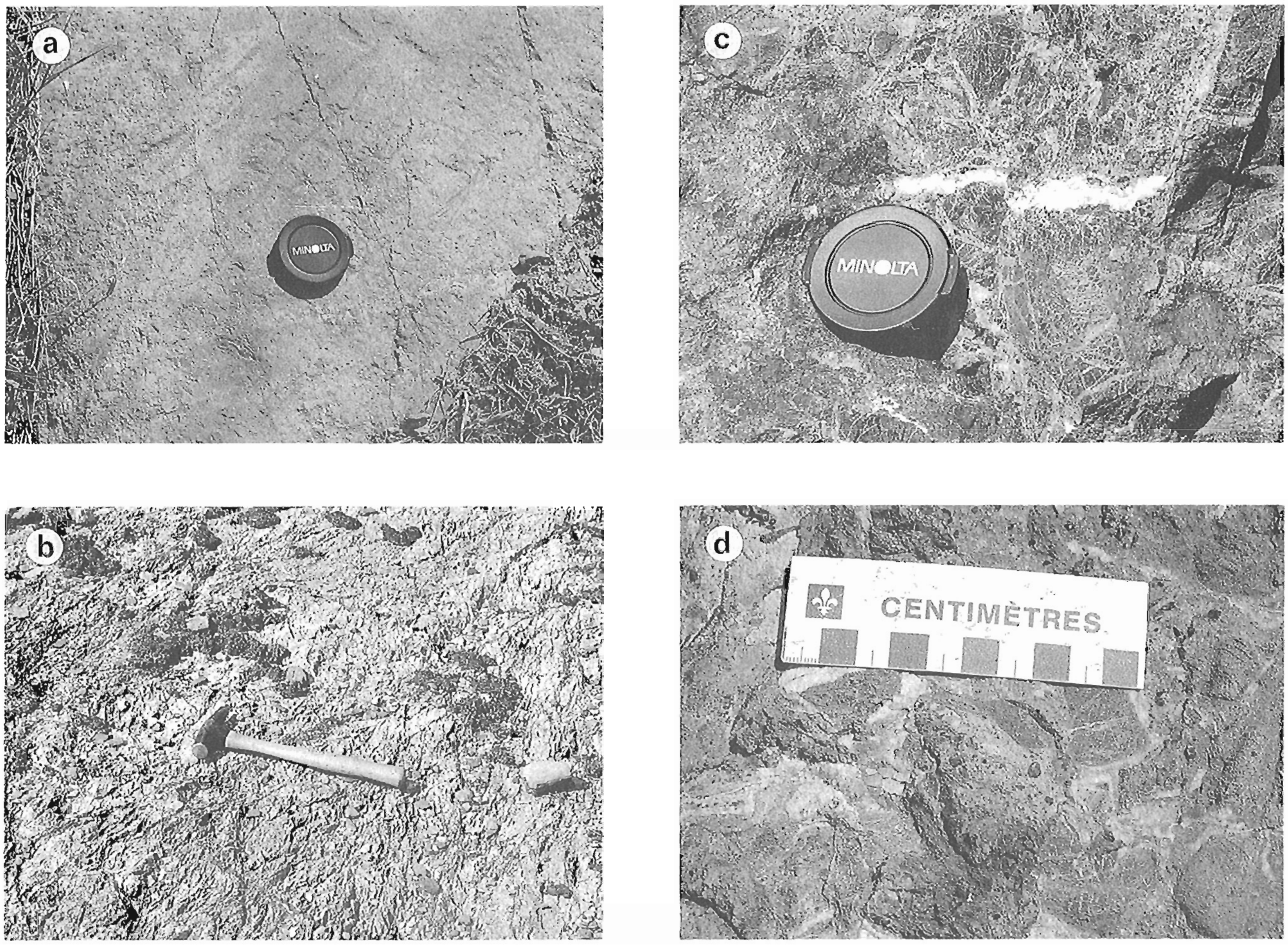


Figure 3. (**a**) Serpentinized and carbonate-free harzburgite of the Rivière des Plantes ophiolitic Mélange. Note the magmatic layering defined by pyroxenes. (**b**) Intensely sheared serpentinite devoid of any major carbonate phase. Hammer is 30 cm long. (**c**) Sheared serpentinite with some carbonate cements sealing fractures. Note the intense microfracturing of the material. (**d**) Typical ophicalcite of the Rivière des Plantes ophiolitic melange. The rock is now mostly composed of carbonates (micrite and calcite cements).

Figure 4. Ophicalcite block protruding into shaly rocks of the Saint-Daniel mélange (contact marked by the stippled line). The Saint-Daniel matrix has been plastically deformed by the intrusion of the ophicalcite block. Hammer is 30 cm long.

occurring phase. The sedimentary nature of the micrite is shown by various current structures such as parallel laminations (Fig. 5a) and few cross-stratifications. Some voids even show faint grading (Fig. 5b). Moreover, the micrite commonly form centimetre-sized geotropic infillings in the various pores of the ultramafic rocks (Fig. 5b).

Micrite deposition was followed by (or synchrone with?) a stage of fracturing and by a cement generation made of isopachous, millimetre- to centimetre-sized, layers of calcite that fill various secondary pores (Fig. 6a). However, the isopachous layers are scarce and as a whole, this second carbonate generation is volumetrically of minor importance, making up to a maximum of 10% of total carbonates.

Finally, the remaining void space is filled by the third generation of carbonates: blocky calcite (Fig. 6b) and coeval, less common, dolomite. These two cements are volumetrically important in large fractures and voids where total occlusion by the two previous carbonate generations was not completed.

All three pore fillings are commonly seen but their relative chronology is generally obscured by complex crosscutting patterns of the fractures. In cases, some fractures containing

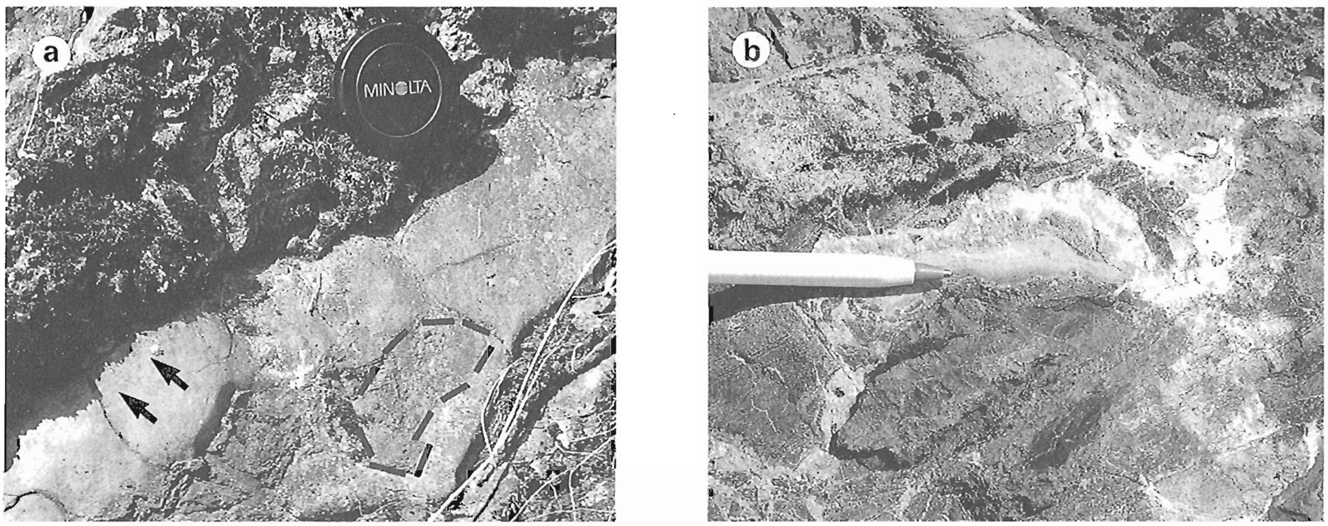


Figure 5. (**a**) Fracture in ophicalcite filled by grey micrite with diffuse parallel laminations (small arrows) and reddish micrite clast (inside stippled line). (**b**) Void in ophicalcite filled by faintly graded, geotropic micrite (pencil tip). The remaining of the void is sealed by blocky dolomite (greyish) and calcite (white).

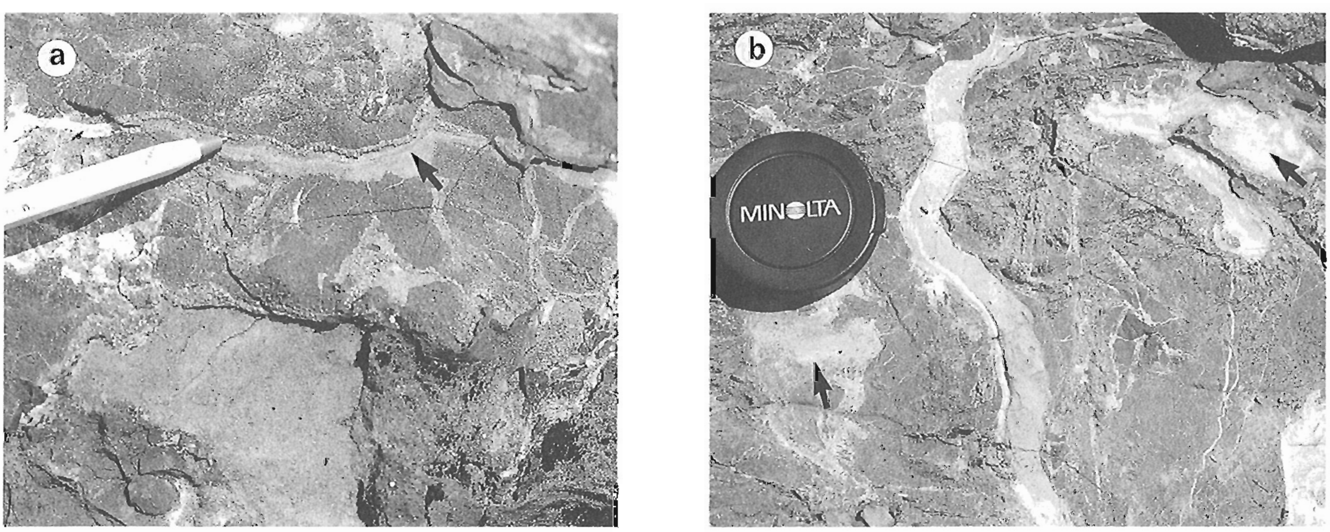


Figure 6. (**a**) Fracture filled by isopachous crust of calcite (pencil tip) overlying laminated micrite (small arrow). (**b**) Blocky calcite (arrows) completely occluding voids in ophicalcite. The adjacent fracture is filled by greyish micrite and white stubby calcite.

the complete spectrum of carbonates cut through older fractures completely occluded by blocky calcite (Fig. 7a). This suggest a complex history of fracturing, filling, and cementation.

An increase in carbonatization is invariably associated with an increase in frequency and size of the fractures. As a result, in the case of the most altered examples, the primary origin of the rock can only be deduced from presence of relicts of ultramafic material, ghost-like pseudomorphs (Fig. 7b), and disseminated chromite grains and serpentinite minerals.

PRELIMINARY CATHODOLUMINESCENCE PETROGRAPHY AND STABLE ISOTOPES GEOCHEMISTRY

In common diagenetic studies dealing with carbonates, cathodoluminescence (CL) helps describing different types and generations of cements even after neomorphism. In that perspective, precise identification of different generations of cement is of prime importance in any attempt to characterize the geochemistry of successive parental fluids. However, CL petrography is rarely reported in studies dealing with the origin of ophicalcites (Früh-Green et al., 1990).

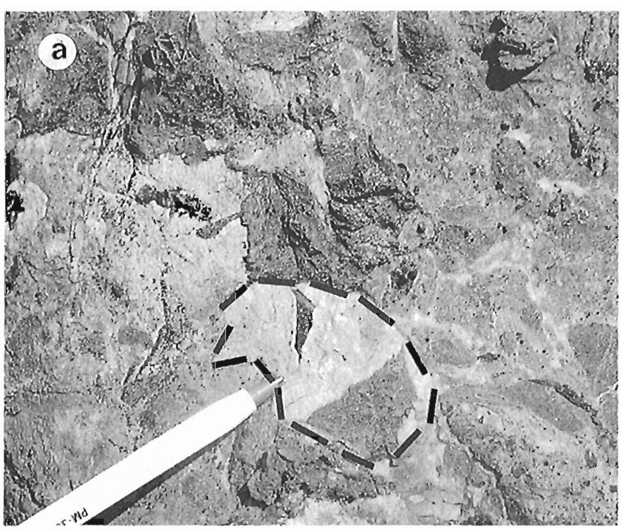
Prior to field work, preliminary CL petrography was carried out on a few selected samples, with emphasis on the late blocky calcite cement. Although it is seen as an uniform phase of non-ferroan, millimetre-sized crystals with conventional petrographic tools, it is composed of two distinct phases under CL.

The first phase is made of granular crystals with dull orange to brown luminescence (Fig. 8a). The subsequent phase is made of xenomorphic crystals displaying a bright yellow luminescence (Fig. 8b). The dull phase represents about 40% of the late cement but is not developed in all fractures. The volumetrically-dominant bright phase occludes the remaining porosity. Bright and dull calcites are both seen in microfractures cutting across pyroxene crystals (Fig. 8c, 8d). They constitute the last generation of cement in the studied samples.

Interestingly, the above petrographic observations agree with those of Fürh-Green et al. (1990) who also used CL in their study of alpine ophicalcites (eastern Switzerland Alps). These authors recognized a similar sequence, with dull to bright late fracture-filling cements cutting through early cements and internal sediment infills. Their late cement is also present in microfractured calcite and serpentinite material.

Without having any information on minor and trace elements for these calcites, it would be highly speculative to consider the cause of luminescence (Machel, 1985; Ten Have and Heijnen, 1985; Reeder and Grams, 1987; Reeder and Paquette, 1989). The interpretation of luminescence patterns will not be addressed here. However, in order to characterize the parental fluids, the two distinct CL zones can be sampled for oxygen and carbon stable isotopes.

Six samples, three for each CL zone, were analyzed at the Derry-Rust Laboratory of the Ottawa - Carleton Geoscience Centre. The precision of the data is ± 0.1 ‰ for both $\delta^{18}O$ and $\delta^{13}C$. The results are shown on Figure 9 where the two CL zones yielded distinct $\delta^{18}O$ values. The early, dull phase has an average $\delta^{18}O_{SMOW}$ value of 17.1‰ whereas the results for the late, bright phase clustered around 15.9‰. In contrast, $\delta^{13}C$ values for both cements have similar average values of -1.1‰ for the early phase and of -0.9‰ for the subsequent one. At this stage, it should be stressed that the data is not statistically significant yet; the profile could



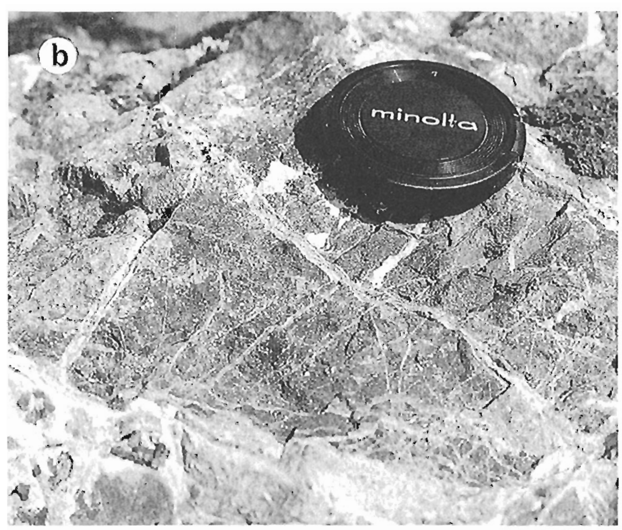


Figure 7. (**a**) Late void (inside stippled line) filled by geotropic micrite (dark grey) overlain by numerous generations of calcite cement. The void is cutting through an older structure filled by micrite. (**b**) Typical cross-cutting relationship of fractures in ophicalcite represented by intensely brecciated material with fractures healed by calcite cements.

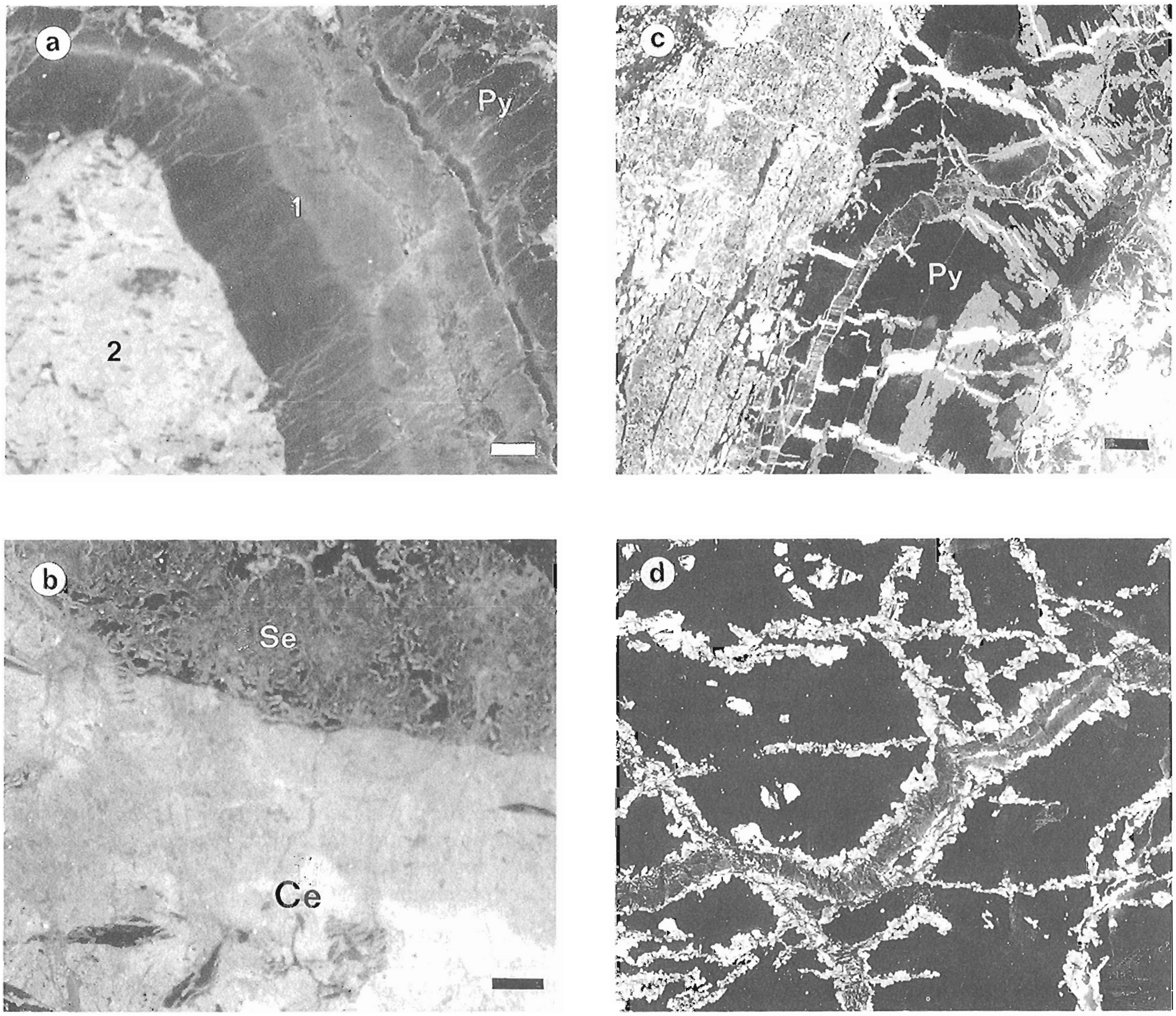


Figure 8. Cathodoluminescence petrography of late cements. (**a**) Large void bordered by a microfractured pyroxene crystal (Py), filled by an early phase of dull luminescent to increasingly darker first generation of cement (1), cut by microfractures filled with dull calcite. This generation is abruptly followed by a brightly luminescent xenomorphic spar (2). (**b**) Large fracture cutting through serpentinized material (Se), healed by brightly luminescent cement (Ce). (**c**) Non-luminescent pyroxene crystals (Py) cutted by numerous microfractures healed in sequence by dull and bright cements. (**d**) Complex microfracturing pattern in non-luminescent pyroxene crystals. Fractures are partly healed by bright cement and the remaining pores are occluded by dull calcite. All scale bars are 1 mm.

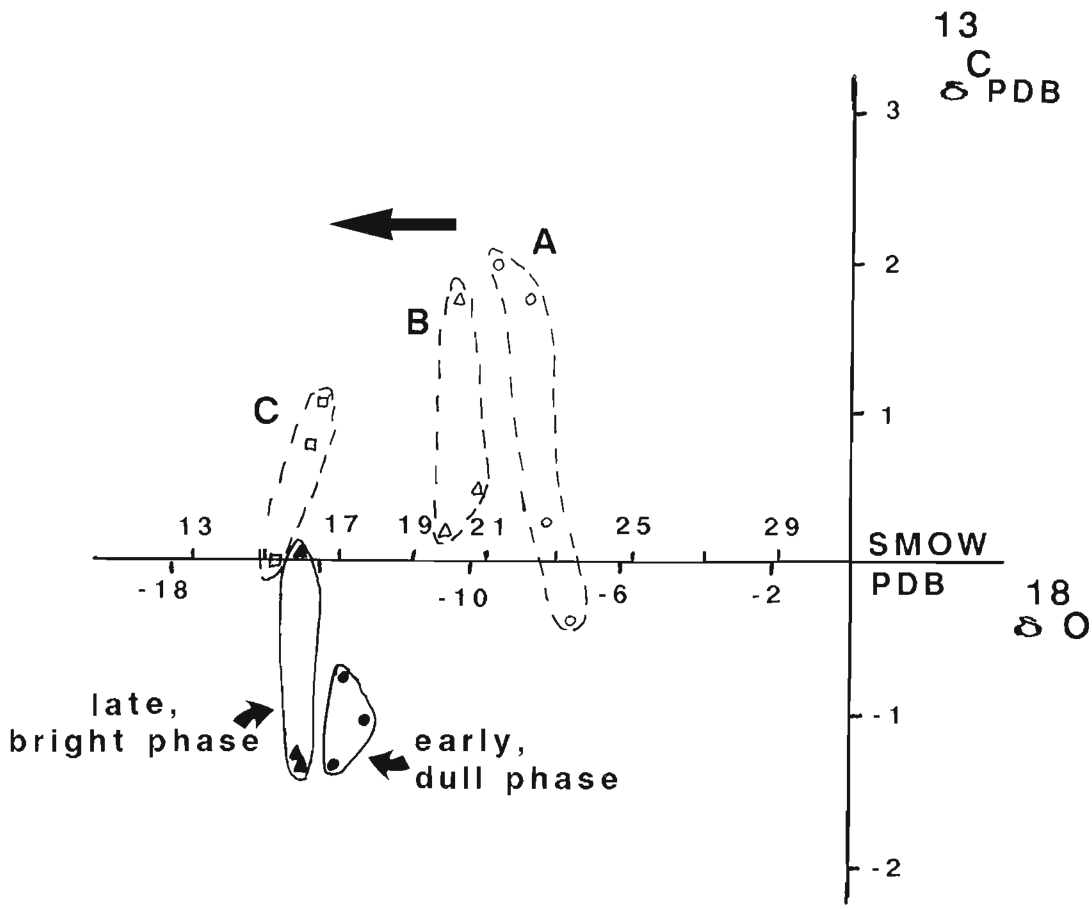


Figure 9. $\delta^{18}O$ and $\delta^{13}C$ crossplot of late, blocky calcite cements from the Rivière des Plantes ophiolitic melange (filled symbols), and data for late, fracture-filling cements (open symbols) from Früh-Green et al. (1990). Arrow indicate increasing metamorphism from prehnite-pumpellite to greenschist facies for three different localities: (A, circles) Parseen, (B, triangles) Arosa and (C, squares) Alp Flix.

change as more data is available. Figure 9 also shows the data of Früh-Green et al. (1990) for comparison (see following discussion).

DISCUSSION

Comparison with other published values for late fracture-filling calcite cements in ophicalcites

Whereas numerous studies report $\delta^{18}O$ and $\delta^{13}C$ values for internal sediment and mixed cement samples (Weissert and Bernouilli, 1984; Barbierei et al., 1979), only few other studies deal with separated phases (e.g. Früh-Green et al., 1990).

These authors partly based their understanding of carbonatization on CL petrography and carbon and oxygen isotopic results for various carbonate components. They recognized a clear relationship between the metamorphic grade of the rock and the oxygen isotopic ratios of the carbonate components. Such a direct relationship exists between metamorphic grade and oxygen isotopes in chert, and in associated pelagic limestones as well. The trend is that low-grade metamorphic samples, from prehnite-pumpellite facies, are less depleted in ^{18}O than samples of the greenschist facies, a slightly higher metamorphic grade. This, without systematic variation for carbon isotopes. Our preliminary results for samples from greenschist facies fall within the alpine population of ^{18}O-depleted samples (Fig. 9).

Before interpreting oxygen isotopic data of carbonate, it is necessary to determine the isotopic composition of marine water from which they likely originate, mostly because significant secular isotopic variations might have occurred in Phanerozoic oceans, as suggested by analysis of former marine cements and brachiopod shells (James and Choquette, 1990). The Saint-Daniel Mélange that host the RPOM is assumed to be of Early to Middle Ordovician age (Cousineau, 1988). However, the timing of carbonatization is not currently known. The upper age limit is constrained by the age of the post-Taconian successor basin fill, the Cranbourne Formation, a Late Silurian unit (Fig. 2). Carbonatization could only be older than this age, possibly a Middle Ordovician event.

The "best" marine components (Popp et al., 1986; Lohmann, 1988) of Middle Ordovician have $\delta^{18}O_{PDB}$ values clustering around -6.5‰ relative to PDB (24.2‰ for SMOW). Cements with highly depleted values of 17.1‰

(SMOW; early phase) and 15.9‰ (SMOW; late phase) could be explained in many ways. For one explanation, deep burial precipitation could account for such values. Assuming a sea water composition for the parent fluids, the necessary temperature of precipitation would be of 58°C for the early phase and of 65°C for the late one (O'Neil, et al., 1969). However, this scenario is unlikely due to the presence of synchronous mechanical sedimentation of micrite in voids and fractures. Precipitation from ^{18}O-depleted meteoric waters could not have resulted in such depleted values in the near surface environment unless the paleolatitude of the terrain was high. Alternatively ^{18}O re-equilibration of seawater-derived cements during later metamorphism, in a fluid-dominated system, could be responsible for the obtained signals. As mentionned earlier, Früh-Green et al. (1990) showed that an increase from prehnite-pumpellite to greenschist metamorphic facies is associated to a lowering of the $\delta^{18}O$ by approximately 6‰ in the carbonates. Finally, near seafloor precipitation from hot (hydrothermal) waters could also explain the ^{18}O-depleted values of the cements. This interpretation is compatible with the synchronous mechanical deposition of micrite. Knowing that modern sites of ophicalcite formation are characterized by hydrothermalism, Früh-Green et al. (1990) proposed this process in order to explain the isotopic signatures of their carbonate cements and serpentinite.

In a preliminary interpretation, it seems that hydrothermal waters and possibly later low-grade metamorphic re-equilibration could have produced the observed isotopic data for the late calcite cements. However, this scenario has still to be tested by further isotopic investigations (stable and radiogenic isotopes).

CONCLUSIONS

Field observations on ophicalcites occurring in the Rivière des Plantes ophiolitic Mélange together with preliminary petrographic and stable isotope works on selected samples yield the following preliminary conclusions:

1) Ophicalcites are the end-product after serpentinization, shearing, brecciation and fracturing of ultramafic rocks, the fractures being filled by various carbonate phases.

2) Three major carbonate generations are present in the ophicalcites as seen on outcrops: grey to reddish sedimented micrite, isopachous layers of calcite cement, and blocky calcite associated to less common dolomite cements.

3) The sedimentary and diagenetic infills of fractures is a complex history as multiple stages of fracturing, internal sedimentation, and cementation crosscut each other.

4) Preliminary cathodoluminescence petrography on the late blocky calcite showed an early dull zone followed by a bright one.

5) Oxygen stable isotopic signatures of both phases are extremely lighter than those of Ordovician marine carbonates, the late phase falling at the most depleted end. This trend is preliminary interpreted as a result of isotopic re-equilibration of hydrothermal precipitates by low-grade metamorphism.

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Sedimentological studies in the Ordovician Miramichi, Tetagouche, and Fournier groups in the Bathurst camp and the Belledune-Elmtree Inlier, northern New Brunswick[1]

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Abstract

The Miramichi Group may representsa submarine fan/fan complex with the psammite dominant Chain of Rocks Formation representing a channelized region, and the pelite dominant Knights Brook Formation a distal position relative to major distributaries. Alternatively the Chain of Rocks Formation may represent a drowned shelf.

In the Tetagouche Group the Vallée Lourdes Formation largely represents both a lower energy shelf slope and a higher energy shallow shelf. Psammite-pelite turbidites in the Patrick Brook Formation are interpreted as slope to base of slope deposits. The Nepisiguit Falls Formation contains little evidence to refine the marine volcanic interpretation. The Boucher Brook Formation contains thin turbidites with black graphitic pelites indicating a deep marine setting, likely a distal position on a submarine fan/fan complex..

The volcanilithic psammites and associated sediments in the Millstream and Pointe Verte formations of the Fournier Group are best interpreted to represent turbidity current deposition off marine volcanic edifices. Felsic volcanism was synchronous with the deposition of both formations.

Résumé

Le groupe de Miramichi est un complexe de cône sous-marin et de cône comprenant la formation surtout psammitique de Chain of Rocks, qui est représentative d'une zone à chenaux, et la formation surtout pélitique de Knights Brook, qui l'est d'une zone distale par rapport aux principaux défluents.

La formation de Vallée Lourdes du groupe de Tetagouche est en grande partie associée à deux milieux : l'un de talus de plate-forme (faible énergie) et l'autre de plate-forme d'eau peu profonde (forte énergie). Les turbidites psammitiques et pélitiques de la formation de Patrick Brook sont considérées comme des sédiments de la zone s'échelonnant du talus à la base du talus. La formation de Nepisiguit Falls présente peu d'indices qui apporteraient de nouveaux éléments à l'inteprétation d'un milieu marin volcanique. Quant à la formation de Boucher Brook, elle se compose de minces turbidites et de pélites graphitiques noires, deux lithologies typiques d'un marin profond correspondant sans doute à la partie distale d'un complexe de cône sous-marin et de cône.

La meilleure interprétation des psammites à fragments volcaniques et des roches sédimentaires connexes des formations de Millstream et de Pointe Verte (groupe de Fournier) indique que ces lithologies doivent leur mise en place en bordure d'édifices volcaniques marins à des courants de turbidité. Le volcanisme felsique est contemporain de la sédimentation de ces deux formations.

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INTRODUCTION

Recent mapping of the Miramichi Highlands has extended the Ordovician tectonostratigraphic zonation of the northern Appalachians from Newfoundland to northern New Brunswick. Accordingly, the Miramichi Group is considered equivalent to the Gander Zone, Tetagouche and Fournier groups are equivalent to the Dunnage Zone (van Staal and Fyffe, 1991). The Tetagouche Group is host to base metal sulphide deposits, iron-formation, and ferromanganiferous pelites in the Bathurst mining camp. Although the revisions to the Ordovician stratigraphy of van Staal and Fyffe (1991) can be applied successfully to the Bathurst camp, detailed sedimentological data have been lacking.

The present study is a sedimentological analysis of the Miramichi, Tetagouche, and Fournier groups. Future petrologic work will attempt to determine detrital compositions for provenance modelling that will test the rifted continental margin tectonic setting as determined by geochemistry of associated volcanic rocks (van Staal et al., 1991).

MIRAMICHI GROUP: DESCRIPTION

Chain of Rocks Formation

The Chain of Rocks Formation, recently defined in van Staal and Fyffe (1991), forms the structural base of the Miramichi Group (Fig. 2). The unit was examined at exposures along the Nepisiguit River (location 8, Fig. 1) and along the Middle River (location 6, Fig. 1). Fine (± coarse) grained, pale-grey weathering psammitic rocks are predominant; grey-green to light-grey pelitic rocks are local. Framework grains are mainly quartz; a labile component is apparently absent. Primary sedimentary structures are generally obscured by penetrative cleavage(s). Hence parallel lamination, normally graded bedding, sharp-based (with scour pockets) contacts of psammite over pelite, and pelite drapes above rippled upper psammite surfaces are only locally preserved. The psammites form decimetre- to metre-scale tabular layers separated by centimetre- to decimetre-scale pelite lenses (Fig. 3). Thicker psammite intervals are considered the result of erosional amalgamation. Locally, psammite horizons display a lensoid morphology.

Exposures along the Middle River indicate that the Chain of Rocks Formation is conformable with, and consistently younging towards, pelitic rocks of the overlying Knights Brook Formation. The contact is covered in an interval of discontinuous outcrop in which the number and thickness of pelitic beds increase upsection at the expense of psammitic beds. The contact with the Knights Brook Formation is not exposed along the Nepisiguit River.

Knights Brook Formation

Pelitic rocks, including black shales, of the Knights Brook Formation (newly defined in van Staal and Fyffe, 1991) were examined in four areas: 1) along Nepisiguit River (location 7, Fig. 1); 2) Knights Brook (location 11, Fig. 1); 3) along Tetagouche River immediately downstream of Little Falls (location 4, Fig. 1); and 4) along the Middle River (locations 5, 6, Fig. 1).

The pelites are characteristically dark grey to black, nonstratified and highly cleaved. Although pelitic rocks are predominant, green-grey or brown-grey, fine- to medium-grained metaquartzites form local interbeds (at the base of centimetre- to decimetre-scale fining-upward sequences) and lenses (Fig. 4). Subhedral to euhedral, volcanogenic quartz grains are a minor detrital component.

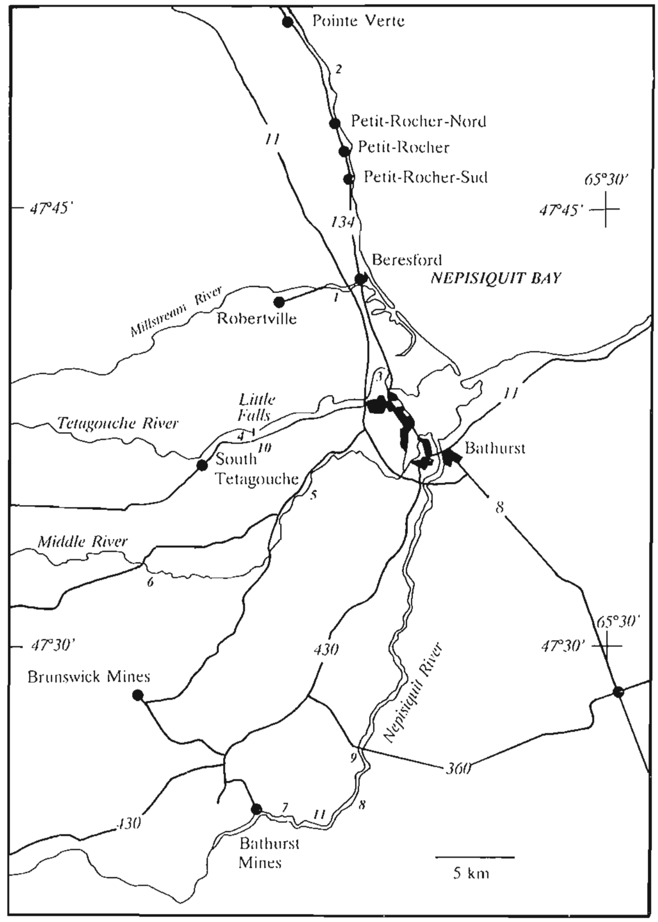

Figure 1. Location map of exposures of the Miramichi, Tetagouche, and Fournier groups examined in the Bathurst region of New Brunswick. **1** - Millstream Fm., Millstream River **2** - Pointe Verte Fm. coastal exposures **3** - Boucher Brk. Fm., Vallée Lourdes Fm., Patrick Brook Fm., Tetagouche River at Vallée Lourdes **4** - Knights Brook Fm., Vallée Lourdes Fm., Patrick Brook Fm., Nepisiguit Falls Fm., Tetagouche River at Little Falls **5** - Patrick Brook Fm., Knights Brook Fm., Middle River adjacent to Middle River Road **6** - Patrick Brook Fm., Knights Brook Fm., Chain of Rocks Fm., Middle River, several kilometres downstream of Rio Road bridge **7** - Nepisiguit Falls Fm., Knights Brook Fm., Nepisiguit River, several kilometres downstream of Bathurst Mines **8** - Chain of Rocks Fm., Nepisiguit River at Chain of Rocks Rapids **9** - Patrick Brook Fm., Nepisiguit River at Middle Landing Rapids **10** - Boucher Brook Fm., Tetagouche River near Tetagouche Falls **11** - Knights Brook Fm., Knights Brook

Owing to high strain, preservation of primary sedimentary structures is poor. Yet parallel stratification, normally graded bedding, and rippled upper psammite surfaces were observed locally.

At Little Falls on the Tetagouche River, the Knights Brook Formation is overlain by calcarenite and minor conglomerate of the Vallée Lourdes Formation (basal Tetagouche Group). A thin (roughly 50 cm) pebble conglomerate, containing siltstone clasts derived from the Miramichi Group, is distributed discontinuously along the contact. Van Staal et al. (1991) suggested that the unconformity implied by this conglomerate formed as a consequence of local lithosphwric doming immediately before back-arc volcanism in the middle to late Arenigian.

Along the Middle River, the Vallée Lourdes Formation is not developed, and the Knights Brook Formation is directly overlain by lithologically similar rocks of Patrick Brook Formation. The key lithological differences between the two units, as exposed in a discontinuous outcrop interval along the Middle River, are that Patrick Brook Formation

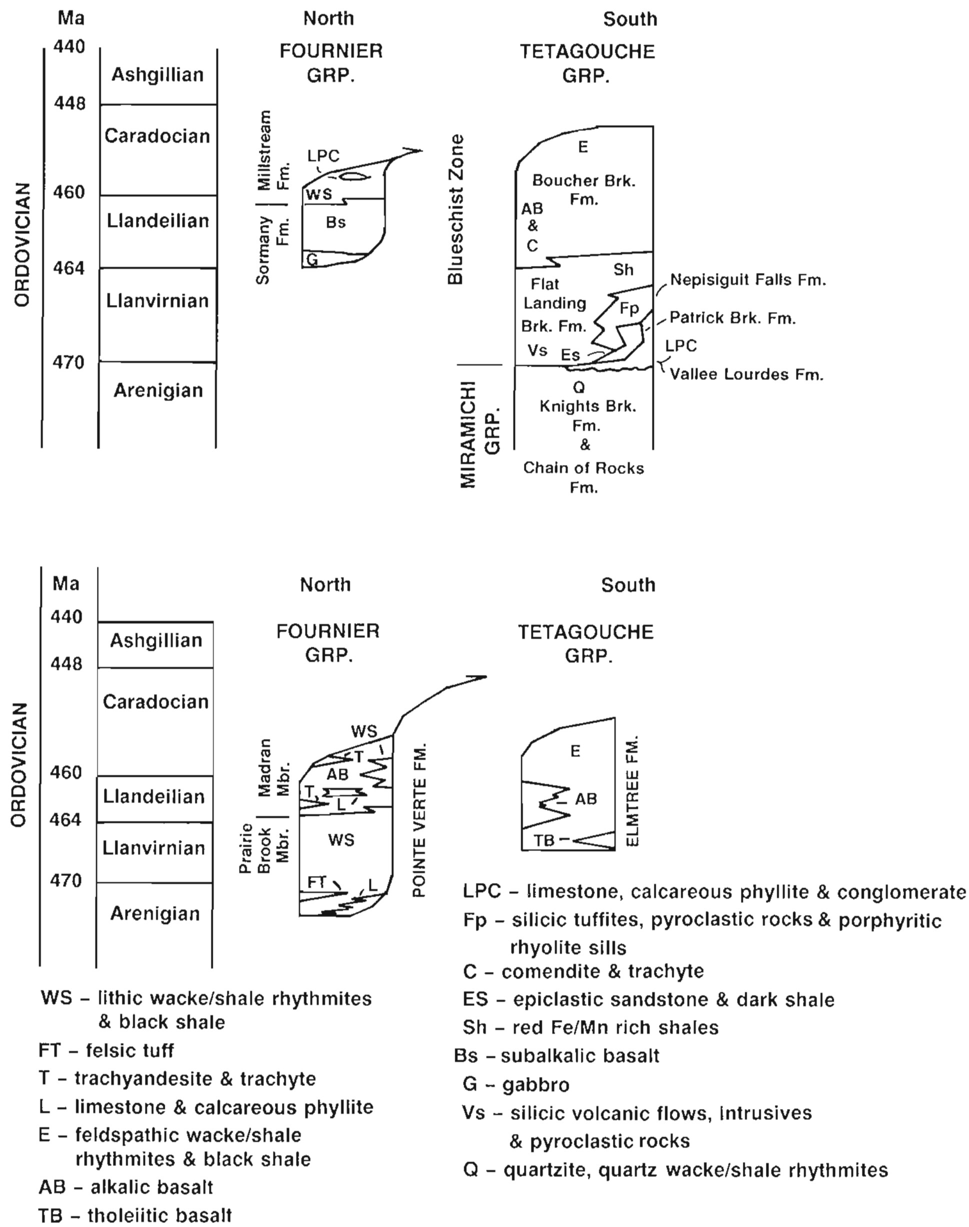

Figure 2. Schematic Ordovician tectonostratigraphy through two areas of the northern Miramichi Highlands. Lower diagram is north to south through the Elmtree Inlier. Upper diagram is north to south through the region west-northwest of Bathurst south of the Rocky Brook – Millstream Fault. Simplified in part from Figure 2 of van Staal et al. (1991).

psammites tend to: 1) have a greater proportion of volcanogenic smoky quartz grains (typically >5%); 2) be more feldspathic; and 3) weather a darker color (dark grey to black).

MIRAMICHI GROUP: INTERPRETATION

The psammite to pelite fining-upward sequences of both the Chain of Rocks (thick sequences) and Knights Brook formations (thin sequences) resemble partial Bouma sequences deposited by waning turbidity current flows. Although there is no direct evidence for marine (rather than lacustrine) standing waters, graptolites have been reported by Fyffe et al. (1983) from turbidites interpreted to be correlative to the Knights Brook Formation (Bright Eye Brook Formation, west-central New Brunswick; van Staal and Fyffe, 1991).

The vertical transition from psammite predominant (Chain of Rocks Formation) to pelite predominant (Knights Brook Formation) turbidites may represent a facies change from a channelized fan region to an area of principally overbank deposition in a deep-sea turbidite fan model. Alternatively, the transition may represent a facies change due to drowning of a siliciclastic shelf represented by the Chain of Rocks Formation (cf. O'Neill and Knight, 1988).

TETAGOUCHE GROUP: DESCRIPTION

The lower part of the Tetagouche Group is a metavolcanic and metasedimentary assemblage in which interfingering of the constituent units (Vallée Lourdes, Patrick Brook, Nepisiguit Falls, and Flat Landing Brook formations) signifies lateral facies equivalence (van Staal and Fyffe, 1991; Fig. 2). The upper part of the group (Boucher Brook Formation) is a predominantly metasedimentary assemblage, bearing only minor metavolcanic rocks.

Vallée Lourdes Formation

The Vallée Lourdes Formation forms lenses close to the contact between the Miramichi and Tetagouche groups extending for 15 km to the southwest of Bathurst (van Staal and Fyffe, 1991). The unit was examined at two locations on the Tetagouche River (locations 3, 4, Fig. 1). In the Vallée Lourdes area on the northern outskirts of Bathurst (location 3), the unit consists of green-grey cleaved calcareous pelite and very fine- to fine-grained calcareous psammite arranged either in fining-upward sequences 2-20 cm thick or as isolated units. Sedimentary structures include: faint stratification, ripple cross-lamination, normal grading, ball and pillow, and pelite rip-ups. Conglomerate containing randomly oriented clasts of carbonate, green-grey subangular pelite, grey rounded psammite and granitic gneiss (to 20 cm) is exposed at a single outcrop.

Superb exposures at Little Falls (location 4, Fig. 1) display a thin (several tens of centimetres) basal conglomerate containing siltstone clasts from the underlying Miramichi Group that is overlain principally by intervals of grey, medium- to coarse-grained calcarenite and interstratified calcarenite and silty calcarenite. Figure 5 presents a stratigraphic section through the Vallée Lourdes Formation at this location. The calcarenite is commonly trough

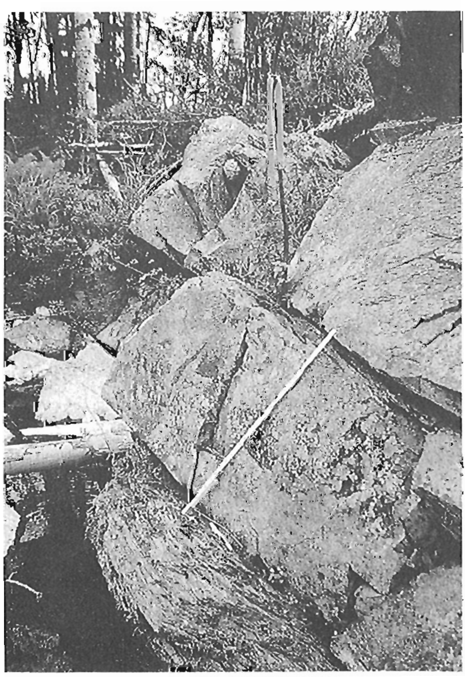

Figure 3. Thick psammites separated by thin pelite horizons in the Chain of Rocks Formation at Chain of Rocks Rapids on the Nepisiguit River.

Figure 4. Interstratified psammites and pelites in the Knights Brook Formation on the Middle River several kilometres south of Rio Road Bridge. Irregular dark grey patches are pools of water.

cross-stratified in 5-20 cm sets that indicate predominantly westward paleoflow and contains pockets of grey silty calcarenite clasts as lags at the base of, or dispersed throughout, individual units. Coarse shell fragments are locally visible and conodonts indicate a late Arenigian-early Llanvirnian age (Nowlan, 1981). Middle-late Arenigian brachiopods (Fyffe, 1976; Neuman, 1984) have been recovered from the formation on the Tetagouche River west of Tetagouche Falls (van Staal and Fyffe, 1991).

In the Vallée Lourdes area, the formation is underlain by the Patrick Brook Formation. The contact is in a covered interval between outcrops of green-grey interstratified psammites and pelites of the Vallée Lourdes Formation and dark grey interstratified psammites and pelites of the Patrick Brook Formation. Patrick Brook Formation psammites are coarser grained and contain volcanogenic quartz phenoclasts. At Little Falls on the Tetagouche River, roughly 11 km west of Bathurst, the Patrick Brook Formation overlies the Vallée Lourdes Formation and the contact is a high strain zone.

Patrick Brook Formation

The Patrick Brook Formation was examined along the Tetagouche River (location 3, Fig. 1), at Middle Landing Rapids on the Nepisiguit River (location 9, Fig. 1), and at two locations on the Middle River (locations 5, 6, Fig. 1). The unit consists of pelite and interstratified pelite and psammite. Pelites are grey to dark grey, highly cleaved, and rarely display original stratification. Psammites are characteristically dark grey to black, very fine to medium-grained, and have both quartz and feldspar as framework grains, commonly with >5% dark volcanogenic smoky quartz phenoclasts. Individual units are commonly nonstratified to faintly laminated and display only local normal grading or occasional ripple cross-lamination. The two rock types define fining-upward sequences 5 cm to several metres thick; psammite units locally form lenses over several metres wide.

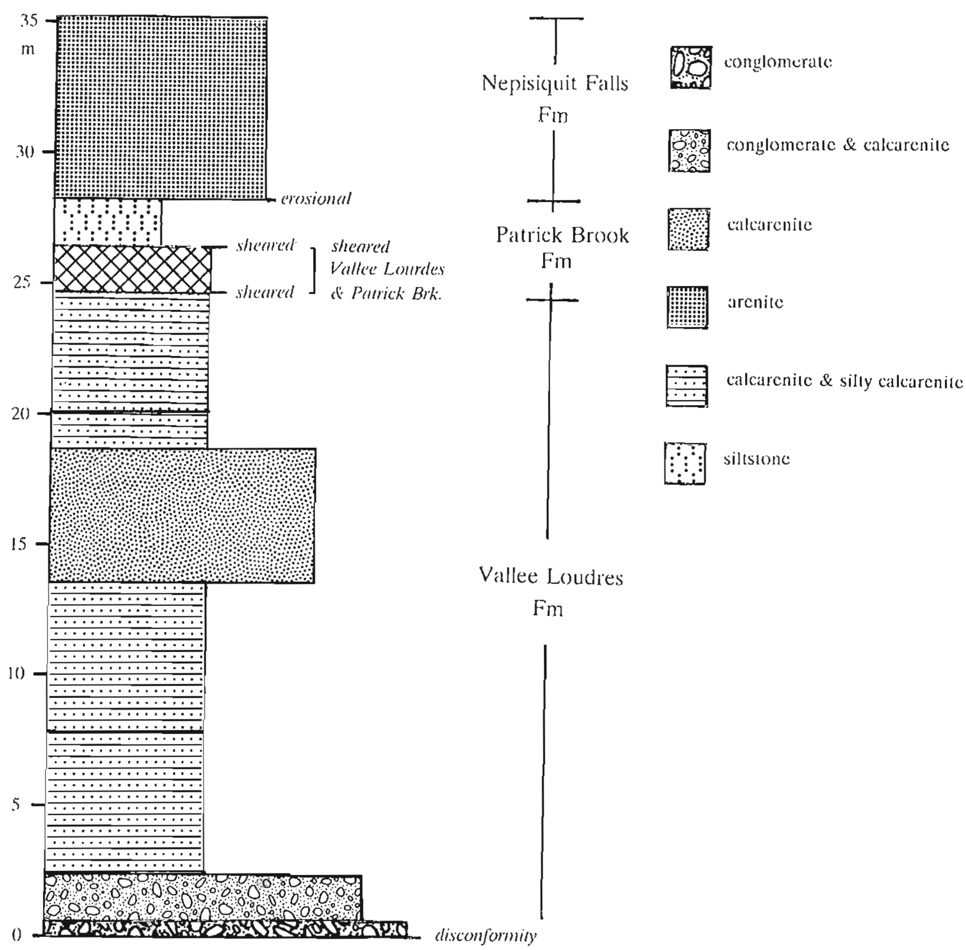

Figure 5. Stratigraphic section through the Vallée Lourdes, Patrick Brook, and Nepisiguit Falls formations at Little Falls, Tetagouche River.

Grey siltstone clasts (to 20 cm), presumably derived from the Patrick Brook Formation, are within the basal 30 cm of the overlying Nepisiguit Falls Formation at Little Falls. Hence the Patrick Brook-Nepisiguit Falls contact is locally considered to be depositional and erosional.

Nepisiguit Falls Formation

Felsic metavolcanic rocks of the Nepisiguit Falls Formation, were examined at the Nepisiguit Falls power dam at Bathurst Mines on the Nepisiguit River and at Little Falls on the Tetagouche River (locations 4, 7, Fig. 1). The unit consists of epiclastic and/or pyroclastic rocks interstratified with flows and/or sills (van Staal and Fyffe, 1991). At the Nepisiguit Falls power dam, the unit is represented by brown-grey, quartz/feldspar-augen schist or quartz-augen schist. The augen have been interpreted as relic volcanogenic phenoclasts (van Staal and Williams, 1984) modified during metamorphism and deformation. Local size grading is interpreted to reflect original normal grading of phenoclasts, and indicates a westward younging direction.

At Little Falls the degree of deformation is less. Here the formation consists of grey, medium-grained psammite. It is feldspathic only within the first two metres of the contact with the Patrick Brook Formation. Above this point feldspar is absent and volcanogenic quartz phenoclasts constitute part of the quartz-rich framework. The psammite is largely nonstratified to locally laminated or very thinly bedded. Other primary sedimentary features are restricted to local normally graded units indicating a north facing direction.

The upper contact of the Nepisiguit Falls Formation with the Boucher Brook Formation is covered at Little Falls; the upper contact was not examined at Bathurst Mines. Ooid and stromatolite-like structures in carbonate facies iron-formation associated with base-metal sulphides are exposed along the boundary between the Nepisiguit Falls and Flat Landing Brook formations (van Staal and Fyffe, 1991).

Boucher Brook Formation

The Boucher Brook Formation is a predominantly metasedimentary package (with minor metavolcanic rocks) that constitute the upper portion of the Tetagouche Group (van Staal and Fyffe, 1991). The formation was examined along the Tetagouche River in the Vallée Lourdes area and near Tetagouche Falls (locations 3, 10, Fig. 1). In the Vallée Lourdes area it consists mainly of pelites interstratified with minor psammite. The pelites are locally red to green and highly siliceous (cherty), but are more commonly dark grey to black (graphitic) and with only local green cherty interbeds. The contact between the red/green cherty pelites and the black pelites is sharp and appears to be conformable in a river-bottom outcrop exposed at low water near the Canadian National Railway bridge over the Tetagouche River. Thin nonstratified psammite beds are common in some black pelite intervals, defining the base of 5-6 cm thick fining-upward sequences (Fig. 6). Similar fining-upward sequences were observed near Tetagouche Falls. Graptolites of Late Caradocian age (clingani zone) have been recovered from the black pelites, at three separate localities, one of which is the Vallée Lourdes area (van Staal et al., 1988; Riva and Malo, 1988).

The upper contact of the Boucher Brook Formation was not observed. The black pelites serve as mechanically weak horizons exploited during imbrication of the Tetagouche Group (van Staal et al., 1990).

TETAGOUCHE GROUP: INTERPRETATION

Sedimentological and stratigraphic characteristics of the Tetagouche Group are consistent with a shelf and shelf slope (Vallée Lourdes) and slope to base of slope (Patrick Brook) depositional settings for the basal group and a deeper starved basin setting for the upper group (Boucher Brook Formation). The depositional setting of the Nepisiguit Falls Formation is more problematic. The available sedimentary evidence is compatible with a shallow shelf interpretation for at least part of the formation, although a deeper marine setting cannot be ruled out. The definition of interfingering relationships for basal Tetagouche Group formations (van Staal and Fyffe, 1991) has proven important in the determination of depositional setting for units lacking diagnostic sedimentological features. For instance we have observed that the Patrick Brook Formation is absent to poorly represented where the Nepisiguit Falls Formation is well developed. This suggests that the Patrick Brook Formation is a distal facies equivalent to the Nepisiguit Falls Formation.

The basal conglomerate of the Vallée Lourdes Formation may represent fluvial sedimentation and record a brief interval of subaereal exposure. Upsection the unit bears

Figure 6. Thin AE/ACE turbidites in the Boucher Brook Formation on the Tetagouche River near Vallée Lourdes.

shallow marine fossils, and erosionally amalgamated lenses of trough cross-stratified calcarenite with pelite rip-up clast lags suggest a energetic shallow shelf setting. A change in abundance of trough cross strata, differences in composition, and changes in style of stratification suggest variable energy conditions which are attributed to relative sea level changes. Along the Tetagouche River near Vallée Lourdes psammite-pelite couplets and a conglomerate unit suggest a lower energy, shelf slope, turbidite-debris flow setting. Psammite-pelite fining-upward sequences in the Patrick Brook Formation probably indicate turbidite sedimentation in a quieter water slope to base of slope setting seaward of the Vallée Lourdes shelf. It is inferred to be marine on the basis of lateral equivalence to marine fossil-bearing rocks of the Vallée Lourdes Formation. The Nepisiguit Falls Formation is assumed to represent largely felsic marine volcanism because the ooid and stromatolite-like structures and manganiferous pelites at the boundary between the Nepisiguit Falls and Flat Landing Brook formations are marine in aspect.

Graptolite-bearing black pelites and psammite to pelite fining-upward sequences in the Boucher Brook Formation signify a relatively starved deep marine setting interrupted by periodic turbidity currents.

FOURNIER GROUP: DESCRIPTION

Fournier Group rocks in the northern Miramichi Highlands are separated northward from lower Silurian rocks by the Rocky Brook-Millstream Fault Zone (RBMFZ). A thin, laterally extensive unit of blueschist-grade metamorphic rocks, phyllonite, and melange (interpreted as a suture) separates the unit from the Tetagouche Group (van Staal et al., 1990). South of the RBMFZ, the Fournier Group consists of two units: basalts of the Sormany Formation, and feldspathic psammites with relatively minor conglomerate, limestone, cherty pelite, and felsic volcanic rocks of the Millstream Formation (Young, 1911; van Staal et al., 1990; and van Staal and Fyffe, 1991).

In the Belledune-Elmtree Inlier north of the RBMFZ, rocks of the Fournier Group are represented by the Pointe Verte and Devereaux formations (Young, 1911; Pajari et al., 1977; and Rast and Stringer, 1980). These divisions were redefined by van Staal and Fyffe (1991) who suggested that they form the upper part of an ophiolite complex on the basis of basalt geochemistry and structural data. Langton and van Staal (1989) defined two members in the Pointe Verte Formation: a lower Prairie Brook Member consisting mainly of feldspathic and lithic psammites, with minor metalimestone, pelite, conglomerate, and felsic pyroclastic rocks, and an upper Madran Member composed principally of alkalic basalts with minor interstratified metasediments.

Millstream Formation

The Millstream Formation was examined along the Millstream River (location 1, Fig. 1). Pelitic units are highly cleaved, grey to dark grey or black, and form units up to several metres thick of pelite, pebbly pelite, and/or interstratified psammite and pelite in thin fining-upward sequences that give a consistent north-younging direction. Psammite units are greenish- to reddish-grey or grey, and very fine- to coarse-grained. Quartz is the predominant framework component, but volcanogenic quartz and feldspar, and volcanogenic rock fragments are locally abundant. Psammites form units up to several metres thick lacking primary sedimentary features. Locally, felsic volcanic flows contain chaotic mixtures of pelitic and psammitic sediment (peperite). Pebble conglomerate units containing either igneous clasts (van Staal and Fyffe, 1991) or angular pelite clasts, are also locally developed.

The Millstream Formation is thought to be Llandeilian to Caradocian based on fossils and U-Pb zircon geochronology of lithologically equivalent rocks in the Belledune-Elmtree Inlier (see below; van Staal and Fyffe, 1991).

Pointe Verte Formation

Only the psammitic metasediments of the Prairie Brook Member, exposed along the coast between Petit-Rocher-Nord and Pointe Verte (location 2, Fig. 1) were examined during the present study. They are brown-grey to dark-grey, coarse- to very coarse-grained, and contain significant percentages of volcanogenic quartz phenoclasts, feldspar, and felsic volcanic lithic fragments. They occur principally as nonstratified units and commonly display normal graded bedding, pelitic rip-up clasts, and up to 1 m wide normally graded lenses of granule-grade detritus 10-20 cm thick. A north facing direction is consistently indicated. Laminated sandy chert/cherty psammite horizons form local interbeds 10-20 cm thick. Minor interstratified pelites are highly cleaved, dark grey and nonstratified. Psammites and pelites display two styles of interstratification: 1) 10-40 cm thick psammites with 10-20 cm thick pelites in fining-upward sequences, and 2) 1-5 m thick psammites separated by local 5-10 cm pelite horizons. Fossils recovered from the Pointe Verte Formation indicate a Middle Ordovician age range (van Staal and Fyffe, 1991).

FOURNIER GROUP: INTERPRETATION

The Millstream Formation of the northern Miramichi Highlands and the Pointe Verte Formation of the Belledune-Elmtree Inlier are lithologically comparable (van Staal et al., 1990) and are considered to have been deposited in a similar depositional environment. Bouma-like fining-upward sequences with abundant volcanic detritus and interbeds of intermediate to felsic volcanic flows with digested sediment suggest a turbidite fan complex adjacent to, or developed on, coeval volcanic edifices. Cherty beds in the Pointe Verte Formation may represent exhalites. Variation in scale of fining-upward sequences is attributed to variation in proximity to fan distributaries. Although fossil evidence corroborating a marine setting is missing from the Millstream Formation, the Pointe Verte Formation contains conodonts and graptolites, and the basalts display an oceanic island geochemical signature (van Staal et al., 1991).

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Geological Survey of Canada Project 900038

An alteration and sulphur isotope study of the Pilley's Island massive sulphides, central Newfoundland[1]

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Santaguida, F., Hannington, M., and Jowett, C., 1992: An alteration and sulphur isotope study of the Pilley's Island massive sulphides, central Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 92-1D, p. 265-274.

Abstract

Massive sulphides on Pilley's Island occur within a 200 m thick, shallow-dipping sequence of altered felsic volcanics. Mineralization in the Old Pilley's Island Mine consists of massive to semi-massive pyrite and chalcopyrite with minor sphalerite. Sulphur isotope ratios range from 2.3 to 6.5 per mil. A local exposure of transported sulphides is more polymetallic and possesses sulphur isotope ratios from 1.1 to 1.8 per mil. This material is chemically and isotopically distinct from sulphides in the Old Mine and likely represents debris from the recently discovered 3B Zone. The sulphide horizons are enclosed by a zoned alteration envelope of stockwork-associated chlorite with widespread quartz-sericite ± K-feldspar. Several stages of chlorite are recognized and mineralization is most closely associated with a distinctive, Fe-rich end-member. Application of a chlorite geothermometer indicates thermally intensifying hydrothermal activity from 240°C to 360°C. Late-stage, Mg-rich chlorite appears to correlate with an influx of cold seawater during the collapse of the hydrothermal system.

Résumé

Les sulfures massifs dans l'île Pilley reposent au sein d'une séquence légèrement inclinée de 200 m d'épaisseur composée de roches volcaniques felsiques altérées. La minéralisation dans l'ancienne mine de l'île Pilley est composée de pyrite et de chalcopyrite de nature massive à semi-massive, avec un peu de sphalérite. Les rapports isotopiques du soufre varient entre 2,3 et 6,5 par mil. Un affleurement local de sulfures transportés est plus polymétallique et présente des rapports isotopiques du soufre variant entre 1,1 et 1,8 par mil. Ce matériau se différencie chimiquement et isotopiquement des sulfures de l'ancienne mine et correspond vraisemblablement à des débris de la zone 3B récemment découverte. Les horizons sulfurés sont entourés par une enveloppe d'altération zonée de chlorite associée à un stockwerk accompagné de quantités abondantes de quartz et séricite±feldspath potassique. Plusieurs stades de chlorite ont été relevés et la minéralisation serait étroitement associée à un membre terminal ferrifère distinct. L'application d'un géothermomètre dans la chlorite révèle une activité hydrothermale à température croissante, passant de 240 °C à 360 °C. La chlorite manganésifère de stade tardif semble correspondre à un apport d'eau de mer froide durant l'effondrement du système hydrothermal.

[1] Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development, 1990-94

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INTRODUCTION

The Robert's Arm-Buchans belt in Central Newfoundland is best known for the Buchans orebodies; total production 16 196 876 tonnes averaging 14.5% Zn, 7.6% Pb, and 1.3% Cu over 57 years (Kirkham, 1987). Mineralization throughout the belt has been compared to the classic Kuroko deposits of Japan (e.g., Swanson et al., 1981). Pilley's Island, at the northern end of the Robert's Arm Group (Fig. 1), hosts a number of small massive sulphide deposits which have attracted extensive exploration for more than one hundred years (Bostock, 1988; Tuach, 1988, 1990). Approximately

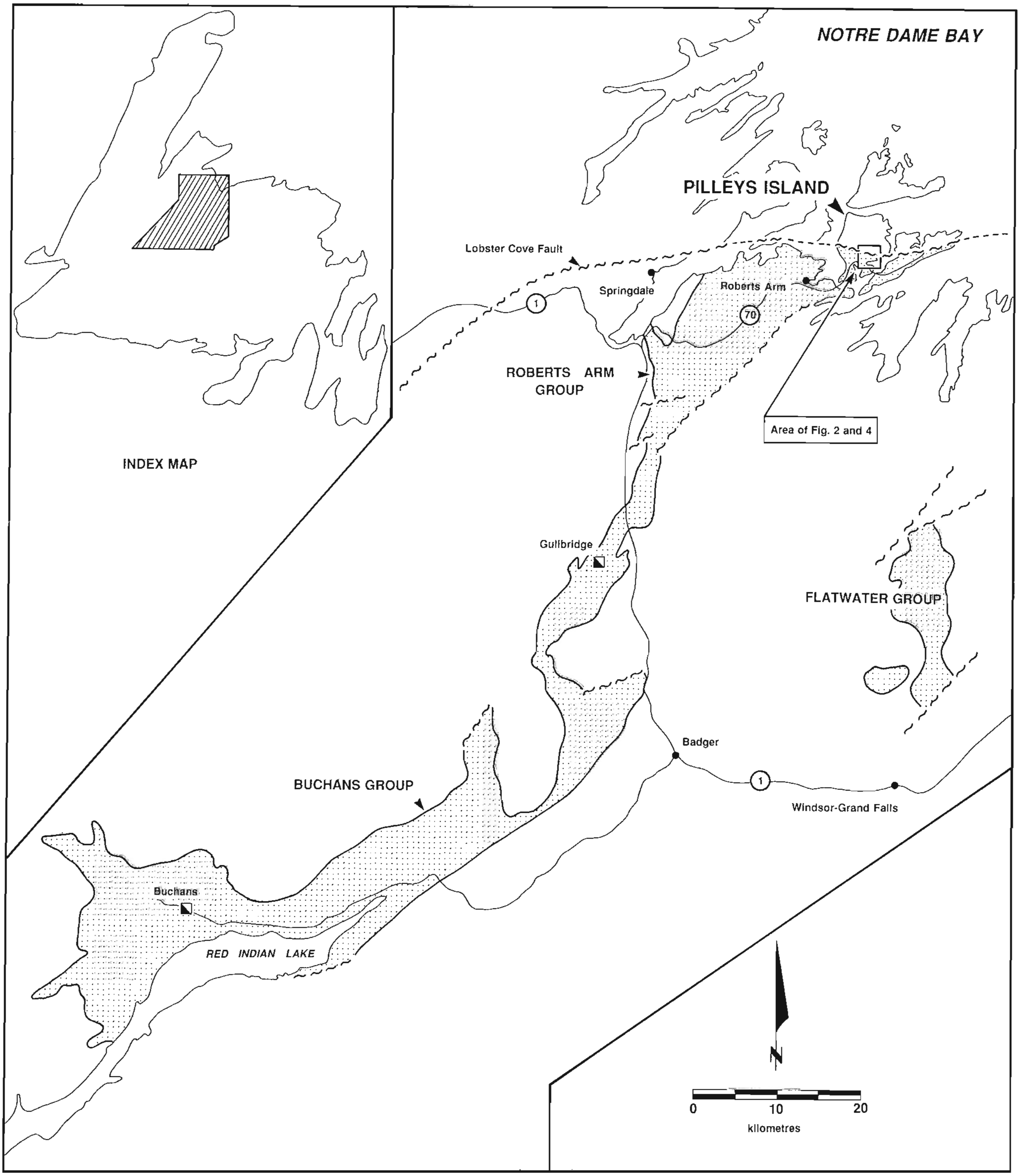

Figure 1. Location map of the Roberts Arm-Buchans Belt (stippled areas) (after Kean et al., 1981).

450 000 tonnes of massive pyritic ore was produced from the Old Pilley's Island Mine between 1891 and 1908, and estimated reserves are 1 M tonnes grading 1.2% Cu (Tuach, 1988; 1990). Exploration drilling was carried out on the property in 1951-56, 1966-68, and 1970, and a new sulphide zone (3B Zone) was located east of the Old Mine during drilling by Brinco Ltd. in 1983-84 (P. Hum and W. Epp, unpub. reports).

The altered felsic volcanic rocks which host the massive sulphides are among the best exposed in Newfoundland, and considerable work has been carried out on the local geology of the deposits by Epenshade (1937), Bostock (1978, 1988), and Tuach (1988, 1990). Strong (1973) described the igneous geochemistry of the felsic and mafic volcanics on Pilley's Island, and Appleyard and Bowles (1978) outlined the alteration present in the Old Pilley's Island Mine. However, systematic studies of the mineralization and alteration mineralogy in the Old Mine and the new 3B Zone have not been pursued. This paper outlines some of the major lithologies and physical volcanology associated with mineralization and describes the preliminary results of work on the alteration mineralogy and sulphur isotope composition of the sulphides. The locations of sulphide showings and drill holes examined in this study are shown in Figure 2.

LOCAL GEOLOGY

The Robert's Arm-Buchans belt is located within the Central Volcanic Belt of Newfoundland and has been interpreted as an allochthonous island arc terrain accreted during the Taconian orogeny (Swinden et al., 1988). Radiometric ages (U/Pb in zircon) for the Buchans and Robert's Arm Groups yield a common age of approximately 473 Ma (Dunning et al., 1987). On Pilley's Island, the Robert's Arm Group is exposed south of the Lobster Cove Fault (Fig. 1), a major lineament which separates the Robert's Arm and Lushs Bight groups. Felsic volcanics at the south end of Pilley's Island belong to the upper part of the Robert's Arm Group, the Boot Harbour terrain (Bostock, 1988), and consist of a sequence of dacite tuff, flows, and volcanic breccia. Regionally extensive mafic pillow lava and breccias comprise the stratigraphic footwall to the felsic volcanics. On Pilley's Island, this sequence has been folded about an east-trending, subhorizontal axis and sulphide mineralization now occupies the core of a northward overturned syncline (Bostock, 1978, 1988). Regional metamorphism is subgreenschist, and little penetrative deformation has affected the primary features of the volcaniclastic units. A number of faults cut the felsic volcanic rocks and have locally juxtaposed footwall pillow

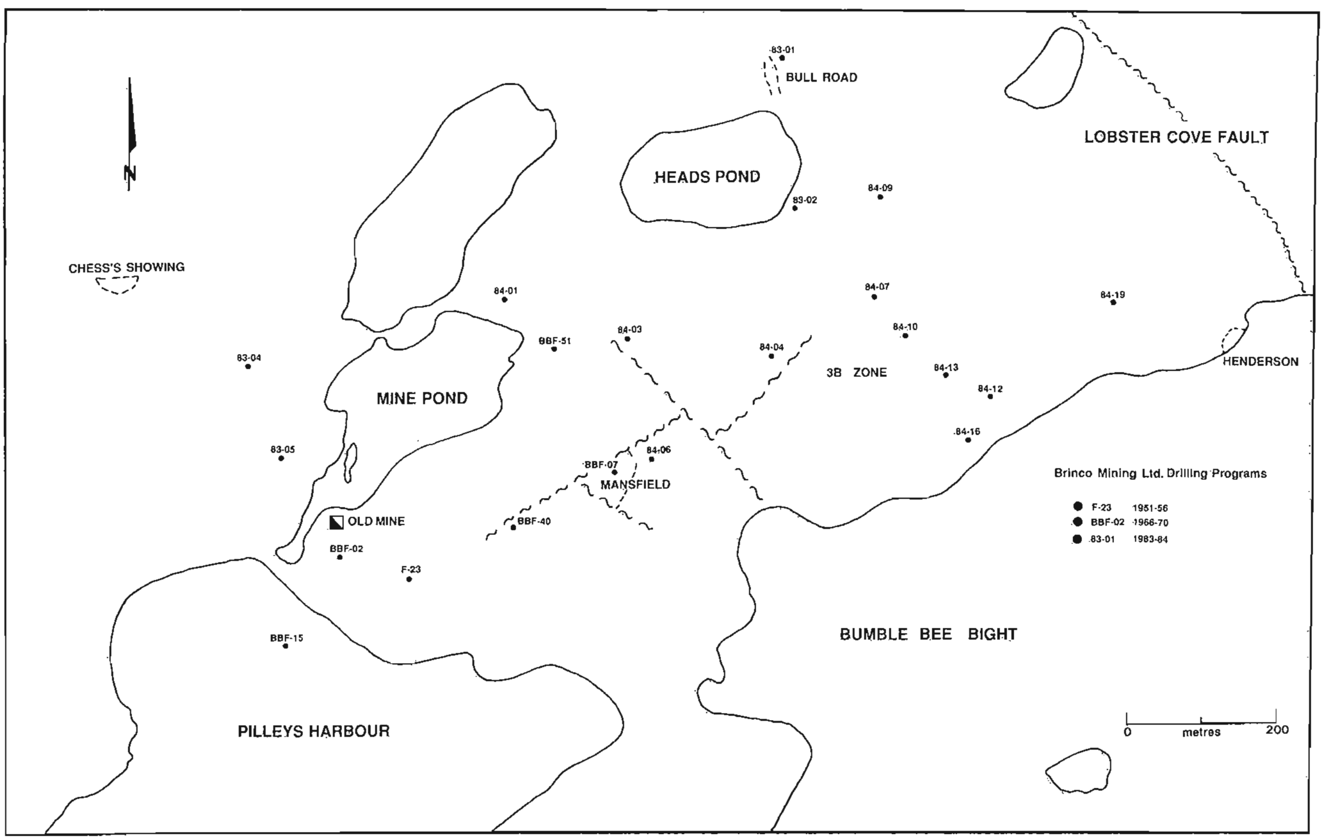

Figure 2. Location of drill holes examined in this study. The Old Pilley's Island Mine is located at the south end of Mine Pond (90 m below surface). The new 3B Zone is located north of Bumble Bee Bight (150 m below surface). Exposures of transported sulphides occur at the Bull Road and Henderson showings. Local stockwork mineralization is exposed at the Mansfield and Chess's showings.

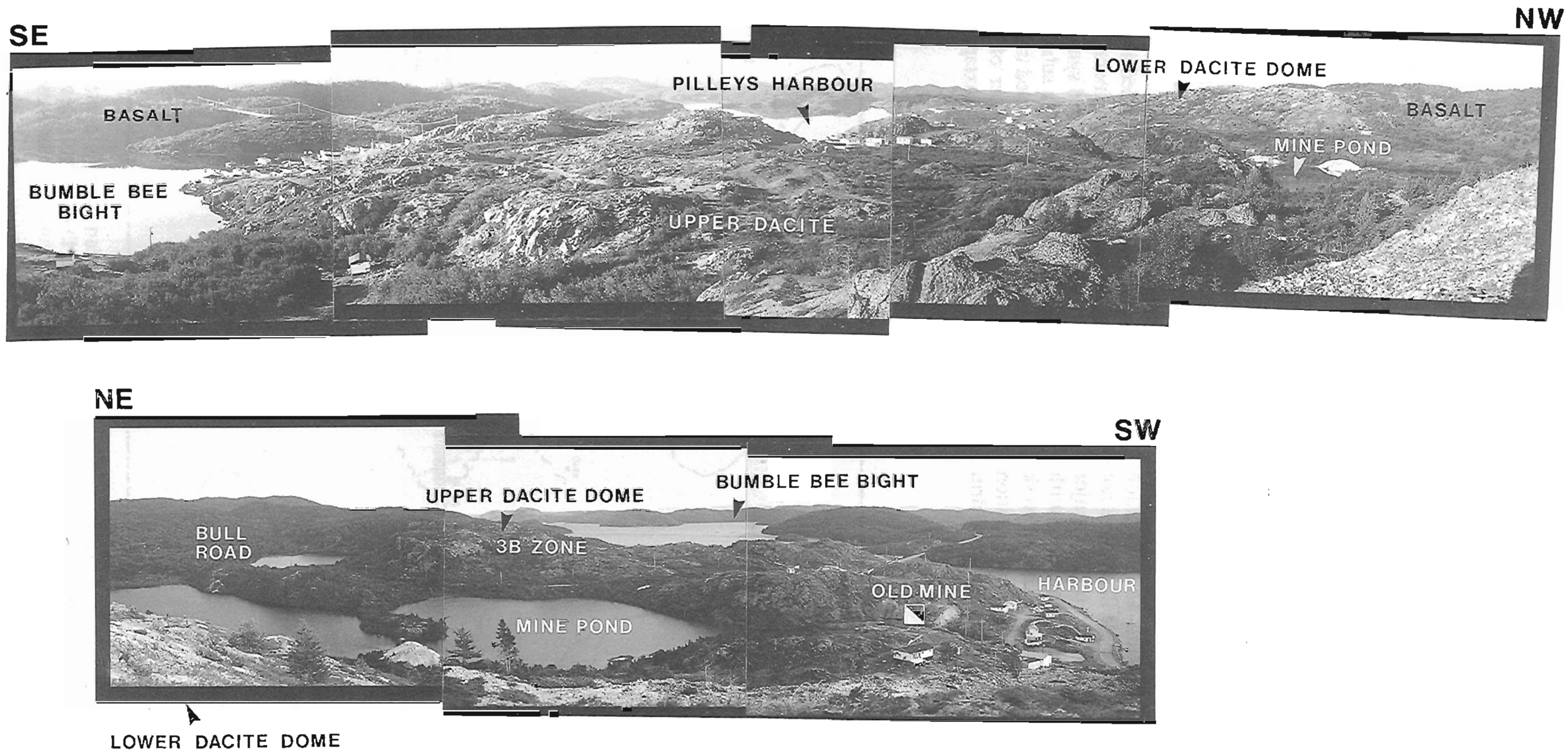

Figure 3. View of Bumble Bee Bight and the Mine pond area of Pilley's Island. Top photograph shows the location of the lower dacite dome. Bottom photograph shows the location of the new 3B Zone and mine workings at the Old Pilley's Island Mine.

lavas and hanging wall tuffs. Felsic and mafic dykes occur along several faults which have been interpreted by Tuach (1988, 1990) as synvolcanic.

HOST ROCK LITHOLOGIES AND VOLCANIC STRATIGRAPHY

The felsic volcanic pile which hosts the Old Pilley's Island Mine is dominated by two dacite lava domes which form prominent topographic features at surface (Fig. 3). The lava domes consist of multiple flow-banded units, amygdaloidal flows and flow breccia, and are separated by a thick pyroclastic unit (the upper dacite pyroclastics: Table 1). The lower dacite flows are pervasively altered, but the upper dacite flows are relatively fresh except near their lower contact. This suggests that the latest eruptive event must have occurred during the waning stages of hydrothermal activity and largely postdates mineralization (Tuach, 1988, 1990).

The pyroclastic rocks consist of an interbedded sequence of welded lapilli tuff, fine grained, silty-ash tuff, and pumice-rich volcanic breccia. Well preserved bedding features (e.g. graded bedding) are observed locally in the lapilli tuff at surface. Subangular to rounded bombs (10 cm to 1 m diameter), bomb fragments, and smaller lithic fragments comprise the coarse units. Large, partially fused volcanic bombs are also exposed and suggest proximity to an eruptive vent (Tuach, 1988, 1990). Sulphide fragments are present locally within the volcanic breccia, indicating that explosive volcanism was contemporaneous with mineralization. Welding of glass shards in the tuffaceous units may suggest shallow water volcanism and possible emergence of the felsic volcanic pile (Tuach, 1988, 1990). In the Old Mine sequence, the lower dacite pyroclastics host mineralization and overlie the basaltic footwall (Table 1). Near the 3B Zone, the upper pyroclastic sequence has accumulated to a maximum thickness of approximately 130 m and appears to occupy a local paleotopographic depression which is bound by synvolcanic faults (Fig. 4).

Several poorly-sorted, epiclastic units occur at the upper and lower contacts of the altered tuff package which hosts the 3B Zone. These are dominantly coarse, polylithic breccias similar to those observed at Buchans (Binney, 1987). Several distinctive breccia units at the thin outer margins of the 3B Zone contain fragments of altered footwall lavas, suggesting that they were eroded from uplifted fault blocks.

MINERALIZATION AND SULPHUR ISOTOPES

Mineralization in the Old Pilley's Island Mine consists of two massive pyritic zones up to 12 m thick and 180-300 m in strike length (Tuach, 1988, 1990). The sulphides consist of massive, fine grained to coarse crystalline pyrite and chalcopyrite with minor sphalerite. Massive sulphides locally display textures typical to seafloor deposition (e.g. colloform banding, sedimentary layering, sulphide talus breccia). However, abundant disseminated and stockwork-like sulphides, together with quartz-rich hydrothermal breccia, occur both in the immediate footwall and hanging wall. A number of perched sulphide lenses also occur in the altered hanging wall tuff. The abundance of vein and disseminated textures in the hanging wall tuff suggests that continuous hydrothermal venting onto the seafloor was frequently interrupted by pyroclastic eruptions.

Drilling in 1983-84 intersected a new sulphide zone, the 3B Zone, to the east of the Old Mine and at about 150 m depth. The sulphide horizons occur within intensely altered tuff and tuff breccia similar to that of the Old Mine horizons. Mineralization in the 3B Zone consists of multiple bands and disseminations of semi-massive sulphide with individual lenses of massive pyrite and chalcopyrite from a few centimetres up to 6 m in thickness. Collectively, the mineralization comprises a 25-45 m sulphide-rich zone containing 30-40% total sulphides. The central portions of the sulphide zone resemble mineralization in the Old Mine (e.g., 1% Cu over 5 m in hole 84-07), but sulphides at its outer margins and in a few discrete lenses above the main horizon are more polymetallic (e.g., 5.5% Zn, 4.3% Cu, and 0.1% Pb over 2 m in hole 84-09).

Several occurrences of transported sulphides (Bull Road and Henderson showings) are exposed at surface to the east of the Old Mine. The results of drilling between 1951 and 1970 could not unequivocally relate the occurrence of transported sulphides at Bull Road to the in situ ore at the Old Mine (Tuach, 1988, 1990). However, recent drilling on the up-dip extension of the new 3B Zone suggests that it is the likely source of sulphide debris on the Bull Road horizon (Fig. 4). The sulphide debris at Bull Road is distinctly polymetallic, with average outcrop assays of 12.4% Zn, 3.8% Cu, and 1% Pb (Tuach, 1988, 1990), and occurs within a polylithic breccia similar to that found in the hanging wall of the 3B Zone. Clastic textures within the sulphides closely resemble textures observed in the transported ores from Buchans (Thurlow and Swanson, 1981; Binney, 1987). Large pyrite boulders (up to 1 m across) are also exposed within polylithic breccia at the Henderson showing, but the source of this debris is not certain. The sulphides at Henderson occur at the eastern margin of the altered tuffs are much closer to the basalt contact than the sulphides at Bull Road.

The isotopic compositions of massive sulphides from the Old Mine and the Bull Road showing were determined using a 602D dual collector mass spectrometer at the University of Waterloo. Sulphur isotope ratios for samples from the Old Mine range from 2.3 to 6.5 per mil (Table 2) and are likely a product of mixing between ascending hydrothermal fluids and seawater. Similar results have been obtained for sulphides from other deposits in the Robert's Arm-Buchans belt (e.g., 2.5-7.6 per mil from Gullbridge: Bachinski, 1978; 2.9-8.7 per mil from Buchans: Kowalik et al., 1981) as well as from deeper volcanogenic stockworks (e.g., 2.3-6.2 per mil from Crescent Lake: Waldie et al., 1991). However, sulphur isotope ratios for samples of transported sulphides from the Bull Road horizon range from 1.1 to 1.8 per mil and are consistent with a source other than the in situ ore at the Old Mine.

Table 1. Volcanic stratigraphy of the mine sequence on Pilley's Island

LITHOLOGY	(m)	DESCRIPTION
6 a Felsic Dyke	-	Glassy, purple colour
b Mafic Dyke	-	Fine grained, porphyritic
5 Upper Dacite Lavas	60-160	Multiple, thick massive flows, flow-banded units, and flow breccia. Massive flows are fine grained (grey-green color) to glassy (purple-pink color) with sparse to abundant amygdules. Individual massive flows are 20 to 40 m thick and can be correlated between holes; banded flows are typically <10 m thick. The upper dacite lava dome forms a prominent topographic feature above the 3B Zone.
4 a Upper Dacite Pyroclastics	10-+180	Interbedded sequence of welded lapilli tuff, pumice-rich volcanic breccia, and fine grained, silty-ash tuff. Tuff breccia and welded lapilli tuff dominate the sequence; fine grained ash tuff occurs locally and is commonly bedded. Large (10 cm - 1 m) subangular to rounded volcanic bombs are exposed at surface. The pyroclastics are locally altered to quartz-sericite with abundant hydrothermal breccia.
b Polylithic Breccia	3-21	Poorly-sorted, coarse grained epiclastic unit dominated by fragments of dacite flow and tuff. Typically occurs at or near the contacts of the upper dacite pyroclastics (4a). Massive sulphide clasts and fragments of altered footwall lava are common in the breccias at Bull Road and Henderson. Occurrences of polylithic breccia tend to be laterally restricted and may be channel-fill deposits.
3 Lower Dacite Lavas	16-145	Massive flows and flow breccia. Flows are fine grained (grey-green color) to glassy (purple-pink color) with few banded units, and individual flows are up to 20 m thick. The flows are strongly altered to quartz-sericite±K-feldspar and contain abundant disseminated sulphides and hydrothermal breccia. The lower dacite lava dome forms a prominent topographic feature to the west of the Old Pilley's Island Mine.
2 Lower Dacite Pyroclastics	15-55	Interbedded sequence of lithic lapilli tuff, fine grained ash tuff, and pumice-rich volcanic breccia. Lithic lapilli tuff with few vitric fragments and fine grained tuff dominate the sequence. The pyroclastics contain abundant disseminated and massive sulphides with quartz-sericite and chlorite alteration, and they are a principal host lithology in the Old Pilley's Island Mine.
1 Basalt	+100	Massive flows, pillow lava, and mafic breccia. Interpillow jasper occurs locally. Mafic breccia consists of pillow and scoria fragments. Massive flows are commonly hematized, with abundant calcite-filled amygdules. Silicified pillow lavas are present in fault contact with the overlying felsic volcanics.

ALTERATION

The extent of alteration associated with local hydrothermal activity is conspicuous at surface by the lack of vegetation on the dacite domes (Fig. 3). Widespread potassic alteration also shows up clearly in regional, airborne gamma-ray surveys (Geological Survey of Canada, 1990). In the Old Mine, alteration extends from the pillow basalt contact well into the hanging wall tuff. Silicified pillow lavas occur in the footwall and are present locally in fault contact with the overlying felsic volcanic rocks. Interpillow jasper in the basalt immediately beneath the massive sulphides has been pyritized. Vesicles in amygdular flows throughout the mafic volcanic sequence are filled mainly by late stage carbonate and locally by chlorite.

The felsic volcanics in the immediate footwall and hanging wall to the massive sulphide horizons are variably sericitized, chloritized, and silicified. Altered flows contain abundant, fine grained sericite and chlorite, together with potassium feldspar and recrystallized quartz. Amygdules in the flows are commonly filled by quartz, pyrite, and lesser amounts of sericite and chlorite. Mineralized fractures, quartz veins, and hydrothermal breccia in the immediate stockwork also contain abundant chlorite. The dacite tuff and tuff breccia which host mineralization have been entirely replaced by quartz-sericite and chlorite assemblages. Alteration in the hanging wall tuff consists mainly of quartz and sericite. Weakly disseminated pyrite (1-2%) occurs throughout the altered felsic sequence; the most intense disseminated sulphide mineralization above and below the massive sulphides typically occurs within a distinctive coarse grained, sericite assemblage. Sulphide bands in the mineralized zones are commonly brecciated and in-filled with quartz, sericite, and late stage carbonate. The alteration assemblages have been mapped in the Old Mine area by Appleyard and Bowles (1978), and a similar alteration envelope occurs around the 3B Zone. Chlorite is less abundant than in the Old Mine, and hydrothermal breccias in the footwall dacite beneath the 3B Zone show more intense potassic alteration. Fragments of the coarse grained potassium feldspar from the altered footwall lavas are found in the sulphide debris and polylithic breccia at Bull Road.

In the Old Mine, the relative abundance of chlorite versus sericite increases systematically with depth through the altered and mineralized horizons, and several stages of chlorite are recognized. Early chlorite (with sericite) in the least altered dacite tuff and flows is distinguished by its light green color, low birefringence, and invariably fine grained nature. Chlorite which completely replaces glass shards in the altered hanging wall tuffs and which occurs within amygdules in the altered dacite lavas commonly possesses anomalous blue pleochroism. Both types of chlorite are relatively depleted in Fe and enriched in Mg compared to chlorite from the mineralized zone (Table 3). The abundance of matrix chlorite in the altered tuffs increases dramatically within the mineralized zone and in the immediate stockwork

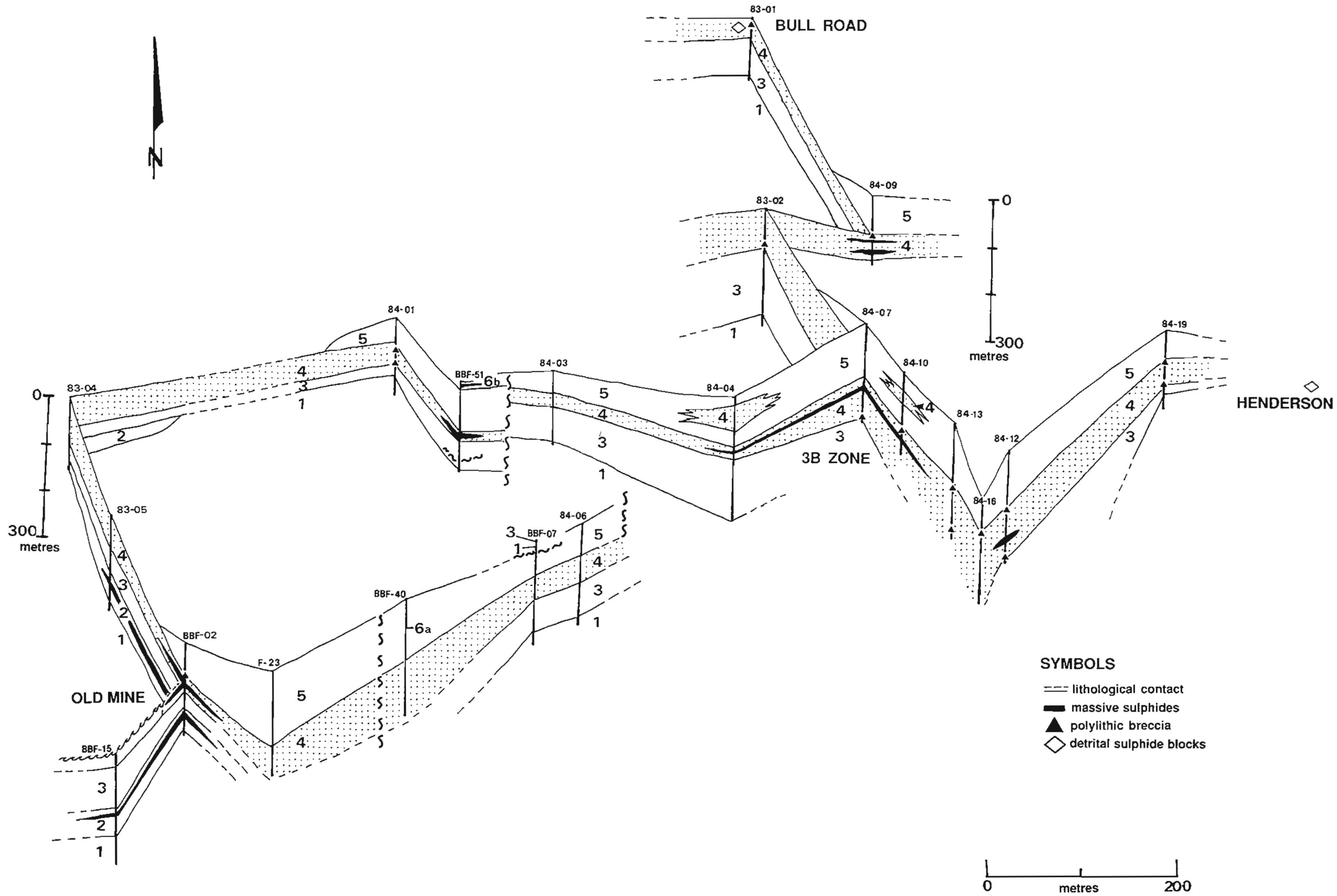

Figure 4. Schematic cross sections of the Old Mine and the new 3B Zone showing local volcanic stratigraphy (see Table 1 for legend). All holes are projected to vertical.

of the Old Mine. This chlorite is relatively coarse grained, and is characterized by its anomalous dark green pleochroism and high Fe content. A late stage, Mg-rich chlorite locally replaces the Fe-rich chlorite and may be related to an influx of cold seawater in the waning stages of hydrothermal activity. Mg-rich chlorite also occurs in the altered footwall basalt as thin blades with straight extinction and first order pleochroism.

Table 2. Sulphur isotope ratios of sulphides from the Old Pilley's Island Mine and transported sulphides from the Bull Road horizon

LOCATION	SULPHIDE	^{34}S
Old Mine	chalcopyrite	2.32
Old Mine	pyrite	3.20
Old Mine	pyrite	6.53
Bull Road	pyrite	1.77
Bull Road	sphalerite	1.73
Bull Road	chalcopyrite	1.11
Bull Road	sphalerite	1.22
Bull Road	chalcopyrite	1.32
Bull Road	pyrite	1.14
Bull Road	chalcopyrite	1.82

The compositions of chlorite correlate with the intensity of hydrothermal alteration and display a distinctive zonation with respect to the mineralization (Fig. 5a). Application of a chlorite geothermometer developed by Cathelineau and Nieva (1985) and modified by Jowett (in press) indicates a thermally intensifying hydrothermal system (Fig. 5b). Temperatures of alteration increase from about 240°C during the early sericite-chlorite stage to about 360°C during mineralization (average calculated temperature for sample 90069).

CONCLUSIONS

The sulphide deposits on Pilley's Island are closely associated with a brief period of felsic volcanism and the extrusion of a dacite lava dome onto regionally extensive basaltic flows (Tuach, 1988, 1990). Preliminary studies of the sulphides and related alteration minerals, together with descriptions of the volcanic lithologies, place additional constraints on the geologic setting of mineralization. Synvolcanic faulting and a major paleotopographic depression appear to have localized mineralization, but continuous pyroclastic eruptions prevented sustained hydrothermal venting onto the seafloor. Stockwork and disseminated sulphides locally dominate the mineralization and suggest sub-seafloor precipitation as a result of mixing of hydrothermal fluids with seawater within the pyroclastic sequence, or possibly due to boiling of the hydrothermal fluids. Examination of the alteration mineral assemblage

Table 3. Microprobe data for chlorite from the alteration envelope around the Old Pilley's Island Mine. Samples are from drill hole 83-05 (N = number of analyses).

REPRESENTATIVE MICROPROBE ANALYSES OF CHLORITE

	Unmineralized Dacite		Mineralized Zone			Basalt			
sample	90055	90061	90065	90069	90070	90073	90074	90076	90077
N	3	1	3	5	3	3	7	3	5
depth (m)	29.8	81.8	114	132	140	158	162	174	181
SiO2	31.69	31.35	28.94	25.79	27.52	30.65	31.52	30.75	30.51
TiO2	0.018	0.080	1.423	0.055	0.052	0.004	0.020	0.005	0.007
Al2O3	17.78	24.18	23.03	21.74	22.12	17.48	16.57	15.19	16.98
Cr2O3	0.015	0.013	0.001	0.056	0.297	0.000	0.535	0.008	0.014
MgO	22.03	21.54	18.43	16.32	18.10	25.00	26.50	22.10	22.64
CaO	0.061	0.075	0.028	0.053	0.022	0.063	0.082	0.095	0.074
MnO	0.586	0.350	0.196	0.399	1.227	0.485	0.487	0.311	0.594
FeO	15.28	10.65	16.43	23.38	19.37	14.10	12.66	18.56	17.24
NiO	0.020	0.000	0.005	0.026	0.009	0.101	0.077	0.033	0.032
Na2O	0.031	0.048	0.029	0.030	0.009	0.024	0.009	0.016	0.024
K2O	0.275	0.996	0.593	0.039	0.155	0.015	0.020	0.021	0.031
H2O	12.01	12.65	12.21	11.52	11.90	12.15	12.31	11.76	11.98
total	99.89	101.9	101.3	99.41	100.8	100.1	100.8	98.84	100.1

shows that high-temperature fluids migrated well into the hanging wall tuffs, and these rocks are locally completely altered. The correlation of chlorite compositions with high-temperature hydrothermal alteration at Pilley's Island is distinctive and constitutes a useful exploration tool.

Polylithic breccia which accumulated during breaks in the pyroclastic volcanism contains debris from the massive sulphide horizons as well as fragments of altered footwall volcanic rocks. These units are thickest at the top and bottom of the pyroclastic sequence and are likely related to the synvolcanic faulting. Samples of the transported sulphides from the Bull Road horizon are chemically and isotopically distinct from sulphides in the Old Pilley's Island Mine and were likely derived from the erosion of the new 3B Zone. Similar studies of the transported sulphides at Henderson, which are isolated from their source, may provide useful information about the possible existence of additional undiscovered ore.

ACKNOWLEDGMENTS

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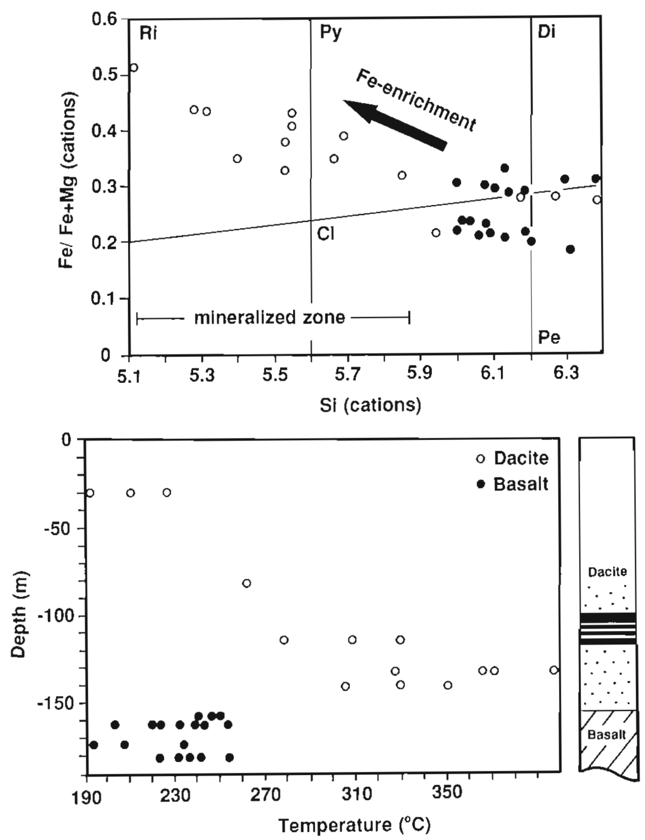

Figure 5. Chlorite chemistry from the Old Pilley's Island Mine (hole 83-05). (a) Si versus Fe/Fe+Mg (chlorite composition diagram from Hey, 1954). Ri-ripidolite, Py-pycnochlorite, Di-diabantite, Cl-clinochlore, Pe-penninite. (b) Calculated temperatures for hydrothermal alteration based on a chlorite geothermometer (Jowett, in press).

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AUTHOR INDEX

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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 scientifique en chef. 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 enregistré sur disquette WordPerfect 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 1992 (Paper 92-1, 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 1992 (Étude 92-1, parties A à D) sont énumérées ci-dessous et vendues séparément.

Part A, Cordillera and Pacific Margin
Partie A, Cordillère et marge du Pacifique

Part B, Interior Plains and Arctic Canada
Partie B, Plaines intérieures et région arctique du Canada

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