Geological Survey of Canada
Commission géologique du Canada

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EASTERN AND ATLANTIC CANADA

RECHERCHES EN COURS PARTIE B
EST ET RÉGION ATLANTIQUE DU CANADA

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GEOLOGICAL SURVEY OF CANADA
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PAPER/ÉTUDE 89-1B

CURRENT RESEARCH, PART B
EASTERN AND ATLANTIC CANADA

RECHERCHES EN COURS, PARTIE B
EST ET RÉGION ATLANTIQUE DU CANADA

1989

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Lower Silurian Goldson Conglomerate with large unsorted clasts of bedded shale and siltstone, New Bay, northeast Newfoundland. Photo by H. Williams
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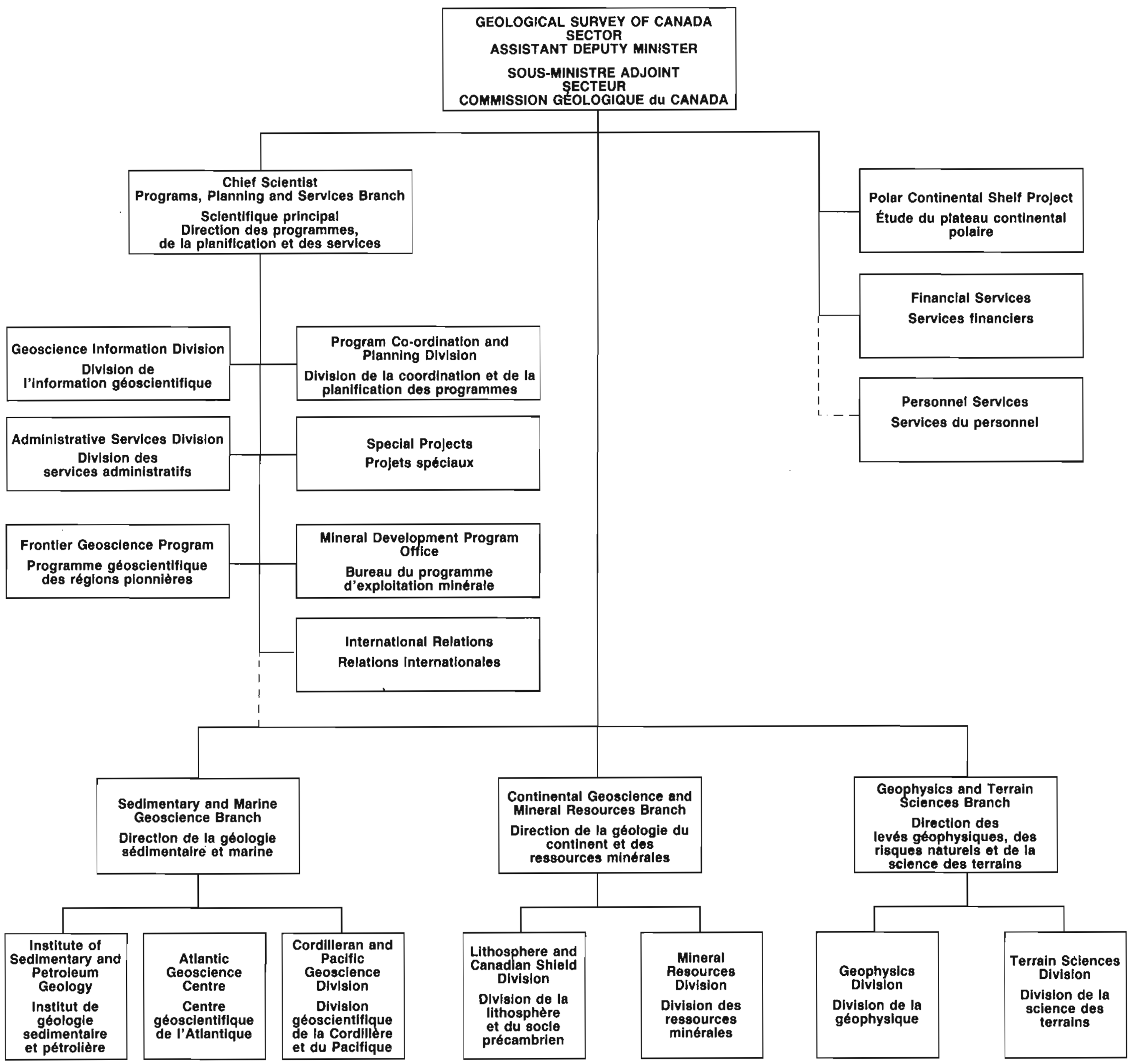
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Aeromagnetic total field, gradiometer, and VLF-EM survey of the Cobequid Highlands, Nova Scotia[1]

F.G. Kiss, E.E. Ready, P.E. Stone, and D.J. Teskey
Geophysics Division

Kiss, F.G., Ready, E.E., Stone, P.E., and Teskey, D.J., Aeromagnetic total field, gradiometer, and VLF-EM survey of the Cobequid Highlands, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 1-5, 1989.

Abstract

An aeromagnetic, gradiometer and VLF-EM survey was flown in the Cobequid Highlands area of Nova Scotia. A comparative study of these geophysical data sets with the mapped geology demonstrates the effectiveness of an integrated aeromagnetic, gradiometer and VLF-EM system in providing useful structural and lithological information in areas where the geology is complex and outcrop is poor.

The Cobequid Fault and the major unconformity that marks the northern limit of the Cobequid complex are particularly well defined. Improved resolution of contacts between large acidic and basic intrusive bodies such as the Wyvern and the Gilbert Mountain plutons, and on a smaller scale accurate positioning of diabase dykes is demonstrated. The mafic volcanic units of the Fountain Lake Group are resolved on the gradiometer map and interpreted to extend beyond their present mapped limits.

Résumé

On a effectué dans les hautes terres de Cobequid en Nouvelle-Écosse un levé combiné aéromagnétique du champ total, gradiométrique et électromagnétique à très basse fréquence (VLF-EM) aérien. Une comparaison entre ces ensembles de données géophysiques et les cartes géologiques révèle qu'un système intégré de levés aéromagnétiques, gradiométriques et VLF-EM renseigne efficacement sur la strucutre et la lithologie dans des régions à géologie complexe et affleurements rares.

La faille de Cobequid et l'importante discordance qui marque la limite nord du complexe de Cobequid sont particulièrement bien définies. Ce système permet en outre une meilleure résolution des contacts entre les grands corps intrusifs acides et basiques comme les plutons de Wyvern et de Gilbert Mountain, et la localisation à petite échelle plus précise des dykes de diabase. Les unités volcaniques mafiques du groupe de Fountain Lake sont définies sur la carte gradiométrique et interprétées au-delà de leurs limites cartographiques actuelles.

[1] Contribution to the Canada-Nova Scotia Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada, Geophysics Division.

INTRODUCTION

A detailed helicopter-borne total field, gradiometer and VLF-EM survey was conducted in the Cobequid Highlands area of Nova Scotia (Fig. 1) in 1986-1987 by Sander Geophysics Limited on contract to the Geological Survey of Canada. This program was funded by the Geological Survey of Canada under the Canada-Nova Scotia Mineral Development Agreement (1984-1989).

The objective of the survey was to provide data which would aid in the geological mapping process by enhancing detail on fault structures and by providing an additional dimension to lithological differentiation based on magnetic and electromagnetic measurements in the Cobequid Highlands area.

In this paper, a number of observations are made from the geophysical data presented in map form on NTS sheets 11E/5 and 11E/12 (Fig. 1).

Survey instrumentation

The helicopter-borne survey was equipped with a magnetic gradiometer and a VLF-EM receiver. Aircraft survey height was maintained at an average of 180 m. The lines were flown in a north-south direction with a spacing of 300 m. The gradiometer consisted of two Sander Geophysics Limited Overhauser magnetometers of 0.005 nT resolution with a vertical separation of 3 m mounted on a rigid boom structure and towed by a cable 30 m below the helicopter. The Sander Geophysics two frequency VLF-EMII receiver mounted on board the survey aircraft utilized VLF transmission stations from NSS, Annapolis, Maryland operating at 21.4 kHz and NAA, Cutler, Maine operating at 24.0 kHz. A Litton LTN-51 inertial navigation system supplemented by a vertically mounted camera for positional fix was used throughout the survey.

Data presentation

The aeromagnetic total field data and gradiometer data are compiled in contour form at a scale of 1:25 000. In addition, these data are also available at a scale of 1:50 000 as colour relief maps. The VLF-EM data in profile form are overprinted on the reverse side of these maps so that they can be directly correlated with the aeromagnetic data on the front side using a light table.

These products may be purchased from the Geological Survey of Canada. The digital data for this survey is also available on magnetic tape from the Geophysical Data Centre, 1 Observatory Crescent, Ottawa, K1A 0Y3.

DISCUSSION

Extensive mapping of the entire Cobequid Highlands area was completed and published in a series of reports by Donohoe and Wallace (1979, 1982, 1985). Later work by Murphy et al., (1988a) along with a map published at 1:50 000 scale (Murphy et al., 1988b) has incorporated revisions to the pre-Carboniferous geology in the eastern Cobequid Highlands. Figure 2 is a composite map of these works in the area of NTS sheets 11E/5 and 11E/12. Examination of this map in conjunction with the aeromagnetic and gradiometer maps and the VLF data shows a number of structural and lithological features that may assist in furthering the understanding of the geology of the area.

Faults and unconformities

The geological map of the area is superimposed on the total field magnetic map in Figure 2. The Cobequid Fault and the major unconformity that separates the Cobequid Highlands from younger Carboniferous rocks to the south (13) and north (14b) respectively are particularly well defined.

The total field magnetic map shows the distinct contrast between the highly magnetic Precambrian Great Village River Gneiss (1b) along with the schists of the Gamble Brook Formation (2a) and the weakly magnetic younger Carboniferous sediments (13, 14a), which underlie the southern flank of the Cobequid Highlands. This contrast is sharply delineated by a narrow linear trend on the vertical gradient map (Fig. 3). Similarly, the Londonderry and Rockland Brook faults are distinct linears, though less prominent.

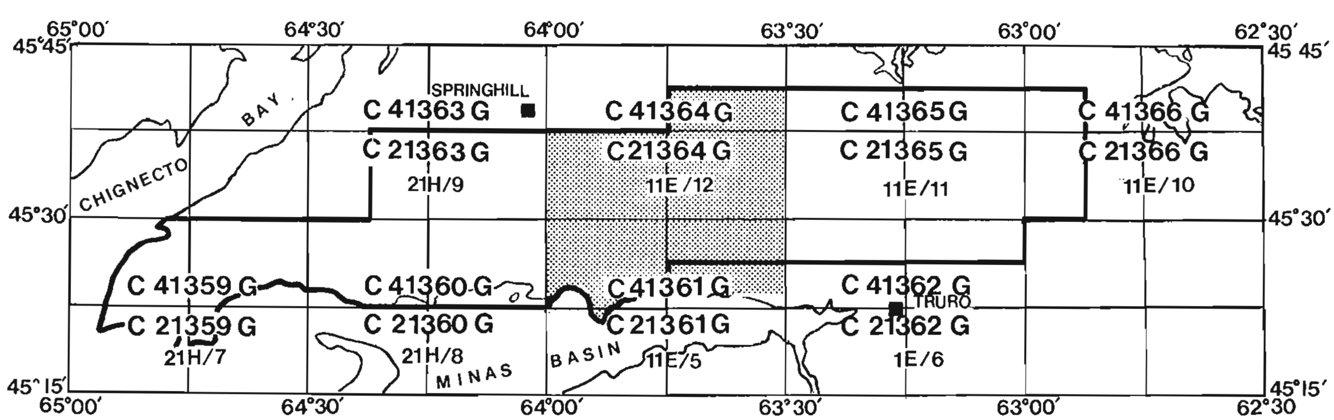

Figure 1. Location map of the Cobequid Highlands survey area in Nova Scotia with the geophysical series map sheet index. Stippling indicates area described in this report.

The unconformity between the lower Carboniferous granites (11b) and the upper Carboniferous sediments of the Riversdale Group (14b) on the northern flank of the Cobequid Highlands is also well defined on both the total field magnetic and gradiometer maps.

Significant VLF-EM responses attributable to geological conductors were detected only over the Cobequid Fault and the unconformity between the sedimentary rocks of the Riversdale (14b) and Cumberland groups (15) with the receiver tuned to the Annapolis transmitter station. Results simultaneously recorded using the Cutler transmitter signal, however, were not considered significant on these map sheets.

Contacts

Major contacts between the gabbros and granites can be delineated approximately parallel to the major fault systems.

An east-west lineation on the vertical gradient map (Fig. 3) separates the area mapped as the gabbroic Wyvern Pluton (WP, 11a) from the granitic Gilbert Mountain Pluton (GMP, 11b). On the black and white reproduction of the total field magnetic map (Fig. 2), the dark areas to the south and the light areas to the north reflect the contrast between the high and low magnetic susceptibility of the gabbros and granites respectively.

The contact between the Carboniferous diorites of the Folly Lake Pluton (FoLP, 11a) and the granitic Hart Lake-Byers Lake Pluton (HLBLP, 11b) is further exemplified in this manner on the total field magnetic map (Fig. 2) at the northern end of Rockland Brook, the headwaters of the Great Village River (e).

Units such as the Triassic North Mountain basalt flows (20) and the Devono-Carboniferous basalt flows of the Fountain Lake Group (9a) can be extended and better

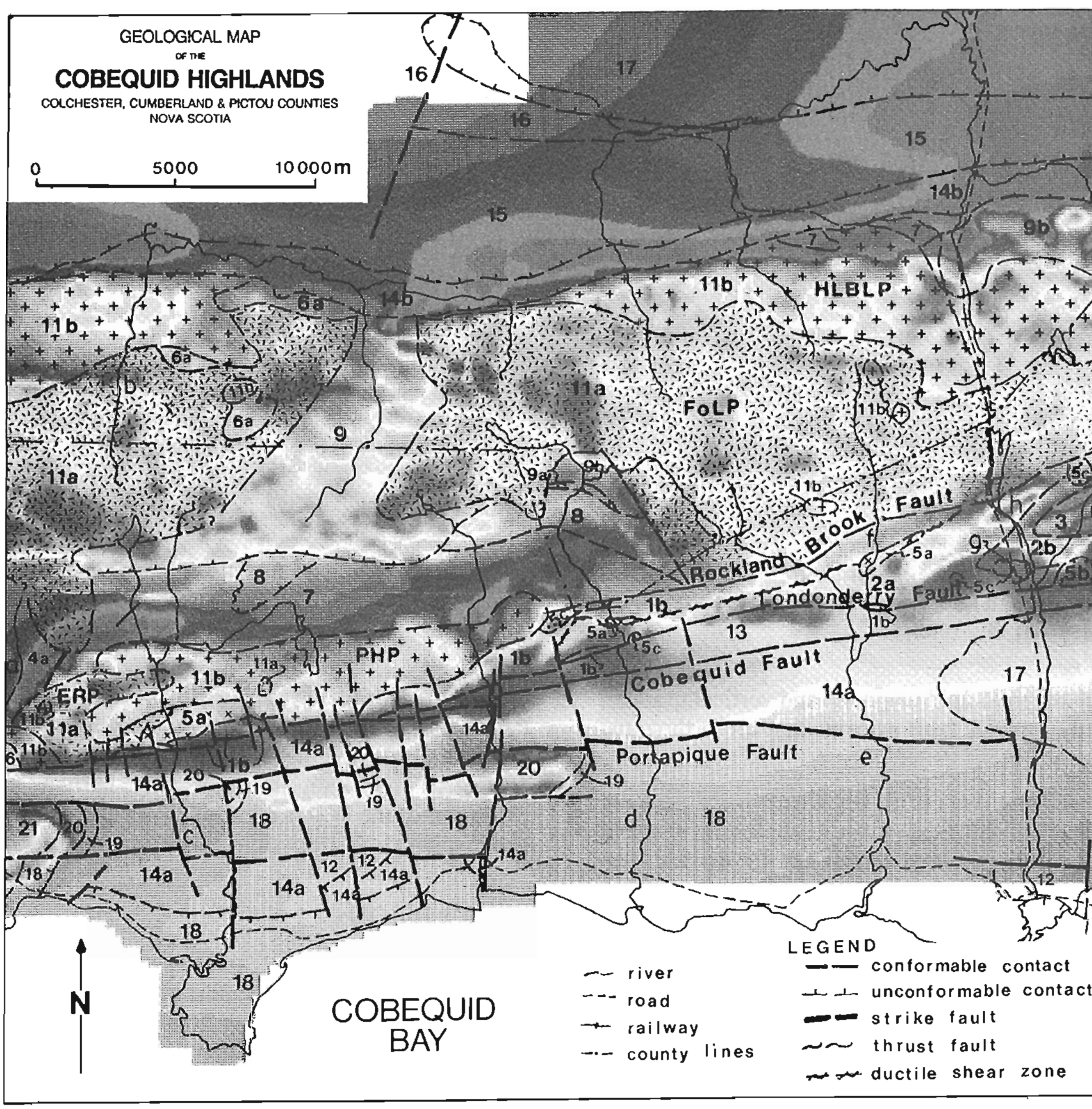

Figure 2. Black and white reduction of the total field magnetic map from 1:50 000 colour original print. Superimposed geology of the Cobequid Highlands prepared from composite data after Murphy et al. (1988a) and Donohoe and Wallace(1982)

Figure 3. Black and white reproduction of aeromagnetic vertical gradient map from 1:50 000 colour original print. Superimposed simplified geological data locating structural and lithological features discussed in this report.

defined on both the total field magnetic and vertical gradient maps, as can the smaller fault systems that are orthogonal to the major contacts.

Numerous narrow diabase dykes which were partially mapped crosscutting the Bass River complex in a northeasterly direction are resolved clearly on the vertical gradient magnetic map.

Buried magnetic bodies

A strong magnetic high is detected over an area mapped by Donohoe and Wallace (1982) and Murphy et al. (1988b) (Fig. 2) as undifferentiated mafic and felsic volcanic rocks of the Fountain Lake Group (9) in the Philip River area between Simpson Lake and Bass River Lake (*see* Fig. 3 for location). However, the area is devoid of outcrop and the lateral extent of these volcanic flows is uncertain. The Fountain Lake Group volcanic rocks in themselves do not appear to be strongly magnetic where they are mapped farther north. An underlying intrusive body, perhaps an eastward extension of the Wyvern Pluton under the volcanic rocks at shallow depth is a possible cause for this magnetic high.

CONCLUSIONS

The application of an integrated aeromagnetic gradiometer and VLF-EM survey in support of detailed geological mapping is effective in resolving lithological and structural features. Detailed study of these data sets should greatly contribute to the understanding of the geological structure of the Cobequid Highlands and thus help to define targets of potential economic significance.

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Preliminary kinematic analysis of the Rockland Brook Fault, Cobequid Highlands, Nova Scotia[1]

Brent V. Miller[2], R. Damian Nance[2], and J. Brendan Murphy[3]

Miller, B.V., Nance, R.D., and Murphy, J.B., Preliminary kinematic analysis of the Rockland Brook Fault, Cobequid Highlands, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 7-14, 1989.

Abstract

The E-W to NE-SW Rockland Brook Fault (RBF) broadly parallels the late Paleozoic Cobequid Fault of mainland Nova Scotia and has been cited as a potential Precambrian terrane boundary within the Avalon Composite Terrane. In its E-W trending western and central portions, the RBF records intense, dextral and locally transpressive ductile shear of late Paleozoic age. At its NE trending eastern end, late Paleozoic strike-slip movement is minimal and the RBF forms a structural front separating Hadrynian volcano-sedimentary successions with a pervasive flat-lying Precambrian cleavage to the west from their upthrown, undeformed equivalents to the east.

This earlier NE-trending segment (ancestral RBF) is believed to have been rotated and reactivated as an E-W ductile shear zone with the arrival of the Meguma Terrane during the late Paleozoic. The ancestral RBF is tentatively interpreted as a Precambrian thrust ramp but is unlikely to have been a major terrane boundary.

Résumé

La faille E-O à NE-SO de Rockland Brook est en général parallèle à la faille du Paléozoïque supérieur de Cobequid de la partie centrale de la Nouvelle-Écosse, et a été citée en tant que limite possible du terrain précambrien à l'intérieur du terrain composite d'Avalon. Dans ses parties ouest et centrale de direction E-O, la faille de Rockland Brook présente un intense cisaillement ductile, à caractère dextre et transpressif par endroits, datant du Paléozoïque supérieur. À son extrémité est de direction NE, le mouvement de rejet horizontal du Paléozoïque supérieur est minime et la faille forme un front structural qui sépare des séries volcano-sédimentaires de l'Hadrynien à clivage précambrien pénétrant à composante horizontale, à l'ouest, de leurs équivalents non déformés et soulevés, à l'est.

Ce segment antérieur de direction NE (faille ancestrale) aurait subi un mouvement de rotation et aurait été réactivé sous la forme d'une zone de cisaillement ductile à orientation E-O avec l'arrivée du terrain de Méguma pendant le Paléozoïque supérieur. Selon une interprétaation, la faille ancestrale de Rockland Brook serait une rampe de chevauchement, mais n'a probablement pas constituée une limite importante du terrain.

[1] Contribution to Canada-Nova Scotia Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada.

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INTRODUCTION

The Rockland Brook Fault (RBF) forms a major late Paleozoic strike-slip structure in the Cobequid Highlands of Nova Scotia (Fig.1). It runs parallel and adjacent to the better-known Cobequid Fault (Donohoe and Wallace, 1980, 1985) which, as part of the E-W Cobequid-Chedabucto Fault System (Minas Geofracture, Keppie, 1982), separates the northern Appalachian Avalon and Meguma terranes (Williams and Hatcher, 1983). A major portion of the RBF's movement history is related to (and probably a product of) movement on the Cobequid Fault since the latter is the dominant tectonic structure in the area and has accommodated major displacements (e.g. Eisbacher, 1969; Keppie, 1982;Mawer and White, 1987). However, the RBF may also record an earlier history of movement (here termed the "ancestral RBF") and has been assumed to form a Precambrian terrane boundary within the Avalon Composite Terrane that separates the proposed Bass and Cobequid terranes (Keppie, 1985).

Despite the potential significance of the RBF, only in its central portion does a consensus exist as to its path. Donohoe and Wallace (1982) trace it eastward to the Cobequid Fault and terminate it to the west at the intrusive boundary of a Carboniferous stitching pluton. Cullen (1984), however, suggested the fault continued westward along the southern margin of this pluton and Murphy et al. (1988) give its eastern portion a northeasterly path. The existence of both NE-SW and E-W trends would be significant in that, with the exception of the E-W structures of mainland Nova Scotia, major northern Appalachian faults trend broadly NE-SW. Hence, the nature and kinematics of the RBF both within and between its E-W and NE-SW segments may bear upon the relationship between these two major fault groups.

This report presents the preliminary results of a detailed examination of the RBF undertaken to determine its kinematic history, assess the significance of its trace and test its role as a potential Precambrian terrane boundary.

PREVIOUS WORK

Donohoe and Wallace (1980, 1982, 1985) describe the RBF as a broad zone of mylonitization in the central Cobequid Highlands which they attributed to predominantly strike-slip ductile shear of mid-Carboniferous age. This zone followed but did not define the northern margin of the Bass River Complex and generally separated this polydeformed Precambrian basement-cover sequence (units 1b, 2, 3; Fig. 1) to the south from undeformed Devono-Carboniferous plutons and volcanic rocks (units 9, 11; Fig. 1) and Silurian volcanogenic sediments (unit 7; Fig. 1) to the north. However, at its eastern extremity, where the zone was considered to merge with that of the Cobequid Fault, the RBF was positioned south of the Mount Thom Complex (unit 1a; Fig. 1) with which parts of the Bass River Complex have been correlated (Donohoe, 1983; Donohoe and Cullen, 1983). Such a path would imply a minimum dextral displacement of about 40 km between the two complexes but, in separating correlated basement units, disqualifies the RBF as a Precambrian terrane boundary. In its central portion, the zone of mylonitization strongly affects the southern margin of the 348±5 Ma (Rb/Sr, whole rock; Donohoe et al., 1986) Hart Lake-Byers Lake granite pluton (HLBLP; Fig. 1). At its western extremity, the RBF was considered by Donohoe and Wallace (1982) to be truncated ("stitched") by the 315±25 Ma (Rb/Sr, whole rock; Cormier, 1980) Pleasant Hills granite pluton (PHP; Fig. 1). Hence, cessation of fault movement was considered to be

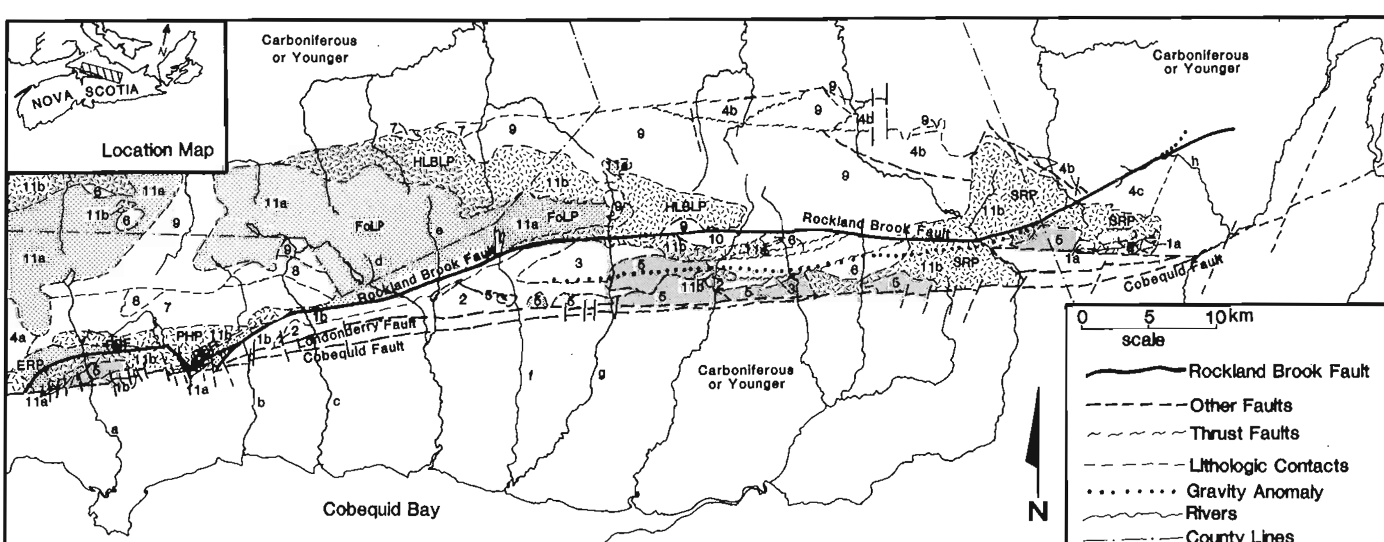

Figure 1. Geological map of the eastern Cobequid Highlands, Nova Scotia, adjacent to the Rockland Brook Fault (RBF) (modified after Donohoe and Wallace, 1982 and Murphy et al., 1988). See Table 1 for legend. Identified Devono-Carboniferous plutons include the Hart Lake-Byers Lake (HLBLP), Pleasant Hills (PHP) and Salmon River (SRP) granites,and the Folly Lake (FoLP) and Economy River (ERP)diorites. Identified locations are as follows: (a) Economy River; (b) Bass River; (c) Portapique River; (d) Great Village River; (e) Rockland Brook; (f) Folly River; (g) Debert River; and (h) Six Mile Brook.

broadly confined to the interval 315-350 Ma. However, as described below, the results of this study refute the interpretation of the Pleasant Hills granite as a stitching pluton and place its ca. 315 Ma age in question.

Cullen (1984) similarly considered the RBF to be a steeply dipping ductile shear zone but concluded that it defined the northern boundary of the Bass River Complex and, rather than being truncated by the Pleasant Hills Pluton, continued along its southern margin. Cullen (1984) also described several localities where pre-existing fabric elements within the Bass River Complex had been rotated into parallelism with the RBF in such a fashion as to indicate dextral shear (e.g. sub area C; Fig. 2). These conclusions are supported by this study.

More recently, Murphy et al. (1988) have reinterpreted the position of the eastern portion of the RBF by linking it with a previously unnamed NE-trending fault near Six Mile Brook (h; Fig. 1). This path is significant in making the RBF one of the few major northern Appalachian faults to show both NE-SW and E-W trends. The path is also more compatible with the proposed Precambrian terrane boundary, as it places basement rocks of the Mount Thom and Bass River complexes within the same (southern) fault block. Furthermore, the path separates deformed Late Hadrynian volcanics and volcanogenic sediments to the northwest (Warwick Mountain Formation, unit 4b, Fig. 1) that are similar to those of the western Cobequid Highlands (Jeffers Group; Pe-Piper and Piper, 1987, in press) from undeformed Late Hadrynian volcanics and volcanogenic sediments to the southeast (Dalhousie Mountain Volcanics, unit 4c; Fig. 1) that are similar to those of the Antigonish Highlands (Keppoch Formation; Murphy,1984, 1985) to the east (Murphy et al., in press). This change in trend was based on a linear gravity anomaly which broadly coincides with the RBF and bends in the area of the Salmon River Pluton (SRP; Fig. 1). However, the path could not be unequivocally demonstrated on the basis of field evidence due to a lack of exposure. Nevertheless, other major fault systems such as the Cobequid-Hollow system (Eisbacher, 1969; Yeo and Ruixiang, 1987) take a similar bend in the same area.

Table 1. Legend to Figure 1 (Stratigraphy modified after Donohue and Wallace, 1982 and Murphy et al., 1988).

11	Devono-Carboniferous plutons a) gabbro/diorite b) granite
10	Nuttby Formation
9	Fountain Lake Group
8	Portapique River and Murphy Brook formations
7	Wilson Brook Formation
6	Silurian volcanics and other undivided Silurian rocks
5	Late Precambrian-Cambrian plutons
4	a) Jeffers Group b) Warwick Mountain Formation c) Dalhousie Mountain Volcanics
3	Folly River Formation
2	Gamble Brook Formation
1	a) Mount Thom Complex b) Great Village River Gneiss

GEOLOGICAL SETTING

Detailed descriptions of the units adjacent to the RBF are given by Donohoe and Wallace (1980, 1985), Cullen (1984), Pe-Piper and Piper (1987), Murphy et al. (1988) and Nance and Murphy (1988), and are only briefly summarized here.

Precambrian units

Precambrian rocks spatially associated with the RBF include the Bass River Complex the Mount Thom Complex and a number of late Precambrian volcano-sedimentary successions (Fig. 1). The Bass River Complex (Donohoe and Wallace, 1980, 1985; Cullen, 1984; Nance and Murphy, 1988) consists of a highly metamorphosed and polydeformed basement succession of: hornblende amphibolites, hornblende granitoid orthogneisses and psammitic paragneisses (Great Village River Gneiss, unit 1b; Fig. 1); metasedimentary platformal metaquartzites, pelitic schists and carbonates (Gamble Brook Formation, unit 2; Fig. 1); syntectonic granite gneisses; and a greenschist facies mafic metavolcanic sequence locally containing sheeted dykes and pillow basalts (Folly River Formation, unit 3; Fig. 1). The Mount Thom Complex (Donohoe and Wallace, 1980, 1985; Murphy et al. 1988) comprises biotite-garnet microcrystalline gneisses, amphibolites and granitoid gneisses, and is considered to be correlative with the Great Village River Gneiss (Donohoe, 1983; Donohoe and Cullen, 1983).

Both the Late Hadrynian Jeffers Group (unit 4a; Fig. 1) and Warwick Mountain Formation (unit 4b; Fig. 1) comprise interlayered felsic and mafic metavolcanics, quartz and volcanic metawackes, metasiltstones and minor marbles, and possess a characteristic flatlying cleavage (Pe-Piper and Piper, 1987, in press; Murphy et al., 1988). The time-equivalent Dalhousie Mountain Volcanics (unit 4c; Fig. 1) are lithologically similar to the Warwick Mountain Formation but lack the penetrative flat- lying cleavage and resemble the Keppoch Formation of the Antigonish Highlands (Murphy et al., 1988).

Paleozoic rocks

Cambrian to Early Ordovician rocks do not occur in the eastern Cobequid Highlands and Late Ordovician to Early Devonian rocks occur only in small thrust slices south of the RBF (Murphy et al., 1988). Silurian strata north of the RBF (Wilson Brook Formation, unit 7; Fig. 1) are not pervasively deformed by the fault, as our mapping has shown that a thin band of Carboniferous granite separates the formation from the RBF where it had previously been interpreted to

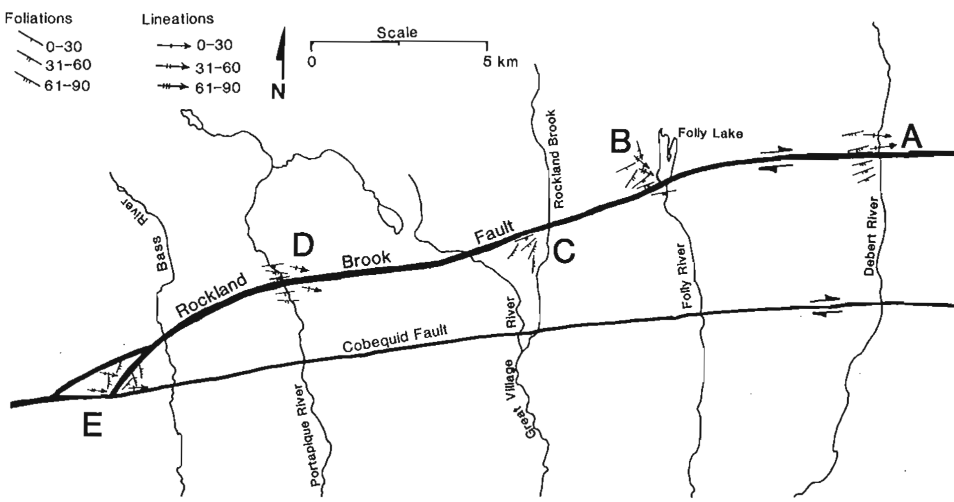

Figure 2. Representative structural data from subareas A through E on the E-W segment of the Rockland Brook Fault. See Figure 3 for stereoplots of subarea data.

be in fault contact with the Bass River Complex (Donohoe and Wallace, 1982). This sliver of granite intrudes the Wilson Brook Formation several tens of metres north of any RBF-related deformation and shows a progressive increase in deformation to ultramylonite as the shear zone is approached.

The youngest units adjacent to the RBF are Devono-Carboniferous plutons (Cormier, 1979, 1980; Donohoe et al., 1986) which occur along most of its length and lie largely to the north of the fault. These include the Pleasant Hills, Hart Lake-Byers Lake and the possibly stitching Salmon River plutons (PHP, HLBLP and SRP, respectively; Fig. 1), which comprise distinctive red to pink granites with large feldspar megacrysts that often show a rapakivi texture. The Folly Lake and Economy River plutons (FoLP and ERP; Fig. 1) are coarse grained gabbros intruded by porphyritic diorite, granodiorite and late fine grained dykes and sills (Murphy et al., 1988).

ROCKLAND BROOK FAULT

For the purpose of this study the RBF can be defined for much of its length as the single major shear zone and related structures that separate Precambrian basement units to the south from Devono-Carboniferous rocks which occur primarily to the north. Although shear zones in the western Cobequid Highlands (e.g. the Kirkhill Fault of Pe-Piper and Piper, 1987) may be structurally related to the RBF, the western limit of the RBF is here taken as the westernmost splay that contains units of the Bass River Complex in the southern fault block (Fig. 1). The present study therefore supports the conclusion of Cullen (1984) that, rather than being truncated by the eastern margin of the Pleasant Hills Pluton (PHP; Fig. 1) as proposed by Donohoe and Wallace (1982), the RBF continues as several splays along the southern edge of the pluton in what is actually a previously unmapped dioritic intrusion (Fig. 1). East of the Pleasant Hills Pluton the central portion of the RBF is the single well-defined mylonite zone defined by Donohoe and Wallace (1982) and shows only minor splaying as noted by Cullen (1984).

The eastern portion of the RBF is less easily defined and disagreement exists as to its position. Donohoe and Wallace (1982) considered the RBF to merge with the Cobequid Fault east of the Mount Thom Complex. However, the continuation they proposed does not mark the course of a ductile shear zone like that of the central portion of the RBF, but rather, is a largely brittle fault. In contrast, Yeo (1985) and Yeo and Ruixiang (1987) show the fault, later considered by Murphy et al. (1988) as the eastern extension of the RBF, to terminate within the Late Namurian to Early Westphalian Riversdale Group east of the Salmon River Pluton (SRP; Fig. 1). Both traces are here considered to define the path of the RBF, but while the former marks the trace of its Late Paleozoic reactivation, the latter path is considered to be that of the ancestral RBF. However, the Salmon River granite, which would be a stitching pluton with respect to the ancestral RBF, may also stitch the main Late Paleozoic ductile phase of movement on the RBF and hence shows only late-stage brittle faulting.

STRUCTURAL GEOMETRY OF THE E-W FAULT SEGMENT

Deformation associated with the western and central portions of the RBF takes the form of an intense dextral mylonitic fabric that strongly affects the ca. 350 Ma Hart Lake-Byers Lake Pluton and locally overprints an earlier fabric within the Bass River Complex that is considered to be of late Precambrian age (Nance and Murphy, 1988). The observed deformation is therefore considered to be late Paleozoic. However, a pre-Carboniferous movement history cannot be discounted although, if present, its effects on pre-Carboniferous rocks are unclear.

In straight segments of the RBF, the mylonitic fabric defines a steep parallel-walled ductile shear zone containing a subhorizontal mineral lineation. The latter is a stretching lineation defined by dimensionally preferred orientations of hornblende, quartz, local feldspars, rare micas and flattened quartzofeldspathic augen, and is taken to record the direction of tectonic transport. However, major departures in this simple pattern occur in curved sections of the fault, such as those in the vicinity of the Pleasant Hills Pluton and Folly River (PHP and f; Fig. 1). Here the RBF develops a number of structural complexities. The patterns are described in the following section from east to west for well-exposed portions of the central and western RBF. Representative data for each of these sub areas are presented in Figures 2 and 3.

In the Debert River area (sub area A, Fig. 2) ductile shear associated with the RBF is contained entirely within the Hart Lake-Byers Lake Pluton, as several hundred metres of undeformed Hart Lake-Byers Lake granite occurs south of the granite ultramylonite that defines the shear zone. In this area the mylonitic foliation trends 240° to 270° and dips steeply south or southeast at angles greater than 60°. The associated mineral lineation defines an east-west trend on mylonitic foliation surfaces and plunges gently east or southeast at angles no greater than 30° (Fig. 3a). The orientation of these fabric elements indicates fault movement to have been primarily strike-slip with only a minor component of NW-vergent oblique-slip as would be expected in areas where the fault follows a straight course and is confined to a single well-defined parallel-walled zone.

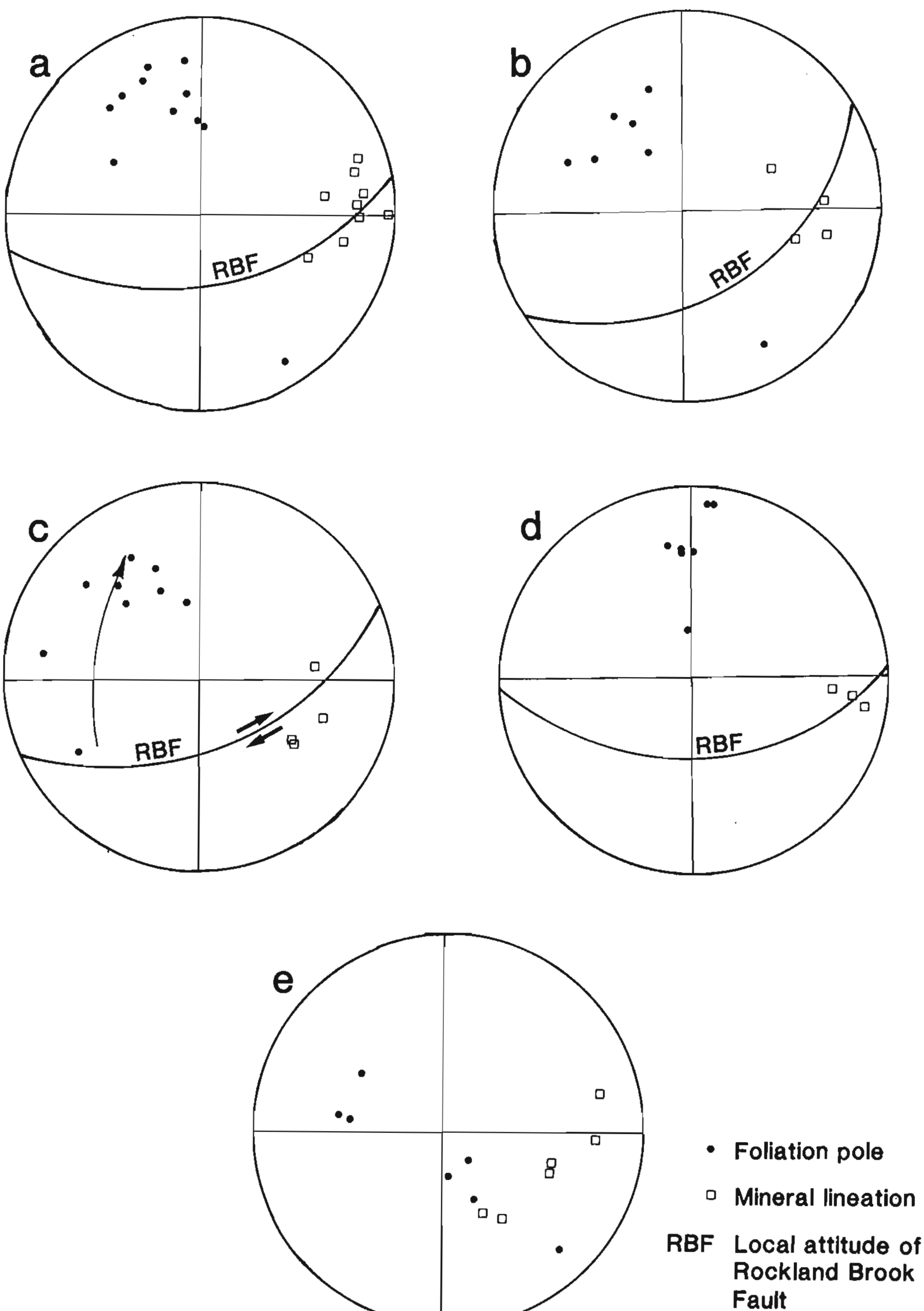

Figure 3. Equal-area stereoplots of structural data from subareas A through E on the E-W segment of the Rockland Brook Fault. See Figure 2 for subarea locations.

In the vicinity of Folly Lake (sub area B, Fig. 2), the RBF separates the massive white quartzites of the Gamble Brook Formation from diorites of the Folly Lake Pluton. Both rock types become increasingly deformed as the centre of the fault zone is approached. However, while the fabric in the Gamble Brook Formation trends from 045° to 075° and dips very steeply southeast, the mylonitic foliation in the diorite north of the fault zone progressively changes from a steeply southeast-dipping (greater than 60°) 030° trend at the fault zone centre to a gently dipping (less than 30°) 060° trend toward its northern margin (Fig. 3b). The fault zone, in this area, defines a curved trace concave to the south, and is interpreted to reflect the development of a half-flower structure (Fig. 4) as a result of dextral transpression within a restraining bend in the fault trace.

West of Rockland Brook (subarea C, Fig. 2), fabric elements within the Great Village River Gneiss are seen to progressively rotate into parallelism with the RBF as the fault zone is approached. This clockwise rotation of fabric elements is indicative of dextral shear (Fig. 3c).

The structural geometry of the RBF in the Portapique River area (subarea D, Fig. 2) is similar to that near Debert River in that it is a cross-strike section of the fault in an area where the fault trace is straight. Here, however, the ductile shear zone separates Carboniferous granite to the north from previously deformed Great Village River Gneiss to the south. The axial surfaces and hinge lines of associated folds in the gneisses are respectively subparallel to the mylonitic foliation and the mineral lineation. Closer to the centre of the shear zone, the earlier gneissic foliation is rotated and overprinted by the mylonitic fabric. North of the fault the rocks in this area were previously mapped as Silurian Wilson Brook Formation (Donohoe and Wallace, 1982). However, our mapping has shown that the Pleasant Hills Pluton continues farther east and contains the majority of the RBF deformation. The mylonitic fabric of the deformed granite strikes broadly east-west, dips steeply south and contains a gentle east-plunging (less than 30°) mineral lineation (Fig. 3d). As with the Debert River area, this portion of the RBF is also interpreted to lie within a straight section of the fault where motion was predominantly of dextral strike-slip geometry with only a minor component of NW-vergent oblique-slip movement.

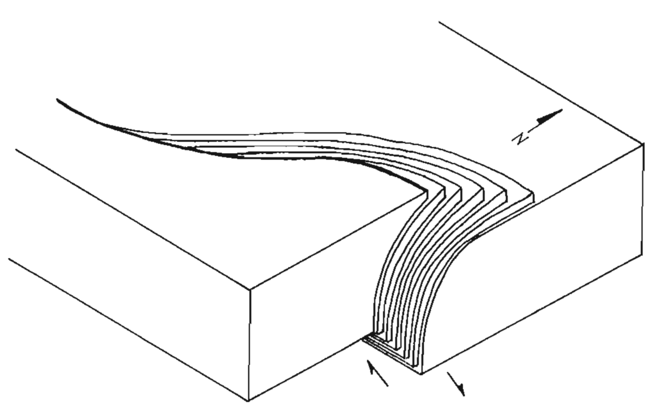

Figure 4. Proposed geometry of half-flower structure developed in the restraining bend on the Rockland Brook Fault near Folly Lake. No scale implied.

West of Bass River (subarea E, Fig. 2) the RBF separates the Great Village River Gneiss from a previously unmapped dioritic intrusion and takes a sharp southwesterly swing to eventually merge with the Cobequid Fault (Fig. 1). Unlike the Folly Lake area, however, this bend was sharp enough for dextral shear to produce low-angle NW-directed thrusting. As a result, the mylonitic fabric in the previously undeformed diorite trends between 010° and 020°, dips at moderate to gentle angles, and contains a down dip mineral lineation that plunges southeast at 36-50° (Fig. 3e).

An interpretative block diagram of the central RBF illustrating possible relationships between the variations in structural geometry described here is shown in Figure 5. Additional structural complexities include the newly recognized splays, extensions and offsets that mark the RBF's trace in the Pleasant Hills Pluton area (Fig. 1). The structural framework relating these to the kinematic history of the RBF has yet to be established. However, the revised position of the RBF in this area separates sample localities used in determining the age of the Pleasant Hills Pluton (Cormier, 1980; Donohoe and Wallace, 1982). Hence, the age obtained (315±25 Ma, Rb/Sr; Cormier, 1980) may no longer be meaningful as the granites north and south of the fault may have belonged to separate plutons now juxtaposed by movement on the RBF.

STRUCTURAL STYLE OF THE E-W AND NE-SW FAULT SEGMENTS

As defined by Murphy et al. (1988), the RBF is one of the few northern Appalachian faults to show both E-W and NE-SW segments. Yet the structures associated with the two fault orientations differ greatly in style. In the E-W section of the RBF (i.e. from Economy River east to the Salmon River Pluton; Fig. 1) the fault is a major boundary separating polydeformed Precambrian rocks from undeformed Devono-Carboniferous intrusives and sediments. Deformation at the fault zone centre is generally marked by 10 to 25 m of ultramylonite which is often sufficiently intense that the protolith cannot be reliably identified on the basis of field observation. Deformation directly attributable to the RBF continues for up to a hundred metres in the previously undeformed rocks north of the fault zone centre. Deformation may extend for similar distances in the rocks to the south of the fault, but because these have a pre-existing polyphase structural history, it is difficult to positively identify fabrics as directly attributable to RBF deformation much beyond several tens of metres from the fault zone centre.

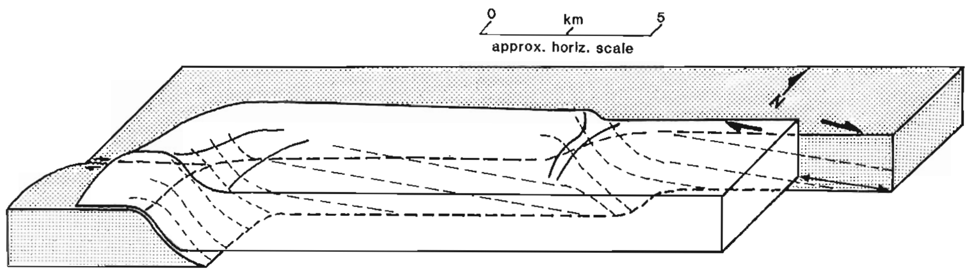

Figure 5. Interpretative block diagram illustrating the structural geometry of the central portion of the Rockland Brook Fault. No vertical scale or total slip implied. Horizontal scale is approximate. Arrow in lower right corner indicates relative movement vector only.

Outcrop and preliminary thin section evaluation of minor fold vergence, C-S fabrics and asymmetrical augen (sigma and delta structures of Passchier and Simpson, 1986) indicate that this section of the RBF is dominated by dextral movement, although kinematic indicators locally show the opposite sense of shear. However, in most cases these sinistral indicators can be explained in terms of antithetic shearing, local folding or complex ductile flow within an overall dextral regime. The conditions of deformation on this segment of the fault vary. In most localities quartz is ductilely deformed, while the feldspars remain relatively rigid. Under dry experimental conditions, this behavior occurs in granite at temperatures between $350\pm50°C$ and $600°\pm50°C$ over widely ranging pressures (Tullis and Yund, 1977). However, more intense conditions are also seen (e.g. subareas A and D, Fig. 2) where both minerals are ductilely deformed.

Within the Salmon River Pluton, Murphy et al., (1988) link the RBF to a previously unnamed fault which trends northeast along the course of Six Mile Brook (h; Fig. 1). This segment of the fault is marked by a quite different structural style and separates the Late Hadrynian Warwick Mountain Formation from the time-equivalent Dalhousie Mountain Volcanics (Murphy et al. 1988). The two units show different structural records but are lithologically very similar. The style of faulting in this area is largely brittle and offsets along individual fractures appear to be minimal. In some localities, volcanogenic sedimentary bedding can be seen at a high angle to the fault trace which suggests that strike-slip displacement along this section of the fault was less significant than that on the E-W portion. Significant post-Carboniferous deformation in the Six Mile Brook area is also precluded by the fact that tilted but otherwise undeformed beds of the Late Namurian Millsville Conglomerate (Gillis, 1964) are locally seen overstepping the predominant fracture trend in the Dalhousie Mountain Volcanics. Furthermore, the larger boulders in this fanglomerate comprise undeformed granite that is very similar to that of the Salmon River and other Devono-Carboniferous plutons of the Cobequid Highlands which are strongly deformed along the E-W segment of the RBF. Hence, late Paleozoic dextral movement along the NE segment of the fault is likely to have been minimal.

Nevertheless, the fault marks the trace of a structural front separating a late Precambrian volcano-sedimentary succession with a pervasive flat-lying Precambrian cleavage (Pe-Piper and Piper, 1987) to the NW, from a lithologically similar but undeformed late Precambrian sequence to the SE. In addition, the fault trace broadly corresponds to the northern margin of a linear positive Bouguer anomaly (Keppie, 1979) that suggests the basement is upthrown on its southeastern side. In view of these relationships, the NE-trending segment of the fault is considered to be that of the ancestral RBF and is tentatively interpreted as a SE-dipping thrust which separates the relatively undeformed Dalhousie Mountain Volcanics from underlying cleaved volcanics of the Warwick Mountain Formation.

SUMMARY AND DISCUSSION

The RBF is recognized as a significant structure within the Cobequid Highlands of Nova Scotia and one of the few in the northern Appalachians to show both E-W and NE-SW trends. In its E-W segment it shows appreciable late Paleozoic movement and lies parallel and adjacent to the Cobequid Fault which, in the Cobequid Highlands, marks the boundary between the Avalon and Meguma terranes. Here the RBF is a major ductile shear zone of dextral sense that separates units of very different lithology and age. In its NE-SW segment, however, it shows minimal late Paleozoic movement but is likely to have undergone pre-Carboniferous displacement. Hence this segment may record the early history of the RBF and is marked by a linear gravity anomaly suggesting the southern side of the fault is upthrown. Along this segment the RBF is predominantly a brittle structure and separates late Precambrian successions that are lithologically similar but which are structurally distinct and have been respectively correlated with widely separate units in the western Cobequid and Antigonish highlands (Murphy et al., 1988). This ancestral RBF is here tentatively thought to mark the trace of a Precambrian thrust ramp along which late Precambrian rocks of the Antigonish Highlands overthrust those of the Cobequid Highlands. If so, the Dalhousie Mountain Volcanics and Warwick Mountain Formation may never have been widely separated, supporting theconclusion of Nance and Murphy (in press) that, despite their present separation, broad correlations are likely to exist between the late Precambrian volcano-sedimentary successions of the Cobequid and Antigonish highlands. Hence, the ancestral RBF is more likely to mark a late Precambrian structural front rather than a major terrane boundary within the Avalon Composite Terrane as implied by Keppie (1985).

Along the present E-W portion of the fault, the ancestral RBF is considered to have provided a plane of weakness for later displacements that are here interpreted to have accompanied movement on the Cobequid Fault during the emplacement of the Meguma Terrane. This movement is thought to have reactivated the RBF and to have been responsible for the rotation of the fault's ancestral NE-SW trend into an E-W orientation. A similar scenario was implied by Keppie (1982) for the development of the Cobequid-Hollow Fault System. Further movement is thought to be responsible for the development of a presumed E-W link to the Cobequid Fault such as that proposed by Donohoe and Wallace (1982). This link would permit sufficient movement to produce the intense deformation now seen in the fault's E-W segment while preserving the ancestral RBF in the NE-SW segment. However, existing links appear to be late-stage brittle faults and may imply that the Salmon River Pluton is stitching with respect to late Paleozoic ductile shear on the RBF.

Cessation of major movement on the RBF is of uncertain timing. However, the mid-Carboniferous age implied by the truncation of the RBF by the Pleasant Hills Pluton (Donohoe and Wallace, 1982) can no longer be accepted, as the recognition of ductile shear zones within this body disqualifies it as a stitching pluton. However, right-stepping en echelon folds affect rocks of Late Visean to Early Namurian age between the Cobequid and Londonderry faults. Movement on the Cobequid Fault and, hence, related movement on the RBF may therefore have continued until at least Early Namurian time. Rare mafic dykes which cut the mylonitic fabric on Bass River, are likely to be of Triassic age.

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Precambrian rocks of the eastern Cobequid Highlands, Nova Scotia[1]

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Murphy, J.B., Pe-Piper, G., and Nance, R.D., Precambrian rocks of the eastern Cobequid Highlands, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 15-21, 1989.

Abstract

In the eastern Cobequid Highlands, Precambrian rocks south of the Rockland Brook Fault (RBF) consist of basement gneisses and amphibolites, structurally overlain by mid Rhiphean (?) platformal clastics (Gamble Brook Formation). A ductile shear zone, dated by a ca. 630 Ma intrusion of syntectonic granite and attributed to sinistral transtension, separates the gneisses from the Gamble Brook Formation. This transtensional environment may herald the deposition of the 630-600 Ma Folly River Formation which unconformably overlies the Gamble Brook Formation and consists of low-grade continental, rift-related mafic volcanics, abundant feeder dykes and distal turbidites. These rocks were probably deposited in a narrow rift basin within a volcanic arc environment. North of the Rockland Brook Fault the late Precambrian Jeffers Group and Warwick Mountain Formation of mafic and felsic volcanics overlain by turbidites may have been deposited in the same basin. The ca. 600 Ma deformation of the Folly River Formation is attributed to dextral transpression which probably closed this rift basin.

Résumé

Dans la partie est des hautes terres de Cobequid, les roches précambriennes au sud de la faille de Rockland Brook sont constituées de gneiss et d'amphibolites de socle, recouverts structuralement de roches clastiques de plate-forme du Rhiphéen (?) moyen (formation de Gamble Brook). Une zone de cisaillement ductile, datée par une intrusion de granite syntectonique d'environ 630 Ma et attribuable à une transtension sénestre, sépare les gneiss de la formation de Gamble Brook. Ce milieu de transtention pourrait annoncer la mise en place de la formation de Folly River vieille d'environ 630 à 600 Ma qui recouvre de façon discordante la formation de Gamble Brook et qui se compose de roches volcaniques mafiques de nature continentale et à faible degré de métamorphisme, liées à un fossé d'effondrement, de nombreux dykes nourriciers et de turbidites distales. Ces roches ont probablement été mises en place dans le bassin d'un rift étroit au sein d'un arc volcanique. Au nord de la faille de Rockland Brook, la mise en place du groupe Jeffers et de la formation de Warwick Mountain du Précambrien supérieur, constitués de roches volcaniques mafiques et felsiques recouvertes de turbidites, aurait eu lieu dans le même bassin. La déformation d'environ 600 Ma de la formation de Folly River est attribuable à une transpression dextre qui a probablement fermé ce bassin de sédimentation lié à un fossé d'effondrement.

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INTRODUCTION AND GEOLOGICAL SETTING

The Cobequid Highlands comprise Precambrian to Early Carboniferous rocks unconformably overlain to the north and east by Late Namurian to Early Westphalian clastic sediments (Donohoe and Wallace, 1982), and bounded to thesouth by the Cobequid Fault (Fig. 1). The Cobequid Highlands lie within the Avalon "Composite" Terrane of the northern Appalachians (Williams and Hatcher, 1983; Keppie, 1985) which forms a distinctive tectonostratigraphic belt characterized by late Precambrian (ca. 650-550 Ma) volcano-sedimentary successions and cogenetic granitic plutons overlain by early Paleozoic platformal successions that contain Acado-Baltic fauna (Williams, 1979; Keppie, 1985). The diversity of the late Precambrian volcano-sedimentary successions and the similarity of the lower Paleozoic histories, suggests amalgamation of the Avalon Terrane was completed by the latest Precambrian, an event that has been attributed to the Avalonian (e.g. O'Brien et al., 1983) or Cadomian (e.g. Keppie, 1985) Orogeny.

PREVIOUS WORK

Previous work in the eastern Cobequid Highlands, discussed in detail by Murphy et al., (1988), is only summarized here. Kelley (1966) produced preliminary maps of the Cobequid Highlands. H.V. Donohoe, Jr. and P.I. Wallace published maps of the entire Cobequid Highlands at a 1:50 000 scale together with a series of reports and field guides (Donohoe and Wallace, 1982, 1985 and references therein). Most of the units described below were originally defined in these reports.

STRATIGRAPHY

The stratigraphy of the eastern Cobequid Highlands (Table 1) has been described by Cullen (1984), Donohoe and Wallace (1985) and Murphy et al., (1988) and is only briefly discussed here. Lithologies vary considerably across the Rockland Brook Fault (RBF), the boundary between the northern and southern highlands. The Great Village River Gneiss, the Mount Thom Complex, the Dalhousie Mountain

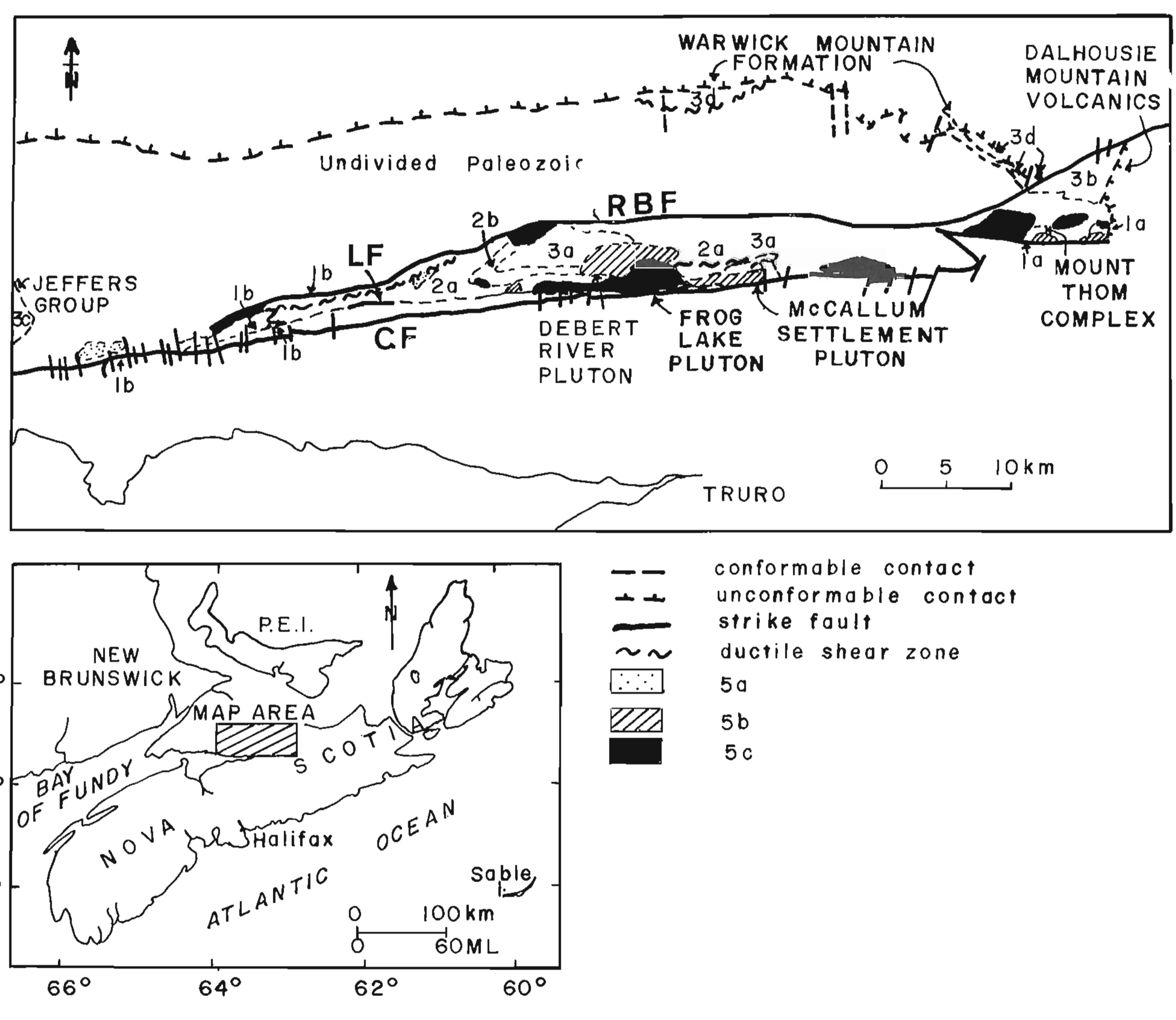

Figure 1. Preliminary map showing the distribution of Precambrian lithologies (see table 1) in the eastern Cobequid Highlands. The location of the eastern Cobequid Highlands is in the inset. Abbreviations used on the map are as follows: CF, Chedabucto Fault; RBF, Rockland Brook Fault; LF, Londonberry Fault.

Table 1. Precambrian stratigraphy of the eastern Cobequid Highlands (modified after Donohoe and Wallace, 1982).

North of RBF	South of RBF
	5c appinitic gabbro, diorite 5b granite, granodiorite 5a granite gneiss
3c Jeffers Group 3d Warwick Mountain Formation	3a Folly River Formation 3b Dalhousie Mountain Volcanics
	2 Gamble Brook Formation 2a orthoquartzite 2b phyllite
	1a Mount Thom Complex 1b Great Villager River Gneiss

Volcanics and Hadrynian plutons occur exclusively to the south of the fault. The late Precambrian Jeffers Group and lithologically similar Warwick Mountain Formation occur only north of the fault. The oldest rocks that unequivocally occur on both sides of the RBF are Devono-Carboniferous plutons.

Precambrian rocks south of the Rockland Brook Fault

Orthogneisses, paragneisses and schists of the Mount Thom Complex (unit 1a) and the Great Village River Gneiss (unit 1b), commonly thought to be the oldest rocks in the Cobequid Highlands, form an amphibolite facies basement structurally beneath the Gamble Brook sequence of platformal sediments (Donohoe and Wallace, 1985). However, the contact between these units is a ductile shear zone that obscures the original relationships (Murphy et al., 1988), so that a basement-cover relationship cannot be confirmed. The Mount Thom Complex and the Great Village River Gneiss occur only in the eastern highlands in a narrow belt between the Londonderry and Rockland Brook faults (Fig. 1). These probably correlative (Donohoe, 1983) complexes form part of the Precambrian metamorphic infrastructure of the Avalon Terrane. Predominantly quartzofeldspathic orthogneisses of the Mount Thom Complex contain a foliation defined by biotite, garnet and muscovite. The Great Village River Gneiss consists of massive to quartz-plagioclase-layered hornblende amphibolites (the Bass River amphibolite of Cullen, 1984) hornblende-bearing quartzofeldspathic orthogneisses (Great Village River orthogneiss of Cullen, 1984), and biotite-garnet psammitic paragneisses (Donohoe and Wallace 1985). A poorly constrained 934 $\pm$ 82 Ma age derived from the Mount Thom Complex (Rb-Sr whole rock, Gaudette et al., 1984) has been interpreted as the age of the D_1 fabric. The ages of the Great Village River Gneiss and Mount Thom Complex are currently being investigated by R. Doig of McGill University using the U-Pb method.

The Gamble Brook Formation (unit 2), the older of two "cover" sequences to the Great Village River Gneiss can be subdivided into two distinct units (Murphy et al., 1988). The lower part (unit 2a) contains abundant orthoquartzites and arkosic quartzites with interlayered biotite-muscovite-garnet psammitic schists and minor carbonate beds or lenses. The upper part (unit 2b) of the Gamble Brook Formation, distinctly more pelitic, is dominated by biotite-muscovite and biotite-garnet schists with minor quartzite and psammitic schists. The orthoquartzite protolith of much of this formation and the association of thin carbonates and overlying pelitic horizons suggest a platformal setting of deposition. The depositional age of the Gamble Brook Formation is unknown. In southern New Brunswick (Donohoe and Wallace, 1985) the Brookville Gneiss is overlain by orthoquartzites and marbles (Green Head Group) that contain stromatolites of mid-Riphean age (Hoffman, 1974). If a correlation between the New Brunswick and lithologically similar Cobequid successions can be made, it suggests a depositional age for the Gamble Brook Formation and a minimum mid-Riphean age for the Great Village River Gneiss (Donohoe and Wallace, 1985). A mylonitic fabric and a well-developed mineral lineation occur adjacent to the contact with the Great Village River Gneiss. In these localities, bedding is transposed into the metamorphic foliation. The intensity of the metamorphic fabric decreases away from the contact and bedding can be recognized.

The Folly River Formation (Fig. 1, unit 3a), interpreted to unconformably overlie the Gamble Brook Formation (Murphy et al., 1988), consists of greenschist facies mafic flows, abundant feeder dykes, hyaloclastites and tuffs, chloritic distal turbidites and minor jasperitic ironstones. The upper contact of the Folly River Formation is not exposed. Lavas show a variety of igneous textures, including ophitic to subophitic, pilotaxitic, porphyritic and vesicular, and consist of plagioclase and actinolite $\pm$augite, $\pm$hornblende, Fe-Ti oxides, chlorite, epidote, biotite and rare quartz. Dykes within the Folly River Formation have a similar mineralogy and display an ophitic to subophitic texture. The basalts and dykes are Fe-Ti-rich, differentiated within-plate continental tholeiites (Pe-Piper and Murphy, in press), suggestive of emplacement during continental rifting. Deposition of the Folly River Formation postdated the $\sim$630 Ma shear zone deformation in the Gamble Brook Formation (Nance and Murphy, in press). Cleavage in the Folly River Formation is heterogeneously developed, particularly in the mafic volcanic rocks and dykes which may vary locally from massive to schistose. A minimum age for the formation is given by the age of the Debert River Pluton (596$\pm$70 Ma, Donohoe et al., 1986, Rb-Sr whole rock) which posttectonically intrudes the Folly River Formation. Deposition of the Folly River Formation may be coeval with that of the volcano-sedimentary Jeffers Group and Warwick Mountain Formation north of the RBF.

The Dalhousie Mountain Volcanics, which occur only in the easternmost part of the area south of the RBF (Fig. 1, unit 3b) comprise interlayered greenschist to subgreenschist mafic flows, felsic tuffs and finely laminated turbidites. Cleavage is heterogeneously developed and is penetrative only in the turbidites. Lithologically and geochemically they strongly resemble the late Precambrian Keppoch Formation of the Antigonish Highlands, from which they are separated by the Devono-Carboniferous Stellarton Gap (Yeo and Gao, 1986), the Jeffers Group (Pe-Piper and Piper, 1987) and the Warwick Mountain Formation of the northern Cobequid Highlands (Murphy et al., 1988). These units are here correlated, although the lack of a pervasive subhorizontal cleavage in the Keppoch Formation and Dalhousie Mountain Volcanics suggests that these units lie within different thrust slices than the Jeffers Group and Warwick Mountain Formation.

Precambrian rocks north of the Rockland Brook Fault

The Jeffers Group (unit 3c) and the lithologically similar Warwick Mountain Formation (unit 3d), described in detail by Pe-Piper and Piper (1987), consist of interlayered felsic and mafic volcanic rocks and turbidites. All rocks characteristically display a ubiquitous and penetrative flat-lying cleavage. Lithologically they strongly resemble the late Precambrian rocks in the Antigonish Highlands. The Jeffers Group is post-tectonically intruded by late Precambrian plutons in the western highlands (Pe-Piper and Piper, 1987). The Jeffers Group contains Fe-Ti-rich continental tholeiitic basalts and a calc-alkalic basalt to rhyodacite suite (Pe-Piper and Piper, in press).

STRUCTURE

We recognize three major deformational episodes (D_1, D_2 and D_3) of Precambrian age. Evidence of D_1 is found only in the Great Village River Gneiss. D_2 is divided into D_{2a} and D_{2b} which are thought to represent phases of a single progressive deformational event that resulted in the development (D_{2a}) and folding (D_{2b}) of the mylonitic fabric associated with the ductile shear zone between the Gamble Brook Formation and the basement. Deformation of the Folly River Formation is kinematically distinct from D_2 (Nance and Murphy, in press) and assigned to a younger polyphase (D_{3a} and D_{3b}) event.

D_1 Structures

Evidence of the earliest phase of deformation (D_1) is confined to basement gneisses where the earliest recognizable folds (to which the D_{2a} mylonitic fabric is approximately axial planar) rarely deform an older metamorphic fabric (Fig. 4 in Donohoe and Wallace, 1985). Elsewhere, the S_1 fabric has been completely overprinted by that of D_{2a}. Where visible, S_1 is an amphibolite facies foliation defined by compositional banding in the gneisses. D_1 folds and linear fabrics have not been observed and the age and significance of the deformation is unknown. The correlation proposed by Donohoe (1983) between the Great Village River Gneiss and the Mount Thom Complex would imply a minimum age for D_1 of about 900 Ma based on the available age data (Gaudette et al., 1984). However, given the uncertainties in this age and correlation, it is also possible that D_1 records the earliest phase of the progressive deformation that later produced D_2. In this case the existence of an unconformity between the Great Village River Gneiss and the Gamble Brook Formation could no longer be argued on structural grounds although their obvious lithological contrasts would remain. However, folds to which the mylonitic fabric is approximately axial planar in the Gamble Brook Formation deform only the bedding. Alternative interpretations cannot be evaluated until the results of geochronological work become available.

D_2 Structures

The Great Village River Gneiss contains moderately inclined, generally shallowly-plunging isoclinal folds of at least two generations (F_{2a} and F_{2b}) that are coaxial, coplanar and probably cogenetic (Nance and Murphy, in press). These folds deform a metamorphic fabric and are thought to be related to the second phase of deformation. The earliest of D_2 generation folds (F_{2a}) contain an axial planar mylonitic fabric and later D_2 folds (F_{2b}) deform this fabric. Fold axes vary locally indicating that sheath folds may exist.

The present contact between the Great Village River Gneiss and the structurally overlying Gamble Brook Formation (unit 2) is interpreted to be a ductile shear zone (Murphy et al., 1988) associated with D_2. The shear zone is characterized by intense local mylonitization, the development of an LS mylonitic fabric, tectonic interleaving of the Great Village River Gneiss and Gamble Brook metasediments, syntectonic intrusion of granite gneiss, development of C-S fabrics and at least two generations of cogenetic small-scale isoclinal folds (F_{2a} and F_{2b}). The southeasterly-dipping mylonitic fabric and east- to southeast-plunging mineral lineation are the oldest recognizable tectonic fabrics in the Gamble Brook Formation and deformation associated with their development is assigned to D_{2a}. The deformation was associated with greenschist to amphibolite facies metamorphism. Spatially associated coaxial and coplanar folds deform the mylonitic fabric, are probably the result of progressive deformation and are assigned to D_{2b} (Nance and Murphy, in press). The ca. 630 Ma age of the syntectonic granite gneiss (Gaudette et al., 1984) is interpreted to be the age of this shear zone. This age is supported by the presence of xenoliths of quartzite containing a mylonitic fabric in the late Precambrian Debert River Pluton. The sense of shear implied by all indicators for D_{2a} and D_{2b} consistently suggests oblique slip, with normal and sinistral components toward the present southeast (Fig. 2a,b). It is interpreted as a transtensional deformation within a sinistral strike-slip regime (Nance and Murphy, in press).

D_3 Structures

The deposition of the Folly River Formation postdates the development of the ca. 630 Ma shear zone fabric in the underlying Gamble Brook Formation (Murphy et al., 1988). However, the Folly River Formation is posttectonically intruded by the 596±70 Ma Debert River Pluton (Gaudette et al., 1984) indicating that deformation of the Folly River Formation occured in late Precambrian time. This deformation (D_3, Nance and Murphy, in press) produced (D_{3a}) and

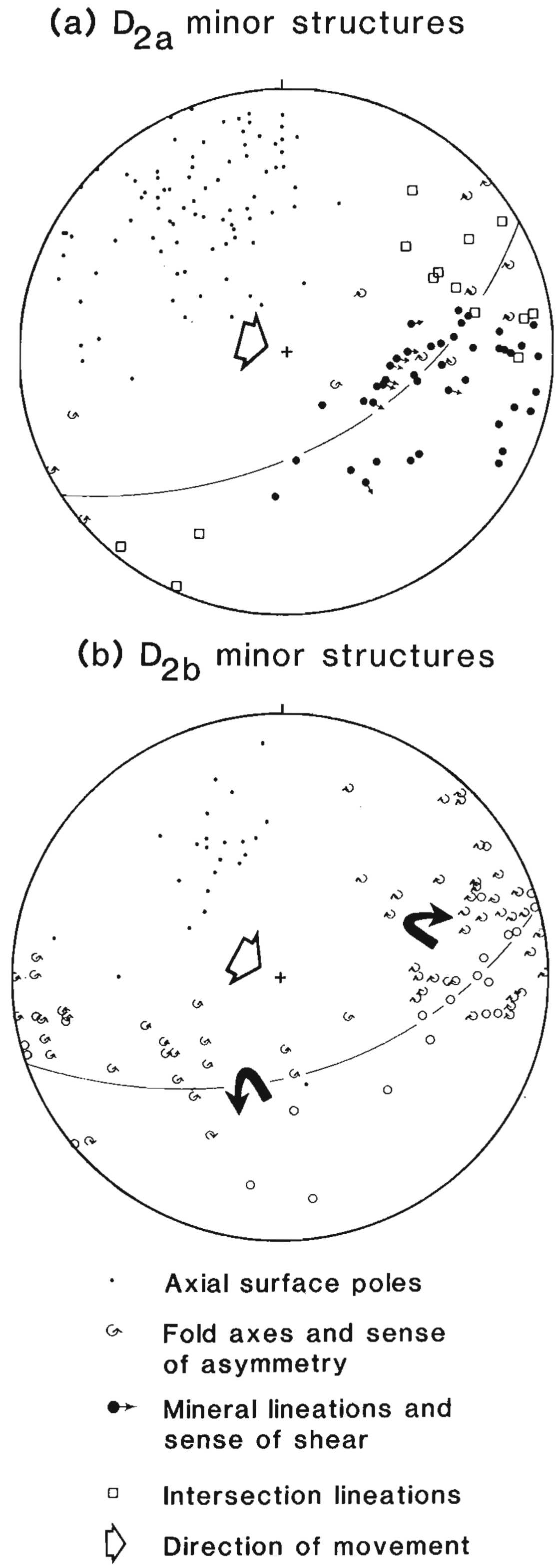

Figure 2. Synoptic equal-area stereographic projections of D_2 structural elements in the Great Village River Gneiss, Gamble Brook Formation and syntectonic granite gneiss bodies within the ductile shear zone. Large arrows indicate the overall sense of asymmetry of F_2 folds, and sense of shear indicators imply oblique slip with normal and sinistral components (after Nance and Murphy, (in press).

later folded (D_{3b}) a low-grade, generally southeasterly-dipping schistosity (S_{3a}) that is axial planar to tight asymmetrical folds (F_{3a}), a mineral lineation (L_{3a}) defined by elongation of low-grade minerals that plunges gently east-southeast (Fig. 3a) and defines a stretching lineation that is considered to record the direction of D_3 transport and widespread asymmetrical folds. F_{3b} folds can be distinguished from F_{3a} since they deform the S_{3a} fabric. F_3b is coaxial and coplanar with F_{3a} (Fig. 3b) and F_3 is interpreted to be the result of progressive deformation. The sense of movement implied by the distribution of F_{3a} and F_{3b} fold symmetry is locally consistent with that implied by C-S fabrics and asymmetrical epidote augen. The sense of shear implied by this kinematic data is that of oblique northwest-directed slip and is interpreted to be related to dextral transpression (Nance and Murphy, in press).

PLUTONIC ROCKS

Precambrian-Early Cambrian plutonic rocks (unit 5) consisting of granite gneiss, granite-granodiorite and appinitic gabbro and diorite occur exclusively to the south of RBF. Granite gneiss bodies (unit 5a), ranging in size from veins and dykes to narrow plutons, occur within the ductile shear zone between the Great Village River Gneiss and the Gamble Brook Formation. The gneisses consist of sericitized porphyroclasts of orthoclase, plagioclase and quartz with rare biotite and muscovite. They contain a mylonitic fabric and a mineral lineation. Contacts with country rocks are welded and concordant on a regional scale, but local crosscutting contacts are also preserved. In contrast to their host rocks, the intensity of mylonitization of the granite gneiss varies considerably over short distances. Larger bodies commonly crosscut the principal (D_2) deformational fabric yet show well-developed C-S fabrics and asymmetrical augen that are coplanar with the stronger fabric of their host rock. These relationships coupled with the available geochronological data (642 ± 15 Ma and 626 ± 22 Ma, Gaudette et al., 1984, Rb-Sr whole rock) imply a probable late Precambrian age and support Cullen's (1984) contention that the emplacement of the granite gneisses was broadly syntectonic with respect to D_2.

Late Precambrian granite and diorite plutons (units 5b and 5c) include the Frog Lake, McCallum Settlement and Debert River plutons (Fig. 1) The Frog Lake Pluton, an appinitic diorite complex of small stocks and sills intruding the Gamble Brook and Folly River formations, consists mainly of plagioclase and amphibole in varying proportions with interstitial biotite, quartz and alkali feldspar. Most specimens are equigranular, but some are porphyritic and a few show an ophitic texture. The pluton resembles the late Precambrian Jeffers Brook diorite in the western Cobequid Highlands (Pe-Piper and Piper, 1987). Pegmatitic granite veins within the pluton are spatially associated with the digestion of xenoliths. The Frog Lake Pluton is cut by probable late Precambrian granites of the Debert River Pluton to the northwest and the McCallum Settlement Pluton to the east. These relationships provide a minimum age for the Frog Lake Pluton.

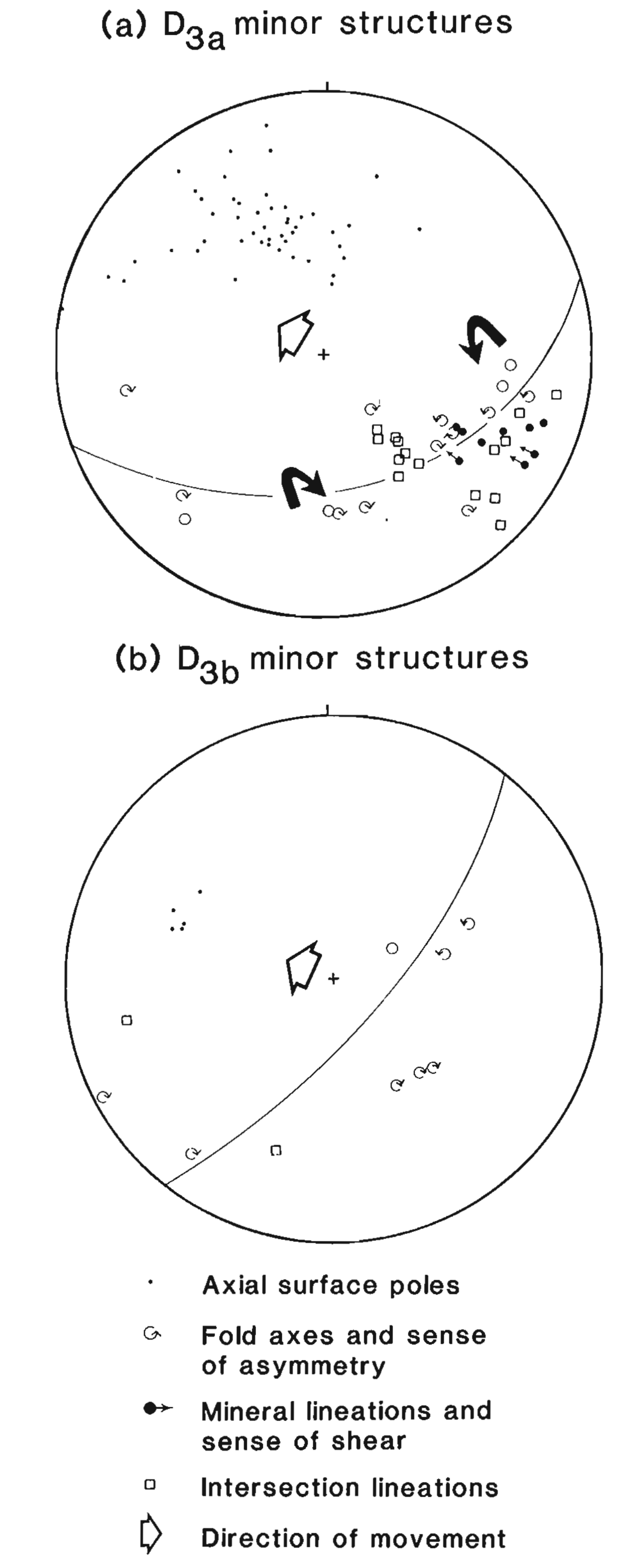

Figure 3. Synoptic equal-area stereographic projections of D_3 structural elements within the Folly River Formation. Large arrows indicate the overall sense of asymmetry of F_3 folds. The sense of shear indicators imply oblique slip with reverse and dextral components toward the WNW (after Nance and Murphy, in press).

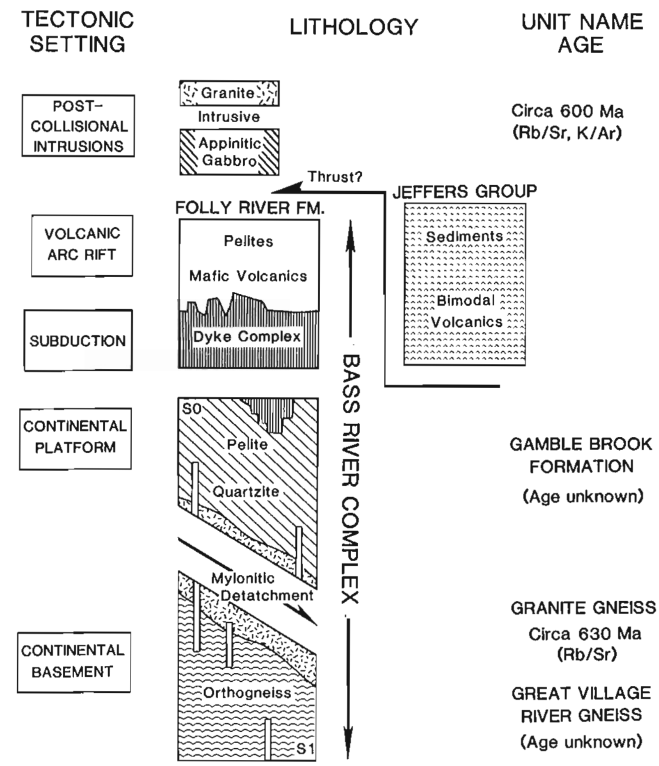

Figure 4. Interpretive tectonostratigraphic column for the evolution of the Precambrian rocks in the eastern Cobequid Highlands.

The Debert River and McCallum Settlement plutons consist of granite and granodiorite, and granite, granodioirite and tonalite respectively. The Debert River body contains abundant xenoliths of mylonitic quartzite thought to be derived from the Gamble Brook Formation, and is cut by stocks of Carboniferous granites. Both bodies appear to be relatively high level with common roof pendants, and consist of medium to coarse equigranular subhedral quartz, plagioclase (An_{30-45}), orthoclase and microcline, with minor biotite (commonly replaced by chlorite), muscovite, hornblende, epidote, opaque oxides, apatite and accessory allanite and zircon. Preliminary geochemical data indicate that they belong to a compositionally expanded calc-alkalic suite displaying volcanic arc affinities on trace element discrimination diagrams.

SUMMARY AND DISCUSSION

An interpretative tectonostratigraphic column is shown in Fig. 4. The ages of the Great Village River Gneiss and Mount Thom Complex are poorly constrained, but Gaudette et al., (1984) interpreted a 934 $\pm$ 82 Ma Rb-Sr whole rock age from the Mount. Thom Complex as a metamorphic age. A minimum age of ca. 630 Ma is provided by granite gneisses which intrude along the contact between the Great Village River Gneiss and the Gamble Brook Formation. The Great Village River Gneiss could be basement to the Gamble Brook Formation, or could alternatively be a deep level equivalent of the late Precambrian volcanic complexes that characterize the Avalon Terrane. U-Pb dating presently in progress will test these hypotheses.

The Gamble Brook Formation is interpreted to be a platformal sequence (Fig. 4), tentatively assigned a mid-Rhiphean depositional age (Donohoe, 1983) on the basis of a correlation with the Green Head Group of New Brunswick. The ca. 630 Ma granite gneiss provides a minimum age. A maximum age can be derived from the age of detrital zircons in the Gamble Brook Formation currently under study.

Deformation at ca. 630 Ma produced a ductile shear zone (Fig. 4) associated with sinistral transtension which may have produced a continental rift environment (Fig. 4) appropriate for deposition of the Folly River Formation whose stratigraphy and geochemistry suggest that it was formed in a narrow pull-apart basin with limited attenuation of continental crust (Pe-Piper and Murphy, in press). The Folly River Formation is broadly correlative with other volcano-sedimentary sequences such as the Jeffers Group of the northern Cobequid Highlands and the Georgeville Group of the Antigonish Highlands which contain abundant calc-alkaline lavas with volcanic arc affinities in addition to rift-related basalts and turbidites (Murphy et al., in press; Pe-Piper and Piper in press). This suggests that the late Precambrian tectonic environment in the Cobequid Highlands may be similar to that of the better exposed Antigonish Highlands where arc-related mafic and felsic lava, interlayered continental tholeiitic basalts and turbidites which occur in the northern and southern highlands are approximately coeval with continental tholeiites and distal turbidites in the central highlands. The occurrence of the late Precambrian arc-related Dalhousie Mountain Volcanics and plutons with volcanic arc affinities to the south of the Rockland Brook Fault (i.e. in the same fault block as the Folly River Formation) supports the interpretion that the Folly River Formation was deposited in an ensialic rift within a volcanic arc environment.

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Aeromagnetic total field, gradiometer, and VLF-EM survey of part of the Dunnage Zone, central Newfoundland[1]

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Tod, J. and Ready, E.E., Aeromagnetic total field, gradiometer, and VLF-EM survey of part of the Dunnage Zone, central Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 23-27, 1989.

Abstract

An aeromagnetic total field, gradiometer, and VLF-EM survey was carried out over part of the Dunnage Zone in central Newfoundland. This portion of the Dunnage Zone is geologically complex and well suited to mapping with aeromagnetics. Aerodat Limited of Toronto carried out the contract in 1986-1987, using a rotary wing aircraft.

Survey results are available at two map scales: 1:25 000 aeromagnetic total field and gradiometer contour maps and 1:50 000 magnetic anomaly and gradiometer colour interval maps. The VLF-EM total field and quadrature profiles are printed on the back of the 1:50 000 maps.

The aeromagnetic results substantiate and expand on the known geology in this part of the Dunnage Zone. The gradiometer colour maps in particular, show clearly defined intrusive features and major transcurrent faults. A series of folded sills, as well as a thrust sheet, can be identified in the middle and northern regions of the survey area, respectively.

Résumé

On a effectué dans ue partie de la zone de Dunnage dans le centre de Terre-Neuve un levé combiné aéromagnétique du champ total, gradiométrique et électromagnétique à très basse fréquence (VLF-EM). Cette partie de la zone de Dunnage est géologiquement complexe et se prête bien à la cartographie par levé aéromagnétique. En 1986-1987, la société Aerodat Limited de Toronto a procédé, à l'aide d'un hélicoptère, à l'exécution du contrat.

Il est possible d'obtenir les résultats du levé à deux échelles cartographiques: les cartes gradiométriques et aéromagnétiques du champ total en courbes de niveaux, à l'échelle de 1/25 000, et des cartes gradiométriques et des anomalies magnétiques avec intervalles en couleurs, à l'échelle de 1/50 000. Les profils VLF-EM du champ total et les profils en quadrature sont imprimés au verso des cartes à l'échelle de 1/50 000.

Les données aéromagnétiques confirment et ajoutent aux connaissances géologiques de cette partie de la zone de Dunnage. Les cartes gradiométriques en couleurs font notamment ressortir des éléments intrusifs bien définis et les grandes failles de décrochement. On peut également reconnaître dans le centre et le nord de la région à l'étude une série de filons-couches plissés, ainsi qu'une nappe de charriage.

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

INTRODUCTION

Under the Canada-Newfoundland Mineral Development Agreement 1984-1989, an aeromagnetic total field, gradiometer and VLF-EM survey was carried out over an area within the Dunnage Zone of central Newfoundland. The location of the survey area is shown in Figure 1. The basement rocks in this area are structurally and compositionally complex with a distinct magnetic signature, making indirect geological mapping through aeromagnetics very successful.

Aerodat Limited of Toronto flew the survey between November 1986 and August 1987, using a rotary wing aircraft. Two Cesium vapour magnetometers, each with a resolution of 0.002 nanoteslas and separated by 3 m, measured the gradient field. VLF-EM total field and vertical quadrature components were recorded with a Herz Totem-2A receiver, using the Annapolis, Maryland transmitter signal. A mean elevation clearance of 150 m was maintained with a flight line separation of 300 m and a control line separation of 5 km.

Survey results are available at two map scales: 1:25 000 aeromagnetic total field and gradiometer contour maps, and 1:50 000 magnetic anomaly and gradiometer colour interval maps. The VLF-EM total field and quadrature profiles are printed on the back of the magnetic anomaly and gradiometer colour interval maps, respectively. This presentation provides a means to view the VLF-EM data jointly with the magnetic data with the use of a light table. In addition, the digital data can be purchased for further processing and interpretation, from the Geophysical Data Centre, Geological Survey of Canada, 1 Observatory Crescent, Ottawa, Ontario, K1A 0Y3 (telephone: (613)992-6438).

Figures 2 and 3 are photo reductions of the magnetic anomaly and gradiometer maps, respectively. Superimposed on each is an overlay emphasizing the major magnetic markers.

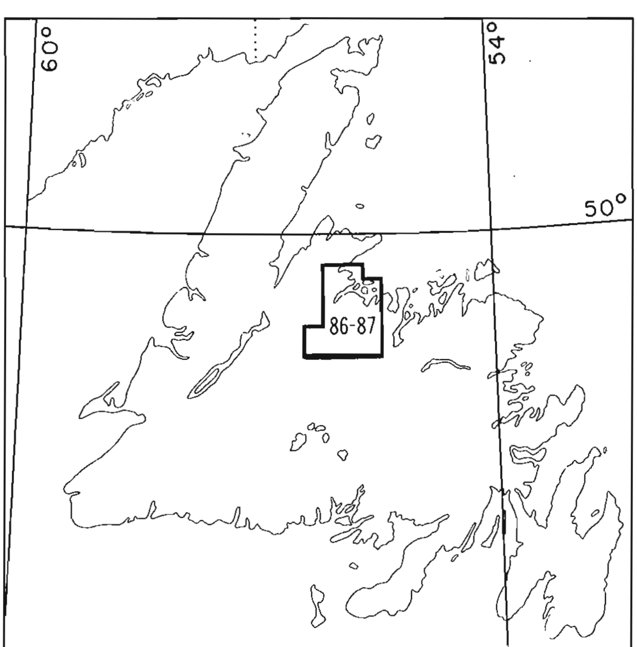

Figure 1. Location map of the study area.

RESULTS

In this central Newfoundland survey area, the magnetic signature not only confirms mapped geology (Dean, 1977) but also provides additional details. A number of structural features stand out clearly on the gradiometer colour maps. Several phases of plutonic intrusion (A, Fig. 2) have occurred in the southern portion of the survey area and these are transected by more recent northeast-striking faults. A folded swarm of sills occurs farther to the north (B). Near the top of the surveyed area there is evidence of a thrust sheet (C). Although all of these features have been previously identified, the specific information on each is enhanced by the geophysical results.

The igneous intrusions in the southern part of the survey area are of Devonian age (Dean, 1977) and are graphically outlined by the magnetics and less so by the VLF-EM. They are identified by a horseshoe-shaped pattern which characterizes the magnetic contrast along their perimeters. The Hodges Hill, Topsails and Dawes Pond granites as well as the Twin Lakes Diorite Complex contacts are particularly well defined (see Dean, 1977 for details on the geology of the area). Within certain intrusive units, there is a secondary magnetic signature indicating either magmatic differentiation or possibly, progressive stages of intrusion.

The gradiometer trace is particularly effective in identifying magnetic contacts, as is the case with these intrusions. The zero gradiometer contour tends to lie directly above a magnetic contact in northern latitudes. For Newfoundland, this contour will be slightly offset due to the inclination of the earth's field, the effect of remanent magnetization and the interference from adjacent sources. VLF anomalies are associated with both contrasts in conductivity between rock units and cultural features. This survey area is dotted with bodies of water, swamps, creeks, etc. which complicate the interpretation of the VLF-EM signature.

In the north the ophiolitic rocks of Lushs Bight Group are thrust faulted against the Silurian Springdale and Ordovician Robert's Arm groups (Kean, 1984). The magnetics trace this fault (Lobster Cove Fault) very clearly where the basaltic pillow lavas are thrust against the Robert's Arm Group to the south. The sharpness of this delineation decreases westward.

Middle and Lower Ordovician diabase and gabbro sills intrude the Wild Bight Group southwest of Badger Bay (see Dean, 1977). They appear magnetically as a cluster of several parallel to subparallel, folded elongated anomalies. The contrast in magnetic response between these units is such that folding patterns can be readily traced. The vertical gradient indicates that the number and extent of these folded, north-northeasterly-trending units is far greater than currently identified from field mapping.

The Topsails granite (see Dean, 1977) is located along the most westerly edge of the survey area. Just to the north are the volcanics of the Silurian Springdale Group which comprise alternately acidic and mafic units. The difference in magnetic signature between these alternating units, helps to define the Burnt Berry Syncline (see Dean, 1977) which trends to the northeast.

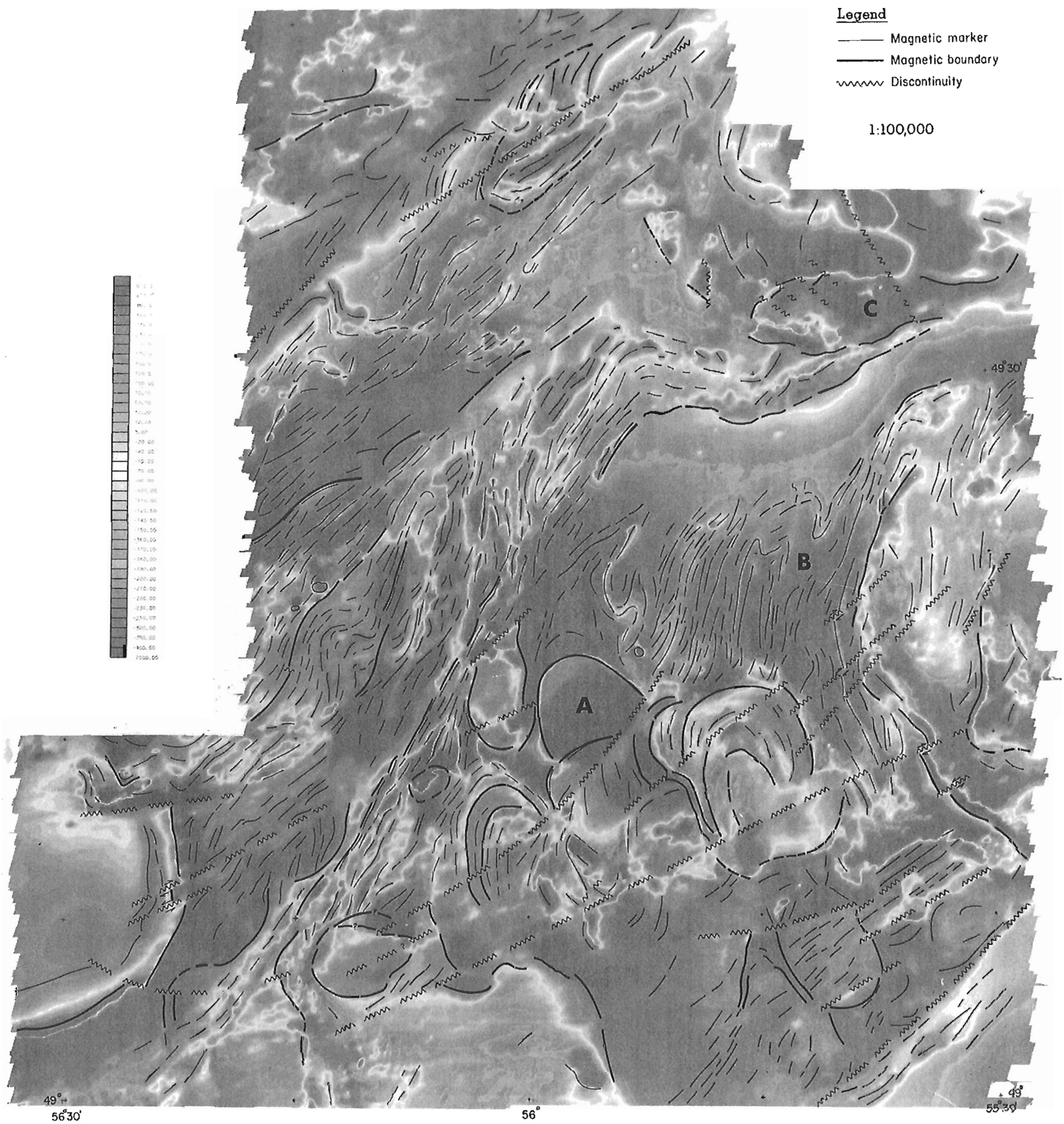

Figure 2. Magnetic anomaly map with major magnetic markers outlined — Central Newfoundland.

Figure 3. Vertical gradient map with major magnetic markers outlined — Central Newfoundland.

A post-Devonian period of major faulting is strongly evidenced in the magnetic data. These east-northeasterly trending faults such as Northern Arm, Long Pond and Four Mile (see Dean, 1977) show up clearly, particularly where they cross magnetic features at almost right angles. They are extensive, readily evident and in several cases, clearly indicate transverse displacement. Several previously mapped faults can be extended or modified on the basis of this new information.

The structural and lithological features that have been highlighted in this paper are only those that are most apparent on the 1:50 000 scale colour maps. Further qualitative details plus a great deal of quantitative information can be derived from the 1:25 000 scale contour maps and the profile data.

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A proposed genetic model for epigenetic Ba-Pb-Zn occurrences not associated with igneous rocks, Taconic Thrust Belt, Quebec[1]

K. Schrijver[2] and P. Rhéaume[2]

Schrijver, K. and Rhéaume, P., A proposed genetic model for epigenetic Ba-Pb-Zn occurrences not associated with igneous rocks, Taconic Thrust Belt, Quebec; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 29-37, 1989.

Abstract

The Cambrian St-Damase Formation in the St-Fabien area contains barite + galena ± sphalerite occurrences of which three are veins and one consists of veins and disseminations. In view of the setting of this vein and vein-associated mineralization, hosted in sedimentary rocks remote from igneous intrusives, the main objective of this paper is to contribute to a metallogenic model, emphasizing the driving forces of fluid circulation.

We propose that these forces were, from early to late, (1) sediment loading, (2) lateral compression, and (3) extension. Late in stage (1), Ba and Pb were released during the leaching of potassium feldspar. In (2), fluid circulation changed. In (3), a hydraulic gradient was established by regional uplift, causing the migration of metalliferous brines. Gravity-driven sulphate-bearing groundwater of marine origin, initially channeled along dilatant faults, mixed with these brines, and sulphate and sulphides were precipitated in secondary fracture and dissolution porosities, possibly in slightly post-Taconic time.

Résumé

La formation cambrienne de St-Damase dans la région de Saint-Fabien contient quatre venues de barytine + galène ± sphalérite, dont trois se présentent sous forme de filons et la quatrième sous forme de filons et de disséminations. Comme cette minéralisation est logée dans des roches sédimentaires éloignées de roches intrusives ignées, le but premier du présent article est de proposer un modèle métallogénique qui mette l'accent sur les forces d'entraînement des fluides minéralisants.

Ces forces se seraient manifestées durant trois périodes selon leur nature : (1) accumulation des sédiments, (2) compression latérale et (3) distension. Vers la fin de la période (1), du Ba et du Pb ont été libérés pendant le lessivage du feldspath de potassium. En (2), la circulation des fluides a changé. En (3), un soulèvement régional a établi un gradient hydraulique, entraînant la migration des saumures métallifères. Entraînées par gravité, des eaux souterraines sulfatées d'origine marine, initialement canalisées par des failles en dilatation, se sont mélangées à ces saumures, et des sulfates et des sulfures se sont déposés dans des fissures secondaires et des porosités de dissolution, sans doute peu après l'époque taconique.

[1] Contribution to the Canada Economic Development Plan for Gaspé and Lower St. Lawrence, Mineral Program 1983-1988. Project carried by the Geological Survey of Canada, Mineral Resources Division, Contract 23233-7-0224/01-SS.

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INTRODUCTION AND GEOLOGICAL SETTING

The St-Fabien area, 235 km northeast of Quebec City, forms part of the Taconic Orogen of the Appalachians, and comprises a conformable sequence of anchimetamorphic Lower Cambrian to Middle Ordovician sedimentary rocks, virtually devoid of igneous rocks (Figs. 1 and 2; Table 1). Structurally, the area is part of the external allochthonous domain of the Quebec Appalachians, characterized by a stack of imbricated nappes, separated by major thrust faults (Fig. 3) (St-Julien and Hubert, 1975; Vallières, 1984; Schrijver et al., in press).

The Des Seigneuries Nappe contains seven known Ba-Pb(-Zn) vein occurrences, all of which but one (Occurrence 1 in Fig. 1) are hosted in the St-Damase Formation. Four of these, as well as several sets of pink baritic veinlets devoid of sulphide (hosted in the grès verts unit and Orignal Formation), occur in the study area (Fig. 2). Detailed mapping (1:2000) of domains centred on the veins has shown that all occur in anticlinal hinges of major folds and are associated with faults that offset the axial planes (Schrijver and Rhéaume, 1988).

The veins are from several millimetres to 150 cm wide, generally steeply dipping and characterized by drusy textures. Excepting the pink baritic veinlets, they form the latest sets of the local system of calcite-quartz veins and veinlets. Moreover, the forms of the mineralized veins indicate that they were emplaced later than the ductile deformation and commonly also the brittle fracture of their contiguous host rocks. Constituents of veins occupying faults or breccias show no evidence of post-ore movements. Alteration of wall rock has not been detected, either in the field, or under the microscope.

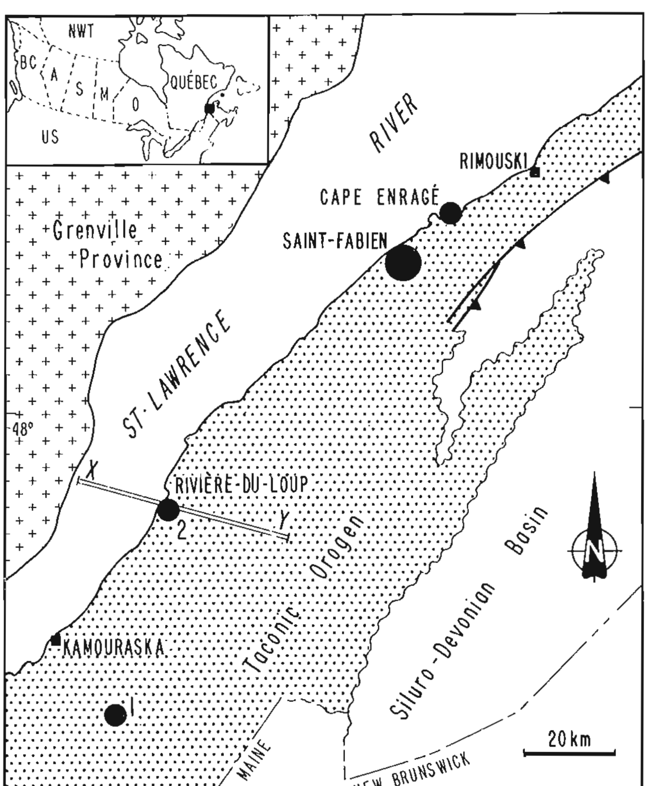

Figure 1. Locations of Ba-Pb(-Zn) occurrences (black dots) in the thrust belt of the Taconic Orogen. The wavy line indicates the discordance between Cambro-Ordovician and Siluro-Devonian sequences; the line in its continuation indicates a thrust fault between the two sequences. The line X-Y locates the cross-section of Figure 3.

In addition to sulphide-bearing veins, abundant disseminated galena (0.2-15 vol. %) and minor disseminated sphalerite and barite are present in one occurrence (the St-Fabien deposit), and traces of disseminated galena are found in Occurrence 6.

OBJECTIVE

In view of the setting of this sediment-hosted vein and vein-associated mineralization remote from igneous intrusives, the main objective of this paper is to contribute to a metallogenic model, emphasizing the driving forces of fluid circulation. Although seemingly a double objective, as it should address the emplacement of both vein and disseminated mineralization, a Pb isotope study of galena in the St-Fabien deposit has shown that the ^{206}Pb, ^{207}Pb and ^{208}Pb to ^{204}Pb ratios in vein and disseminated settings are virtually identical to each other, with coefficients of variation less than 0.06 % (Fig. 4). From these data, it was concluded that the two varieties of galena precipitated simultaneously (Beaudoin, 1987; Beaudoin et al., 1988; Schrijver et al., in press). Thus, a single system of fluid circulation should explain all mineralization.

THE ST-FABIEN DEPOSIT

The following summary draws heavily upon Beaudoin (1987), and therefore only a few points essential to subsequent parts of this paper are given here.

The host rocks

The host rocks are: (1) sandstones, mainly consisting of carbonate-cemented, sand-size quartz grains and 15 to 50 % potassium feldspar and plagioclase; interbedded with (2) conglomerates, mainly consisting of carbonate-rich clasts, up to several centimetres in diameter, in a matrix similar to the contiguous sandstone. Characteristically, detrital plagioclase in both rock types is nearly exclusively albite (Ab = 99.4 mol. %), similar in composition to authigenic feldspar in carbonate clasts (Ab = 99.5 mol. %). Evidence that the composition of detrital plagioclase is due to in situ albitization is provided by rare, up to 50 cm thick, mafic igneous dykes of diabasic texture that consist of albite laths (Ab = 99.6 mol. %) in a chloritic matrix. These dykes are contained in the St-Damase Formation and constitute a very rare type of host for mineralization.

Limestone clasts and calcite cement are pervasively dolomitized in irregular and discordant bodies that commonly coincide with the most highly fractured and mineralized zones. Contacts of dolomitized and nondolomitized domains are knife-sharp to gradational over several millimetres.

The sulphide-bearing veins

These veins show a complete gradation from en echelon tension cracks, to planar regular veins, to highly irregular networks of dilation breccia (Roehl, 1981). Translucent vein constituents are commonly arranged in zones parallel to the contact of the vein and the wall rock, thus suggesting progressive and/or episodic opening during the precipitation of successive zones. The ideal vein consists, from either wall inward, of dolomite (Fe = 0.4-1.3 wt. %), ferroan dolomite (Fe = 1.9-3.1 wt. %), barite and calcite (Fig. 5). Any one of the zones may be missing but no open space remains. The zones are regarded as representing the sequence of precipitation from dolomite to calcite.

Sulphides, mainly galena, preferentially occur in the dolomitic zones of the veins, and may be in contact with the wall rock. Sulphides are normally rare in barite but are locally abundant where the apex of radiating aggregates of this barite is located on selvedges of the vein or on enclaves of the host rock (Schrijver and Beaudoin, 1987, Fig. 2C).

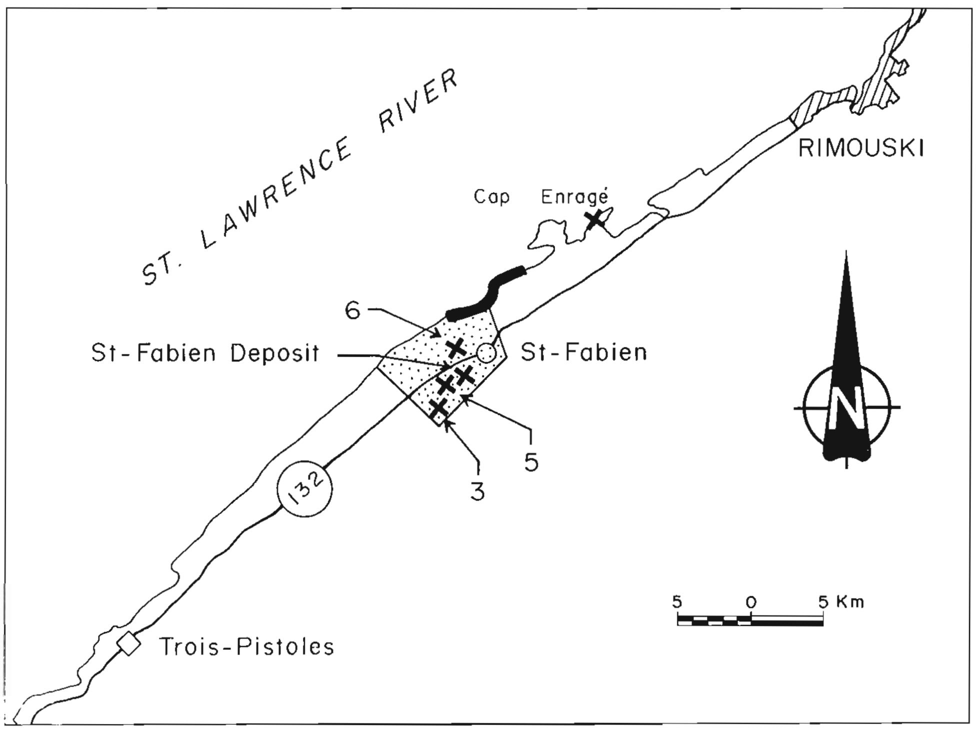

Figure 2. Location map of the study area (patterned); mineral occurrences (crosses) other than the St-Fabien deposit and the Cap Enragé occurrence (discovered August 1988) are numbered according to Schrijver et al. (in press). The heavy black line indicates the location of the pink barite veins.

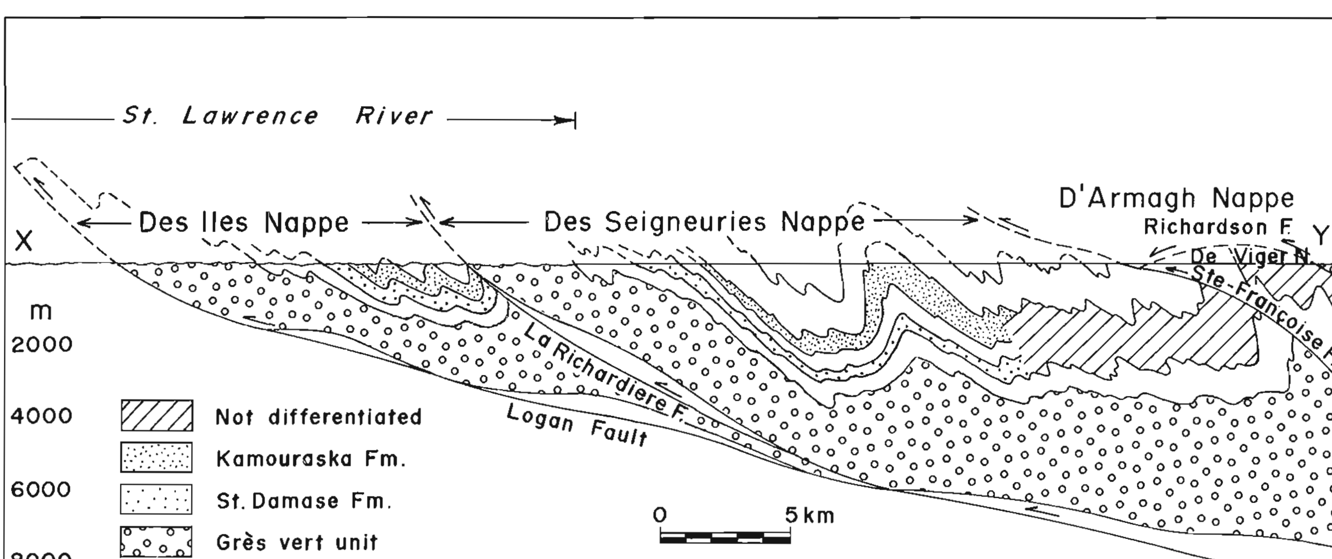

Figure 3. Schematic vertical section near Rivière-du-Loup with names of nappes and major thrust faults after Vallières, 1984). Units consisting dominantly of sandstones are patterned. Fieldwork by Schrijver and Rhéaume (1988) has shown that the gross features of this section are applicable to the study area.

Table 1. Summary of lithostratigraphy and tectonic interpretation of evolution of continental margin; approximate formation thickness at Rivière-du-Loup (after Vallières 1984).

PERIOD	EPOCH	Lithology	Thickness (m)	Interpretation
Ordovician				↑
Ordovician	Middle / Lower	TOURELLE FM. Top of pre-Taconic sequence, grading from quartzitic (base) to poorly bedded, massive, quartzofeldspathic sandstones with interlayered pelites; turbidites.	4500	Emersion and erosion; destruction of continental margin subsequent to suturing. Detrital chromite indicates sedimentation subsequent to emplacement and erosion of ophiolites. Sediment transport from southeast, i.e. from older Paleozoic rocks.
Ordovician	Lower	RIV. OUELLE FM. Multicolored pelites, minor sandstones and rare carbonate rocks; common bioturbation.	3500	? —— Subduction, collision and closure —— ? —— ? —— ? —— ? —— ? —— ? —— ? —— ? —— ? ——
Ordovician / Cambrian	Lower / Upper	KAMOURASKA FM. Sandstones and minor calcareous conglomerates; similar to St-Damase Fm but with little feldspar and no turbidites.	2700	Second phase of convergence. Sediment sources are Grenville basement of subdued relief, and craton-rimming-carbonate platform.
Cambrian	Upper	RIV.-DU-LOUP FM. Shales interlayered with minor turbidites.	2500	Probable interruption of convergence.
Cambrian	Upper	ST-DAMASE FM. Sandstones and calcareous conglomerates in major base-of-slope submarine channel (Hein and Walker 1982); turbidites (host of Ba-Pb-Zn occurrences).	2300	First phase of convergence and development of unstable margin. Sediment sources are Grenville basement and craton-rimming carbonate platform.
Cambrian	Middle	ORIGNAL FM. Multicolored pelites, minor sandstones; marked absence of turbidites.	2000	Oceanic expansion and development of stable margin.
Cambrian	Lower	GRES VERTS UNIT. Arkoses similar to St-Anselme Fm, interlayered with minor pelites (host of Ba-Pb occurrence).	1700	Further separation of basement with formation of sedimentary prism in one or more intracratonic basins at western continental margin of Proto-Atlantic Ocean. Sources of sediment are high-relief parts of Grenville basement within the basin(s).
Pє	Hadrynian	ST-ANSELME FM. Alkaline basalts interlayered with immature arkoses.	400	805(?)-570 Ma: initial rifting and block-faulting of highly metamorphosed crystalline Grenville basement.

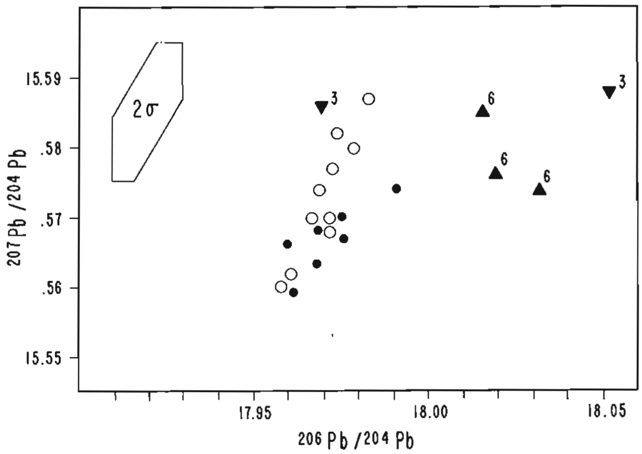

Figure 4. Pb-isotope ratios of vein and disseminated galenas of the St-Fabien deposit (filled and open circles, respectively), and of vein galenas of occurrences 3 and 6 (triangles with occurrence number). The linear distribution of the St-Fabien ratios reflects the ^{204}Pb error.

Galena, sphalerite and barite, where in contact with dolomite crystals, assume the euhedral form of the latter; the sulphides may show tiny apophyses into dolomite rhombs.

Disseminated mineralization

Disseminated sulphides, mostly galena and minor sphalerite and, rarely, disseminated barite and chalcopyrite, occur as cement (Schrijver and Beaudoin, 1987, Fig. 2A), and characteristically form straight-edge contacts with quartz and carbonate crystals. Earlier cements, including phyllosilicates and carbonates, were partially replaced or dissolved and removed, in the latter case creating a secondary porosity. Some framework constituents, including carbonate clast, detrital potassium feldspar and, rarely, plagioclase in both sandstones and the sandy matrix of conglomerates, were also partially replaced by galena, barite and sphalerite (Schrijver and Beaudoin, 1987, Figs. 2B,D).

Gradations exist from isolated sulphide grains, to linear strings of grains, to grains partially embedded in discontinuous miniscule veinlets (e.g. 0.5 mm wide, mm long) of carbonate and/or barite, to sulphides in zoned veins (e.g. 5 mm wide).

Paragenesis

The paragenesis summarized in Figure 6 (dolomitization and ore-stage) is based essentially on the sequence of precipitation deduced from gangue and ore minerals in veins, and is assumed to be similar to that of the disseminated mineralization. In the latter, the paragenesis is less clearly defined, since no progressive episodic opening can be assumed for minerals that precipitated, at least initially and probably only in part, in a secondary dissolution porosity.

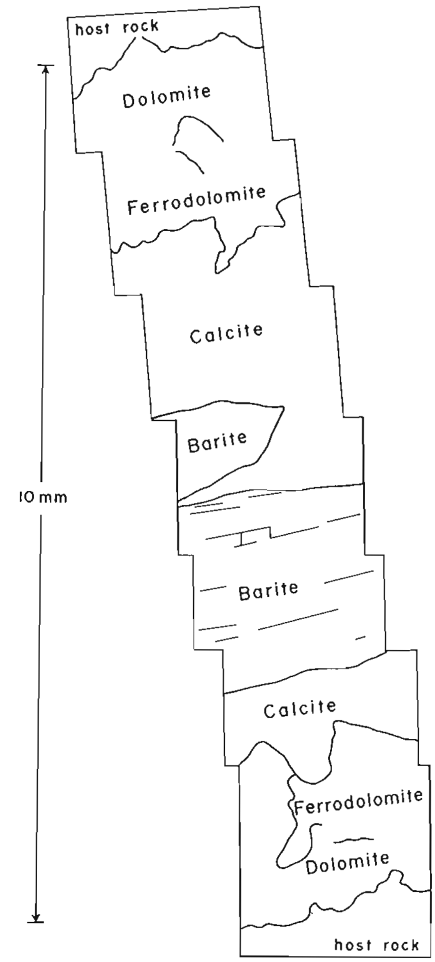

Figure 5. Zoned mineralized barite veinlet from drift of the St-Fabien deposit. Drawing from mozaic of photomicrographs. Further along this veinlet, calcite forms the central cement.

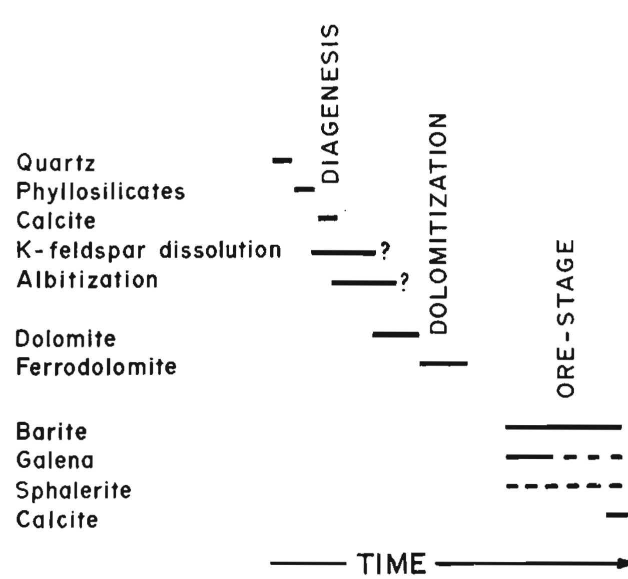

Figure 6. Summary of parageneses of the St-Fabien deposit, including diagenesis, dolomitization and ore precipitation (adapted from Beaudoin, 1987).

Precipitation of sulphate and sulphides

The presence of ferroan dolomite indicates a reducing environment, with a low activity of aqueous sulphur species prior to sulphate-sulphide precipitation. The latter event marks the influx of a metal- and sulphur-bearing fluid or the mixing of a reduced metalliferous and an oxidized sulphate-bearing fluid, both channeled by an interconnected network of incompletely cemented and/or reopened fractures. Textural evidence suggests that barite and sulphides precipitated penecontemporaneously in these remaining or reopened spaces, thus forming mineralized veins. Within the fluid, the process of sulphate reduction and sulphide precipitation in the presence of organic matter, according to the simplified reaction below, would lead to the production of carbonic acid (CH_4 represents organic matter). The acid would cause partial dissolution of carbonate constituents, thus creating a secondary porosity or enlarging pre-existing pores. Feldspar, however, would probably not be affected by this acid, particularly if it were rapidly neutralized by carbonate dissolution. The pores acted as nucleation sites of sulphides (cf. de Boorder, 1986) that may also have replaced portions of the pore walls.

$$So_4{}^{2-} + CH_4 + Pb^{2+} = PbS\downarrow + CO_2 + 2H_2O$$

If organic matter (now present as 0.03 to 0.32 wt. % total organic carbon in barren siltstones) acted as reductants, this would explain the dominance of disseminated sulphides in dissolution porosities and the relative paucity of sulphides in major barite veins. Finally, deposition of calcite in the remaining open spaces, upon depletion of the hydrothermal fluid in either or both metal and sulphur, blocked all transport channels.

OTHER Ba-Pb(-Zn) OCCURRENCES

Outstanding differences of these occurrences (3, 5 and 6 in Fig. 2) with the St-Fabien deposit are: (1) the very minor quantity of disseminated sulphides; (2) the absence of dolomite as a gangue mineral (thus, mineralized veins are not or poorly zoned) and its rarity as a host rock constituent; and (3) the close spatial association between the mineralization and major strike-slip faults. Similarities are (1) the constituent mineral species, galena and barite; (2) the preferred location of galena in the selvedges of baritic veins, thus in contact with wall rock; (3) their structural settings in hinges of major anticlines (Occurrence 5 is in the same anticline as the St-Fabien deposit); and (4) the galena-Pb isotope ratios of occurrences 3 and 6, although grossly similar to those of the St-Fabien deposit, are relatively heterogeneous within and between themselves (Fig. 4).

The times of emplacement of vein mineralization of these occurrences are similar to that of the St-Fabien deposit relative to regional structural elements, but the emplacements had simpler, more obvious and better defined structural controls. As noted above, Pb isotope ratios of galena in occurrences 3 and 6 are grossly similar to those of the St-Fabien deposit, and it seems likely that their galenas crystallized contemporaneously with the St-Fabien mineralization. The relative heterogeneity of the ratios, as compared with the pronounced homogeneity of the St-Fabien galenas, is due, more likely, to slightly different sources of lead than to grossly different times of extraction of lead from these sources. The described similarities with the St-Fabien deposit, however, contrast with the paucity of disseminated galena in occurrences 3, 5 and 6. To explain this, it is proposed that an interconnected network of incompletely cemented and/or reopened fractures did not exist. Thus, the production of carbonic acid would have been limited, and little or no secondary porosity was developed or enlarged.

SOURCES, RECEPTACLES, AND TRANSPORT CHANNELS

Sources

Based on the upper crustal signature of the Pb isotope ratios of all barite-galena occurrences but one (Occurrence in Fig. 1; the newly discovered Cap Enragé occurrence has not been analyzed) in and southwest of the study area, it has been tentatively concluded that the immediate sources of lead in the metalliferous fluids were local siliciclastic rocks (Schrijver et al., in press). The dominant extraction process would have been the diagenetic alteration of feldspar in either or both the St-Damase Formation and the Grès verts unit. The observed features of dissolution of potassium feldspar, a mineral known to contain measurable quantities of lead (3-122 ppm; Smith, 1974, p. 101), support this inference. Potassium feldspar could also have been the main source of barium in view of the fact that significant quantities (2600-4800 ppm) of this element are present even in leached and partially dissolved grains of the St-Fabien deposit (Beaudoin, 1987).

Since the sources of sulphur are unknown, it can only be speculated that oxidizing sulphate-rich fluids originated in evaporitic paleoenvironments known to have occurred, among others, in Early Silurian (Llandoverian) time in the Siluro-Devonian basin of the northern Appalachians (P.-A. Bourque, pers. comm., 1988). Closest to the study area, suspected relicts of evaporites are present as bedding-parallel nodular laminae of supratidal to intertidal calcareous siltstones lying on, and slightly above, the Cambro-Ordovician — Silurian discordance.

Receptacles

Receptacles, i.e. fracture porosity (cf. Currie, 1977) and dissolution porosity, have been described and discussed to some extent in previous sections of this paper. In nonorogenic settings, secondary porosity is most commonly developed under conditions of deep burial diagenesis (Schmidt and McDonald, 1979). Giles and Marshall (1986, p. 243), however, found from mass-balance calculations that "meteoric water penetration and mixing corrosion may be able to create significant volumes of secondary porosity in the shallow subsurface" ("telogenesis" by decarbonatization, Schmidt and McDonald, 1979, p. 120).

The decarbonatization porosity described in this paper is unlike that referred to above since open pores and fractures are virtually nonexistent; instead they are nearly invariably occluded by carbonate and, locally, sulphate and sulphide. In summary, the present problem finds itself between that of the literature cited and that of the geological evolution of ancient orogenic basins (e.g. Davy and Gillet, 1986). In such basins, the evolution of hydraulic conditions and geothermal gradients is governed not only by normal diagenetic burial, but also by superimposed thrusting, folding and fracturing. It is conceivable that most pores that formed during normal early and burial diagenesis were obliterated by these later events, thus greatly reducing permeability. A period of uplift, after nappe emplacement, would seem to be a more proper time for the formation of decarbonatization porosity that was subsequently occluded by cement.

Conversely, feldspar alteration, including dissolution and albitization, are processes well known to occur during normal burial diagenesis. In the Gulf of Mexico basin, "albitization of plagioclase is observed to proceed to completion slightly deeper than the point at which K-feldspar disappears completely" (Land et al., 1987, p. 209). Whether or not this is generally applicable, and potassium feldspar dissolves at all or disappears completely, would depend on: (1) the presence of a medium that can dissolve aluminium silicates and transport Al (e.g. carboxylic acid solutions forming Al-organic complexes; Surdam et al., 1984); and (2) the varying ratio a_{Na^+}/a_{K^+} in, and thus temperature, sources and rate of movement of, diagenetic brines. High ratios would favour both potassium feldspar dissolution and albitization (L.S. Land, pers. comm., 1987). Returning to the sources of lead and barium, it seems likely that leaching and further alteration of feldspars took place relatively early (pre-Taconic) and regionally. The liberated Pb and Ba were removed by circulating brines, locally concentrated in late fractures, mixed with sulphate-bearing waters and precipitated barite and sulphides according to the reduction process discussed previously.

Transport channels

All barite-galena occurrences in, and southwest of, the study area are confined to those sandstone-dominant units that are characterized by albitization of feldspar and, at least in the St-Damase Formation, by the dissolution of potassium feldspar. Diagenetic feldspar alteration in sandstones higher in the stratigraphic column is relatively minor (Lajoie et al., 1974). Thus grès verts and St-Damase sandstones constitute potential sources of Ba and Pb. Moreover, the St-Damase Formation, sandwiched by shale-dominant formations, contains all but one (occurrence 1 in Fig. 1) of the barite-galena occurrences. Thus, it is likely that this formation was permeable up to and including a late brittle stage of its deformation.

At a more detailed scale, fracturing in a dilatant stress field is our preferred mechanism for the formation of transport channels (e.g. Phillips, 1972; Sibson et al., 1975; Roehl, 1981). This general mechanism would explain (1) the fact that the mineralized veins occupy fractures along which no or at most slight movement appears to have taken place, and (2) the occurrence of baritic dilation breccias which most clearly developed in the St-Fabien deposit.

Furthermore, hinges of anticlines which are more nearly concentric than similar, are zones of extension, and thus potentially dilatant. The initial stages of the development of major strike-slip faults may have been controlled by fold shape and/or by the stress field at the time of folding. Rather than being single continuous fault planes, these faults probably consist on a finer scale of numerous discontinuous segments along which zones of compression and extension (i.e. potentially dilatant) alternated (Fig. 7). The coincidence of dilatant zones, originated by folding and faulting, would explain the fact that all baritic vein occurrences in the study area (Fig. 2) are located in anticlinal hinges where these hinges are cut by late faults and fissures.

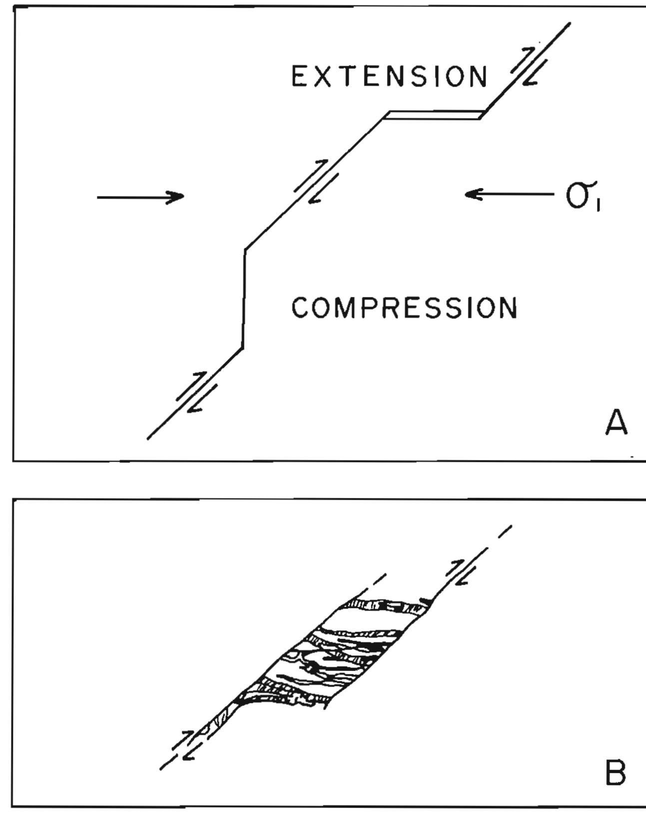

Figure 7. Zones of compression and extension along strike-slip faults. A: Model. B: Example (after Sibson, 1987).

The role of faults as transport channels is indicated, at least for a segment of the La Richardière Thrust, by the increasing density of a greatly dispersed system of calcite veinlets toward the fault (in a St-Damase calcareous conglomerate). Contiguous to the fault plane, this rock loses its identity and is mostly (~70 %) replaced by a diffuse network of calcite patches and stringers. It is speculated that such faults, in post-thrusting time, may have acted as channels of oxidizing sulphate-bearing waters (cf. Fyfe and Kerrich, 1985).

PROPOSED SEQUENCE OF DRIVING FORCES OF FLUID CIRCULATION AND SYNTHESIS

Bethke (1986, p. 540) stated that "migrations [of hot saline groundwaters] have been attributed to sediment compaction, compression during continental collision, and orogenic uplift ...". Although this statement refers to settings other than that of the present study area, we propose that these three major events took place sequentially and, together, played an important role in the genesis of the epigenetic Ba-Pb-Zn occurrences discussed.

Compaction

The driving force of fluid circulation during early diagenesis was undoubtedly simple physical compaction under the increasing load of overlying sediments, with a contribution from chemical compaction. Most likely, quartz and calcite cements were deposited in primary pores. At some considerable depth of burial (2-3 km?; cf. Land et al., 1987) and elevated temperatures (50-100°C?), the fluid became more saline and reducing, dissolved cements, leached feldspar, partially dissolved potassium feldspar and brought about volume-for-volume replacement of plagioclase by albite. At this stage, pelitic beds would be tightly sealed, and coarser grained sediments (sandstones, conglomerates) would become more permeable providing an interconnected network of secondary pores had developed. The grès verts unit and the St-Damase Formation could thus have constituted aquifers relatively early in the diagenetic history of the sedimentary sequence.

Convergence, subduction and collision

Convergence, subduction and collision involving continent-size processes and deep-seated energy sources, and probably having culminated in Early Ordovician time (Table 1), must have affected the extant sedimentary pile and probably initiated its "transformation" from autochthonous to allochthonous. Furthermore, depending on the sector in which subduction and collision took place with respect to the position and configuration of the pre-Taconic sedimentary pile, an addition to the fluid volume may have occurred from upward migrating fluids released from the subducting slab (e.g. Peacock, 1987). In summary, the forces involved greatly changed the circulation and possibly the composition of diagenetic-hydrothermal fluids.

Thrusting, folding and faulting

Thrusting, folding and faulting, probably driven by the same sources of energy that effected subduction and collision define the Taconic Orogeny. This is proposed to be the earliest time that mineralization associated with late faults (e.g. strike-slip) could have occurred, culminating during the subsequent event of the establishment of a dilatant stress field.

Uplift, fracturing and dilation

The Middle Ordovician Tourelle Formation (Table 1) records the uplift and erosion of the Taconic Thrust Belt. At and subsequent to this time, the structural configuration induced a hydraulic gradient and provoked an increase in the rate of migration of fluids in directions depending upon extant permeability, dilatant or potentially dilatant structures and regional topographic relief. Gravity-driven sulphate-bearing fluids — possibly of surficial marine origin — would mix with migrating metalliferous brines — initially of deep burial origin — and precipitate most or all of the sulphate and sulphides in, and southwest of, the study area.

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Revised late Precambrian stratigraphy near Saint John, New Brunswick

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Abstract

Latest Precambrian strata can be divided into a basal felsic quartz-feldspar porphyry, ignimbrite and tuff unit, an overlying basalt unit and an upper redbed unit. The redbeds pass with slight disconformity, or by transition, into the lowermost Cambrian Ratcliffe Brook Formation. The Ratcliffe Brook Formation and the other three units form a mainly terrestrial sequence spanning the Cambrian-Precambrian boundary, which should be distinguished from the marine Cambro-Ordovician Saint John Group.

The underlying Coldbrook Group of subduction-related basalt and intermediate volcanic rocks, mainly fragmental, and minor sedimentary rocks has not been stratigraphically subdivided. The structure of the Coldbrook Group is imperfectly known, but an unconformity separates it from overlying strata.

Turbidite-basalt sequences atypical of either Coldbrook or Eocambrian strata are tentatively assigned an older (780 Ma) age. Relations between these rocks and other late Precambrian strata are unknown.

The proposed threefold division of Eocambrian strata resembles the current subdivision of the late Precambrian strata of the Avalon Peninsula of Newfoundland, and suggests possible detailed tectonostratigraphic correlations.

Résumé

On peut subdiviser les strates du sommet du Précambrien en un porphyre quartzo-feldspathique felsique basal, en une unité d'ignimbrite et tuf, en une unité basaltique sus-jacente, et en une unité supérieure de lits rouges. Les lits rouges passent avec une légère discordance stratigraphique, ou de façon transitionnelle, à la formation de Ratcliff Brook de la base du Cambrien. La formation de Ratcliffe Brook et les trois autres unités forment une séquence principalement terrestre, qui couvre la limite entre le Cambrien et le Précambrien, et que l'on doit distinguer du groupe cambro-ordovicien de Saint John, de caractère marin.

Le groupe sous-jacent de Coldbrook qui se compose de basaltes formés en rapport avec un épisode de subduction, de roches volcaniques intermédiaires surtout clastiques et de quelques roches sédimentaires, n'a pas fait l'objet d'une subdivision stratigraphique. On connaît assez mal la structure du groupe de Coldbrook, mais une discordance stratigraphique le sépare des strates susjacentes.

On a provisoirement attribué un âge plus élevé (780 Ma) à une séquence de turbidites et basaltes atypiques de Coldbrook ou des strates éocambriennes. On ne connaît pas les relations existant entre ces roches et les autres strates précambriennes supérieures.

La division proposée en trois parties rappelle la subdivision actuelle des strates du Précambrien supérieur de la péninsule d'Avalon à Terre-Neuve, et suggère la possibilité d'établir des corrélations tectonostratigraphiques détaillées.

INTRODUCTION

Avalonian terranes, defined by little metamorphosed late Precambrian volcano-sedimentary strata and Cambrian strata with an Acado-Baltic fauna, fringe much of Atlantic Canada. Recent stratigraphic (Jenkins, 1984) and radiometric (Jamieson et al., 1986) studies show that late Precambrian rocks of Avalonian terranes, both supracrustal and plutonic, fall into three well-defined divisions which in Canada give U/Pb emplacement ages of about 760 ± 20, 615 ± 20 and 560 ± 20 Ma. This subdivision requires that a revision of the stratigraphy for Avalonian terranes be made. Around Saint John, New Brunswick, sedimentary Eocambrian strata occur below the Tommotian Ratcliffe Brook Formation of the Saint John Group, but above the volcanic rocks of the Coldbrook Group (Currie, 1984), and criteria for distinguishing these strata from superficially similar sedimentary rocks of differing ages have been published (Tanoli, 1987). However, little consideration has been given to distinguishing volcanic rocks of differing ages, or to identifying strata of 760 Ma age, although plutons of about this age are known to be present (Olszewski et al., 1980). The present communication addresses these questions. The general distribution of the units and the deduced stratigraphic column appear in Figure 1.

STRATIGRAPHY OF THE EOCAMBRIAN STRATA

Hofmann and Patel (1988) have shown that the Ratcliffe Brook Formation, presently assigned to the Saint John Group, is of Tommotian age, that is, basal Cambrian. Redbeds underlie the Ratcliffe Brook Formation in a number of places around the Saint John region, as detailed by Tanoli (1987). Volcanic rocks associated with high-level granite plutons giving ages of 565 ± 8 Ma occur west of Saint John (Currie, 1987) and are considered to be correlative to these latest Precambrian (Eocambrian) strata. Reexamination of all known localities of Eocambrian strata reveals a simple threefold stratigraphic division. At the top there occurs the sequence of redbeds described by Tanoli (1987). These redbeds, which vary in thickness from 30 to 120 m include distinctive blood-red tuff, various sand and siltstones charged with feldspar laths and conglomerate with clasts of quartz-feldspar porphyry, felsite and rhyolite. Thinly banded greenish siltstone occurs in the upper part of the section. The resemblance to the Ratcliffe Brook Formation is very close, and a good case could be made for including these beds with the Ratcliffe Brook Formation. Where strata of both units are exposed together their attitudes are conformable, or less than 10° nonconformable. The structural history of both units appears to be the same, and they are consistently spatially associated.

The redbed sequence locally contains vesicular basalt within it (for example at Devils Back north of the Long Reach), and is underlain by commonly vesicular basalt. These basalts appear remarkably fresh and undeformed in many outcrops. Vesicular flows are locally accompanied by dark tuffaceous strata, presumably basaltic tuffs, for example in roadcuts on Highway 7 south of Welsford. Slaggy, reddened flow tops suggestive of subaerial extrusion can be recognized on some flows. The thickness of the basaltic interval reaches as much as 150 m, but generally does not exceed 50 m, and may be absent.

The lower part of the Eocambrian section consists of distinctive red to pale green quartz-feldspar porphyry, rhyolitic flows and welded ignimbrite and fragmental rocks. Much of this unit is a monotonous quartz-feldspar porphyry with 5-15 percent of 1-3 mm quartz and feldspar phenocrysts in a homogeneous, very fine grained matrix. Quartz phenocrysts consistently outnumber feldspar phenocrysts, and the latter may be absent. These massive rocks locally contain inclusions up to 30 cm across of rhyolite strongly banded on a millimetre scale, as well as blobby or lenticular aphanitic red streaks reminiscent of fiamme. The porphyritic rocks appear to grade to fragmental and tuffaceous material, although the amount of outcrop is not sufficient to examine these transitions in detail. The felsic magmatic rocks appear to consist of a complex mixture of high-level intrusives, or perhaps domes and tholoïds, with welded ignimbrites and minor fragmental material. The thickness of this felsic unit is unknown because no complete single section has been found, and composite sections are very difficult to measure accurately because of the massive nature of much of the material. A single roadcut on Highway 7 south of Welsford exposes an apparent thickness >1000 m.

STRATIGRAPHY OF THE COLDBROOK GROUP

No stratigraphic contact between Eocambrian strata and the underlying (?) Coldbrook Group has been discovered. Wherever the two units outcrop together, the contact is clearly faulted. However three lines of indirect evidence suggest unconformity between the two units:

(a) The Saint John Group rests with unconformity on the Coldbrook Group on Hanford Brook, east of the mapped area (Hoffmann and Patel, 1988). Since the Eocambrian strata underlie the Saint John Group conformably, or with slight disconformity, it follows that they probably overlie the Coldbrook with unconformity.

(b) Eocambrian strata exhibit a consistent association with the Saint John Group, but the Coldbrook Group does not. It follows that distribution of the Coldbrook Group may be controlled by different structures than those controlling outcrop of the Eocambrian strata.

(c) Where Coldbrook Group and Eocambrian strata outcrop in close proximity, for example at the Westfield interchange of Highway 7, or at the southern end of the airport expressway, the Coldbrook Group consistently shows a pervasive cleavage and epidote-chlorite metamorphism whereas Eocambrian strata show only patchy local epidotization and cleavage.

In broad terms the Coldbrook Group consists of three easily distinguished "packages" of rocks, namely massive to hyaloclastic basalts, locally amygdaloidal, fragmental volcanic rocks of intermediate composition and sedimentary rocks, commonly grey- green rhythmically banded siltstones, but locally including conglomerate. These units occur in various combinations throughout the exposed region of Coldbrook Group rocks. Attempts to construct a local stratigraphy based on distinctive fragmental units and sedimentary strata failed due to lack of outcrop and lensoid units. Few reliable facing directions have been found in the Coldbrook Group because of tectonic overprinting. Structure is therefore poorly known. Based on lithologic mapping, sparse facing directions and cleavage relations, the belt northeast of Saint John appears to form a gently northeast-plunging syncline.

The age of the Coldbrook Group likewise remains uncertain. Dating by Cormier (1969) gave a poorly constrained Rb/Sr age of about 750 Ma, but Cormier regarded this age as unreliable due to internal inconsistencies. Stukas (1978) obtained Ar/Ar data on plagioclase from the Kingston Peninsula. Several samples gave plateau ages of about 640 Ma, but most either failed to yield plateaus or gave badly scattered data ranging from 300 to 800 Ma. Sample selection may also be a problem with this data. The Coldbrook Group is here assumed to be roughly correlative to plutons of the Golden Grove suite which gave ages of 625 Ma east of Saint John (U-Pb on zircon, Watters, 1987) and 615 Ma to the west (Rb-Sr isochron, Olszewski et al., 1980).

Basalt of the Coldbrook Group forms lenticular masses up to a hundred metres thick and several kilometres in length. Most weathered surfaces look essentially massive, with a network of inconspicuous fractures causing the surface to have a chipped appearance. Veins and football-size masses of epidote are ubiquitous. On blasted surfaces or favourably glaciated surfaces it can be seen that plagioclase phenocrysts up to 3 mm in length are ubiquitous, and that many outcrops carry disc-like chloritic fragments up to several centimetres in size, or obviously clastic rounded quartz grains. Locally, sinuous rounded masses of quartz, calcite and epidote presumably form the remains of vesicles. In one location, purplish-black plagioclase-phyric basalt fills cracks in a greenish intermediate rock, but basalts appear in general to be older than the intermediate rocks. Preliminary chemical analyses show the basalt to be typical of arcs (Currie, 1988). Basalt forms about 15 percent of exposure.

Dark intermediate rocks are interbanded on a 10-100 m scale with pale green, buff or pinkish volcanic rocks. The strong contrast in colour has led to suggestions that much of the Coldbrook Group is bimodal, but this is not supported by chemical analyses which show the bulk of the material to be intermediate in character (McCutcheon in Ruitenberg et al., 1979). In outcrop, the intermediate rocks are commonly fragmental, varying from sand or granule size fragments to blocks 30 cm across set in a finer grained matrix. Like the basalt, the fragmental nature of the rocks is seldom easily visible on weathered outcrops which have the usual "chipped" appearance. The intermediate rocks which are not fragmental appear to be mainly sills or dykes, typically with plagioclase phenocrysts. Many felsic specimens contain virtually no mafic minerals and a high content of very fine grained quartz. However the high content of epidote suggests that the compositions were originally dacitic or rhyodacitic, rather than rhyolitic. Intermediate rocks form about 75 percent of the exposure of the Coldbrook Group.

Grey-green siltstone, rhythmically banded on a centimetre scale, forms a minor but highly typical part of the Coldbrook Group. The beds commonly show grading from a basal pale grey sandstone with a sharp bottom through darker upper siltstone. Typically the beds are very regular in thickness over the length of an outcrop. Sedimentary intervals up to 100 m thick occur, but 10-20 m is more usual. Northeast of Saint John the siltstone passes to slightly coarser sandstone containing an interval of conglomerate 20 m thick. Fragments in the conglomerate are predominantly fragmental volcanics of familiar type, but there is also some vein quartz present. Sedimentary rocks, excluding tuffs, form about 10 percent of Coldbrook exposure in the Saint John region.

The lower boundary of the Coldbrook Group may be exposed in a large roadcut on a new airport expressway near Upper Golden Grove. This roadcut begins in basalt which contains large (to $10 \times 40 \times 3$ m^3) inclusions of marble of the Proterozoic Green Head Group, then passes gradually into more amphibolitic material, and eventually into an enclave-bearing granodiorite typical of the Golden Grove suite. This roadcut is interpreted to support the view that the Coldbrook Group unconformably overlies the Green Head Group, but grades to, or is equivalent to the Golden Grove suite. Other exposures of the Coldbrook-Green Head contact are faulted, but greenish felsic dykes typical of Coldbrook lithologies cut the Green Head Group and the Brookville gneiss basement in several places, suggesting a probable unconformity.

OTHER POSSIBLE LATE PRECAMBRIAN STRATA

Strata younger than the platformal marble-quartzite-pelite shelf sequence of the Green Head Group and older than the Cambrian Saint John Group could fall into one of the threefold divisions of the late Precambrian noted above. In addition to the two units noted above, two other rock units fit this definition, namely the Martinon Formation, and sedimentary enclaves separating late Precambrian granitoid plutons north and west of Grand Bay. Ruitenberg et al. (1979) considered the Martinon Formation to form part of the platformal Green Head Group. This assignment seems unlikely since a spectacular roadcut on Highway 7 exhibits a debris flow in the Martinon Formation with numerous large clasts of Green Head limestone, many exhibiting pressure solution phenomena. Much of the Martinon Formation consists of hornfelsed turbiditic siltstone in which distal Bouma cycles can be distinguished on favourable exposures. The formation contains a significant component of basalt flows and sills which are fine grained and dense, thus differing from either the Coldbrook or Eocambrian basalts. These basalts closely resemble in petrography dykes found in the Green Head Group. The Martinon Formation formed by slumps on an unstable slope subsequent to lithification of the Green Head Group, but prior to its metamorphism about 780 Ma. Basaltic volcanism was active during its formation.

Northwest of Grand Bay, Devonian and late Precambrian rocks intrude curvilinear belts of dark grey siltstone with basalt sills. Much of this material is hornfelsed to a

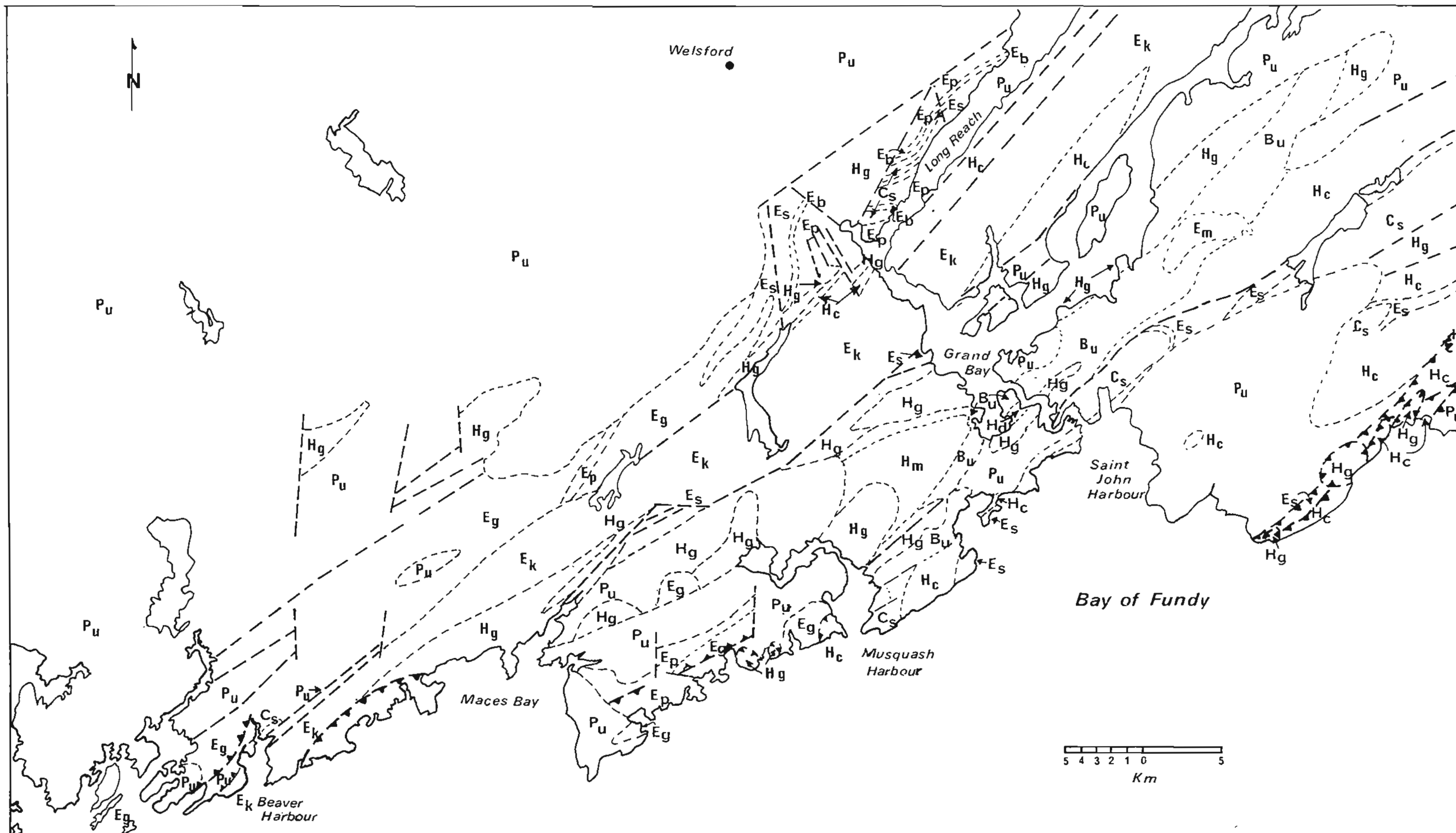

Figure 1. Geological map of two Saint John region.

Map legend showing revised late Precambrian stratigraphy, Saint John region.

SILURIAN AND YOUNGER

Pu — undivided post-Cambrian units (Silurian to Quaternary)

Supracrustal units

CAMBRIAN AND LOWER ORDOVICIAN

Cs — SAINT JOHN GP. (excluding Ratcliffe Brook Fm.) littoral to shallow marine grey-green sandstone and siltstone passing upward into black shale. RATCLIFFE BROOK FM. red and grey-green siltstone, conglomerate and thin tuffaceous beds

— conformable to disconformable contact —

EOCAMBRIAN

Es — red tuff and siltstone, pink sandstone, pink conglomerate with rhyolite and porphyry cobbles; minor grey-green siltstone; includes minor Eb, Ep

— conformable contact —

Eb — vesicular basalt, locally with reddened slaggy flow tops, minor tuffaceous siltstone

— conformable contact —

Ep — red to pale green quartz-feldspar porphry, rhyolite, flow-banded rhyolite and ignimbrite

— unconformity —

HADRYNIAN

Hc — COLDBROOK GP. basalt, fragmental intermediate volcanics; green siltstone, conglomerate

— unconformity(?) —

Hm — MARTINON FM. turbiditic siltstones, marble debris, basalt flows and sills

— unconformity —

MIDDLE PROTEROZOIC AND OLDER

Bu — undifferentiated Proterozoic and older basement

Plutonic units

Em — gabbro, ultramafics

— intrusive contact —

Eg — granitoid plutons (565 Ma

— gradational contact —

Ek — bimodal sheeted dykes (KINGSTON COMPLEX)

— intrusive contact —

Hg — GOLDEN GROVE STE. (620 Ma) diorite to granite

Hd — diorite gneiss, basalt dykes (780 Ma)

geological contact . . fault, high angle or motion unknown . fault, thrust . .

considerable degree, but on favourable exposures partial Bouma cycles appear to be preserved. These rocks form separators between late Precambrian plutons. Their age is therefore greater than 565 Ma, age of the youngest Precambrian plutons, but otherwise uncertain. Lithologically the rocks resemble the Martinon Formation.

STRUCTURE OF THE LATE PRECAMBRIAN ROCKS

Folding on a metre scale has been observed in the Coldbrook Group in numerous places, and on a kilometre scale in the Martinon Formation and Eocambrian section. The extent, significance and timing of this folding is doubtful. West of the Long Reach, Eocambrian siltstones exhibit northwest-trending drag folds (and cleavage) in opposite senses on opposite sides of a horst(?) of older granite. Five kilometres to the east the Eocambrian rocks can be traced around the nose of an east-west-trending fold. Prominent cleavage wraps around this fold, which does not itself display an axial plane cleavage. Farther northeast along the Long Reach, cleavage in the Saint John Group bends around an angle of 45° in five kilometres. South of the Long Reach, the trend of cleavage in the Coldbrook Group is equally irregular. In belts of Coldbrook and Eocambrian rocks east of Saint John, cleavage and minor fold directions associated with pre-Carboniferous deformation likewise tend to be erratic. Along the Bay of Fundy, an older structure is strongly overprinted by intense northeast-trending cleavages of Carboniferous age (Currie and Nance, 1983; Watters, 1987). At present it is unknown whether the erratic directions of cleavage and minor folding represent: (a) local, and presumably late, deformation associated with the movement of relatively small blocks; or (b) more pervasive, possibly older, deformation which has been variably broken up and rotated by younger movements. Probably both possibilities are represented locally. Both alternatives suggest that the distribution of outcrop is mainly due to faulting.

Faults of at least four distinct types occur in the Saint John region: (a) northeast-trending steep ductile faults (mylonite zones) with late Precambrian motion; (b) northeast-trending steep brittle faults with Silurian to Carboniferous motion; (c) north- to northwest-trending steep brittle faults whose time of motion is uncertain; and (d) northeast-trending shallow to steeply dipping brittle-ductile Carboniferous faults.

The old mylonite zones rarely affect Coldbrook and Eocambrian volcanic strata east of Grand Bay. However, Eocambrian sedimentary strata overlie the mylonite southwest of Grand Bay. This relationship shows that fault movements were of Precambrian age.

Northeast-trending brittle faults control much of the outcrop pattern of Coldbrook and Eocambrian strata. There are two subsets of such faults, a slightly older set trending about 070° and a younger set trending about 040°. Both sets exhibit mainly normal motions. Movement on the older faults was overby Late Silurian to Early Devonian time as defined by relations to dated plutons (Currie, 1988). Movement on the younger faults persisted into the Early Carboniferous, as demonstrated by relations to Carboniferous strata. North of the Long Reach both sets of faults exhibit systematic north-side down displacement. South of the Long Reach the net effect of both sets of faults has been to produce grabens from a few hundred to a few thousand metres across, within which the rocks have been down-dropped a few hundred metres at most, preserving outliers of Eocambrian and younger strata. The position of some of the larger brittle faults appears to be controlled by older ductile faults. The younger faults tend to follow roughly the same trends, but cut across the older fabrics at a low angle.

North- and northwest-trending faults have only recently been recognized in the Saint John region. South of Welsford they affect distribution of Eocambrian rocks in a spectacular fashion which can be observed in roadcuts. The net effect is to produce alternating northwest-trending horsts and grabens a few hundred metres wide, which expose older granites and Eocambrian rocks respectively. These faults also appear to have produced significant lateral motions, as evidenced by drag folds. Commonly the drag folds alternate in sense across paired faults, suggesting that the net lateral displacement may be small. At present it is uncertain whether the faults trending about 000° are a separate set from those trending about 335°. The age of motion of these faults likewise remains uncertain, but the 335° set appears to cut Lower Carboniferous strata in Grand Bay.

Carboniferous thrust faulting along the Bay of Fundy has been described in some detail by Currie and Nance (1983) and Watters (1987). The net result has been a westward imbrication of the late Precambrian stratigraphy. Within the narrow deformed zone along the Bay of Fundy both Eocambrian and Coldbrook strata certainly occur, but many of the criteria for distinguishing units break down in these severely deformed rocks, and the identification of some individual outcrops remains dubious.

DISCUSSION

Stratigraphy and radiometric age dating show conclusively that volcanic and sedimentary strata of two differing late Precambrian units occur in the Saint John region, namely Eocambrian strata and the Coldbrook Group.

Eocambrian strata can be divided into three discrete units, the youngest of which appears to grade to the lowermost Cambrian Ratcliffe Brook Formation. Recent studies by Tanoli (1987) on the stratigraphy of the Saint John Group show that formations above the Ratcliffe Brook belong to a single environment (barrier bar complex) which did not produce the Ratcliffe Brook Formation. It would therefore appear logical to include the Ratcliffe Brook Formation and the three underlying units in a different group spanning the time period from latest Precambrian to earliest Cambrian. These rocks accumulated under subaerial conditions in an environment dominated by gradually decreasing volcanic activity. Radiometric ages on high-level plutons gradational to porphyries (Currie, 1987) suggest that the lowest part of this package has an age of 565 $\pm$ 8 Ma. The base of the Cambrian must be significantly younger than this. The age and stratigraphy closely resemble those of the Musgravetown Group of the western Avalon Peninsula in Newfoundland (O'Brien and Knight, 1988). The ages are also strikingly similar to igneous ages obtained in southwestern Newfoundland (Dunning et al., 1988) and Cape Breton Island (Jamieson et al., 1986). The significance of this latest Precambrian event is not yet completely clear, but it may represent a breakup event leading to the formation of the lower Paleozoic Iapetus Ocean.

The stratigraphy, structure and age of the Coldbrook Group remain incompletely resolved. Strong circumstantial evidence, supported by some new outcrop, suggests that the volcanics are correlative to Golden Grove plutons dated at 615-625 Ma. This would make the Coldbrook Group correlative to the lithologically similar Harbour Main volcanics of Newfoundland (Krogh et al., 1988). Coldbrook volcanic rocks clearly form a subduction- related suite (Currie, 1988). Around Saint John the Coldbrook Group is dominated by volcanic rocks, but to the northeast sedimentary rocks are much more abundant (Giles and Ruitenberg, 1977). Studies of these rocks may permit construction of a satisfactory stratigraphy for the Coldbrook Group, and hence a determination of the structure. The character of the sedimentary and volcanic rocks in the Saint John area suggest a marine origin.

Radiometric determinations (Olszewski et al., 1980) show the presence of an igneous and metamorphic event at about 780 Ma. It remains uncertain whether any supracrustal rocks formed during this event. The Martinon Formation falls into the proper time interval, and the lithologies (turbidites sampling a carbonate shelf, basaltic igneous activity) strikingly resemble those of the Burin Group of Newfoundland, dated by Krogh et al. (1988) at 762 Ma. Assignment of the Martinon Formation to this age bracket is speculative, but more reasonable than the present assignment to the Green Head Group. The hornfelsed siltstone-basalt sequences northwest of Loch Alva are clearly of Precambrian age, but their assignment to the Martinon Formation is speculative. Such an assignment appears the most reasonable alternative.

Revision of the late Precambrian stratigraphy around Saint John emphasizes the resemblance with the Avalon Peninsula of Newfoundland. Some recent authours have suggested that Avalonian terranes are only loosely related, and that significant motions occurred between present exposures (Williams and Hatcher 1983, Keppie, 1984;). However recent studies show striking similarities in age and lithology between the various terranes, suggesting that they form parts of a single whole, only minimally disrupted.

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Preliminary report on a classification of Newfoundland granitic rocks and their relations to tectonostratigraphic zones and lower crustal blocks[1]

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Abstract

Newfoundland granitic rocks are separated into nine categories, each named after a type example as follows: Long Range, Holyrood, Round Pond-Cross Hills, Twillingate, Burgeo, Middle Ridge, Mount Peyton, Topsails and Ackley. Distinctive categories of granitic rocks occur within discrete tectonostratigraphic zones or coincide with lower crustal blocks.

Precambrian plutons (Long Range, Holyrood) that predate the Appalachian orogenic cycle occur on opposing sides of the orogen in the Humber and Avalon zones, respectively. Late Precambrian to Early Cambrian plutons (Round Pond-Cross Hills) in the Humber and Avalon zones are interpreted as synrift with respect to the Appalachian cycle. Deformed early Paleozoic plutons of oceanic or island arc affinity (Twillingate) relate to constructive phases of the Dunnage Zone. Deformed middle Paleozoic plutons that follow metamorphic belts of the Gander Zone (Burgeo, Middle Ridge) probably relate to accretionary events. Other middle Paleozoic plutons (Topsails, Mount Peyton, Middle Ridge) that cross local zone boundaries occur on opposite sides of the Dunnage Zone and may reflect differences in lower crustal blocks. A few others (Ackley examples) cut boundaries of both zones and lower crustal blocks.

Résumé

Les roches granitiques de Terre-Neuve sont subdivisées en neuf catégories. Chacune est nommée d'après un exemple type: Long Range, Holyrood, Round Pond-Cross Hills, Twillinggate, Burgeo, Middle Ridge, Mount Peyton, Topsails et Ackley. Les catégories distinctives de roches granitiques apparaissent à l'intérieur de zones tectonostatigraphiques discrètes, ou coïncident avec des blocs de la croûte inférieure.

Les plutons précambriens (Long Range, Holyrood) mis en place avant le cycle d'orogénèse des Appalaches occupent les versants opposés de l'orogène dans les zones de Humber et d'Avalon respectivement. Les plutons d'âge précambrien supérieur à cambrien inférieur (Round Pond-Cross Hills) dans les zones d'Humber et d'Avalon ont été interprétés comme constituant un fossé d'effondrement syntectonique dans le cadre du cycle de l'orogénèse appalachienne. Les plutons déformés, d'âge paléozoïque inférieur, d'affinités océaniques ou typiques des zones d'arc insulaire (Twillingate) se rapportent aux phases constructives de la zone de Dunnage. La mise en place des plutons déformés d'âge paléozoïque moyen qui suivent les zones métamorphiques de la zone de Gander (Burgeo, Middle Ridge) concorde probablement avec des épisodes d'accrétion. D'autres plutons d'âge paléozoïque moyen (Topsails, Mount Peyton, Middle Ridge) qui traversent les limites locales de zones occupent des versants opposés de la zone de Dunnage, et reflètent peut-être des différences dans la nature des blocs de la croûte inférieure. Quelques autres (exemples d'Ackley) recoupent les limites des deux zones et les blocs crustaux inférieurs.

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INTRODUCTION

Granitic rocks occur in all tectonostratigraphic zones of the Newfoundland Appalachians and make up about 30 percent of the bedrock. They range in age from middle Proterozoic to upper Paleozoic. Nine categories are recognized, modified after the classification of Hayes et al. (1987) and based on relative age, structural style, form, internal makeup, associations with specific country rocks, mineralogy, texture and chemical affinity. Each category is named after a type example (Long Range, Holyrood, Round Pond-Cross Hills, Twillingate, Burgeo, Middle Ridge, Mount Peyton, Topsails and Ackley). Their distribution is shown in Figure 1, with designations of the best known examples and a sampling of available isotopic ages and references indicated in the caption. Geochemical data are available for all nine categories and some individual plutons have been studied in detail. A more complete bibliography is available from the authors on request.

Tectonostratigraphic zones in Newfoundland, defined on contrasts among early Paleozoic and older rocks, are outlined in Figure 1 (inset), with zone boundaries shown on the plutonics map (Fig. 1). The Humber Zone is the Appalachian miogeocline or early Paleozoic continental margin of eastern North America. Zones farther east are suspect terranes accreted to the ancient NorthAmerican margin (Williams and Hatcher, 1983). The Dunnage Zone is divided into the Notre Dame and Exploits subzones and the Gander Zone into the Gander Lake, Mount Cormack and Meelpaeg subzones (Williams et al., 1988).

Spatial relations between granitic rocks and zones have been described previously (Strong and Dickson, 1978; Whalen et al., 1987; Williams et al., 1988), but only for particular classes of plutons and for parts of Newfoundland. We compare the distribution of all major granitic plutons with updated tectonostratigraphic zones and lower crustal blocks for the entire island. No attempt is made to explore petrogenetic implications, although the distribution of most plutons can be rationalized in plate models for the orogen. Renewed interest in the distribution of granitic rocks accompanies the recent clarification of zone boundaries (Williams et al., 1988) and offshore seismic reflection studies that define lower crustal blocks (Keen et al., 1986; Marillier et al., in press). This preliminary paper is presented in anticipation of further work that will test our classification and observations. It is also presented in anticipation of an onland deep seismic reflection profile sponsored by the Canadian Lithoprobe Project that will test the nature and geometry of lower crustal blocks.

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Figure 1. Distribution of Newfoundland granites with respect to tectonostratigraphic zones and lower crustal blocks. Numbered plutons, with type examples in bolder print, are described in the caption.

LONG RANGE: Middle Proterozoic plutons of Grenville inliers; 1. mainly granites of Long Range inlier, about 1040 Ma U/Pb, Erdmer, 1986. 2. Steel Mountain anorthosite. 3. Indian Head Complex.

HOLYROOD: Late Proterozoic granite-granodiorite-gabbro suites; 4. Holyrood, 620 Ma U/Pb, Krogh et al., 1988. 5. Swift Current, 580 ± 20 Ma U/Pb, Dallmeyer et al., 1981.

ROUND POND-CROSS HILLS: Late Proterozoic alkali granites; 6. Round Pond (Humber), 602 ± 10 Ma U/Pb, Williams et al., 1985. 7. Hare Hill, 617 ± 8 Ma U/Pb, van Berkel and Currie, 1988. 8. Cross Hills (Avalon). 9. Louil Hills.

TWILLINGATE: Early Paleozoic mainly deformed plagiogranites; 10. Twillingate, 510 +17/−16 Ma U/Pb, Williams et al., 1976. 11. Hungry Mountain, 467 ± 8 Ma U/Pb, Whalen et al., 1987. 12. Burlington, 461 ± 25 Ma U/Pb, Hibbard, 1983. 13. Southwest Brook, 456 ± 3 Ma U/Pb, Dunning et al., in press.

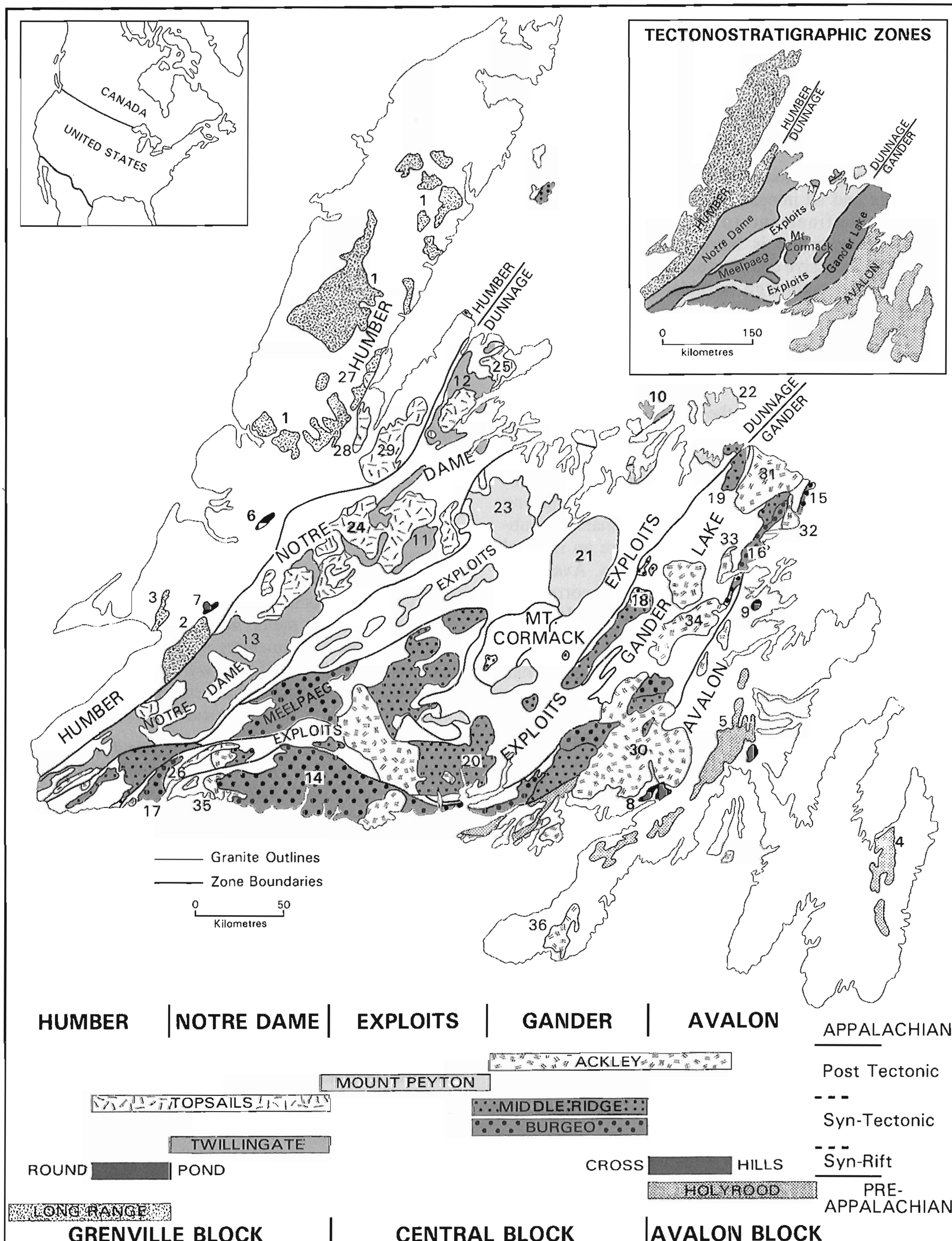
BURGEO: Middle Paleozoic foliated to mylonitic, medium to coarse grained feldspar-phyric biotite granites; 14. Burgeo, 428 ± 4 Ma U/Pb, O'Brien et al., 1986a. 15. Cape Freels, 414 ± 5 Ma Rb/Sr, Bell and Blenkinsop, 1975. 16. Lockers Bay, 460 ± 20 Ma U/Pb, Dallmeyer et al., 1981. 17. LaPoile, 416 ± 4 Ma U/Pb, Chorlton and Dallmeyer, 1986.

MIDDLE RIDGE: Middle Paleozoic foliated to massive garnet-muscovite-biotite granite; 18. Middle Ridge, 383 ± 15 Ma Rb/Sr, Bell and Blenkinsop, 1977. 19. Ragged Harbour, 380 ± 20 Ma Rb/Sr, Bell *in* Strong and Dickson, 1978. 20. North Bay, 427 ± 12 Ma Rb/Sr, Elias and Strong, 1982.

MOUNT PEYTON: Middle Paleozoic massive composite plutons with early mafic and later granitic phases; 21. Mount Peyton, gabbro — 420 ± 8 Ma Ar/Ar, Reynolds et al., 1981, granite — 390 ± 15 Ma Rb/Sr, Bell et al., 1977. 22. Fogo Island, 380 ± 16 Ma K/Ar, Wanless et al., 1964. 23. Hodges Hill.

TOPSAILS: Middle Paleozoic alkali granites associated with Silurian volcanic suites; 24. Topsails, 429 ± 3 Ma U/Pb, 417 ± 4 Ma Rb/Sr, Whalen et al., 1987. 25. Cape Brule, 404 ± 25 Ma Rb/Sr, Bell and Blenkinsop, 1977. 26. Hawks Nest, 410 ± 20 Ma U/Pb, Dallmeyer, 1980. 27. Devils Room, 398 +27/−7 Ma U/Pb, Erdmer, 1986. 28. Gull Lake, 398 Ma U/Pb, Tuach, 1987a. 29. Wild Cove Pond, 392 ± 16 Ma K/Ar, Wanless et al., 1972 (may belong to Mount Peyton or Ackley categories).

ACKLEY: Middle and late Paleozoic massive coarse grained feldspar-phyric to medium grained equigranular biotite granite; 30. Ackley, 355 ± 5 Ma Rb/Sr, Bell et al., 1977; 355 ± 10 Ma Ar/Ar, Dallmeyer et al., 1983, Tuach, 1987b, 367 Ma Ar/Ar, Tuach and Kontak, 1986. 31. Deadmans Bay, 400 ± 13 Ma K/Ar, Stevens et al., 1982. 32. Newport, 344 ± 44 Ma Rb/Sr, Bell et al., 1979. 33. Middle Brook, 374 ± 7 Ma Ar/Ar, Dallmeyer et al., 1981. 34. Maccles Lake, 367 ± 10 Ma Rb/Sr, Bell and Blenkinsop, 1977. 35. Chetwynd, 372 ± 5 Ma Ar/Ar, Chorlton and Dallmeyer, 1986. 36. St. Lawrence, 326 ± 5 Ma Rb/Sr, Bell et al., 1977.


CANADA
UNITED STATES
TECTONOSTRATIGRAPHIC ZONES
HUMBER
DUNNAGE
GANDER
Notre Dame
Exploits
Mt. Cormack
Gander Lake
Meelpaeg
AVALON
0 150
kilometres
HUMBER
NOTRE DAME
EXPLOITS
MT. CORMACK
MEELPAEG
GANDER
LAKE
AVALON
Granite Outlines
Zone Boundaries
0 50
Kilometres
HUMBER
NOTRE DAME
EXPLOITS
GANDER
AVALON
APPALACHIAN
ACKLEY
MOUNT PEYTON
TOPSAILS
MIDDLE RIDGE
BURGEO
TWILLINGATE
ROUND POND
CROSS HILLS
HOLYROOD
LONG RANGE
Post Tectonic
Syn-Tectonic
Syn-Rift
PRE-APPALACHIAN
GRENVILLE BLOCK
CENTRAL BLOCK
AVALON BLOCK

RELATIONSHIPS BETWEEN GRANITIC ROCKS AND TECTONOSTRATIGRAPHIC ZONES.

Figure 1 portrays a marked spatial correspondence between Newfoundland plutonic rocks and tectonostratigraphic zones. Affiliations are clearest in the northeast where zones are wide and easily defined. Where zones are telescoped and difficult to define in southwestern Newfoundland, the correspondence between granites and zones is less pronounced.

The Long Range plutons (1-3, Fig. 1) are middle Proterozoic in age and an integral part of the Grenville basement. Their confinement to the Humber Zone or Appalachian miogeocline is therefore obvious. Granitic rocks range from massive coarse grained biotite granite with feldspar megacrysts, through hornblende-biotite granodiorite, to foliated orthopyroxene-bearing types of granulitic affinity (Owen, 1987). Coarse grained anorthosites and gabbros make up the Steel Mountain inlier (2) and much of the Indian Head inlier (3).

Holyrood plutons (4, 5) are the oldest of the Newfoundland Avalon Zone. These are composite plutons and are associated with late Precambrian volcanic rocks. The type example (4) is a calc-alkaline suite with a complete differentiation sequence of gabbro, granodiorite and granite (Strong and Minatides, 1975). It is overlain nonconformably by subhorizontal Cambrian strata. These plutons are an integral part of the pre-Appalachian, late Precambrian, Avalonian orogenic cycle that affected the Avalon Zone and correlative terranes throughout the North Atlantic borderlands (O'Brien et al., 1983).

The Round Pond-Cross Hills plutons are relatively small and of alkaline to peralkaline affinity. They occur in the eastern Humber Zone (6, 7) and western Avalon Zone (8, 9), respectively. Ages of the Humber Zone examples are defined isotopically (Williams et al., 1985; van Berkel and Currie, 1988) and the type example (6) is coeval with late Precambrian bimodal volcanic rocks and diabase dykes. All of these units are interpreted as synrift and tied to the initiation of the Appalachian cycle (Williams et al., 1985). The Cross Hills pluton of the Avalon Zone (8) may relate to the same rifting event. Ages there are unconstrained, except for a possible consanguineous relationship with nearby peralkaline Precambrian volcanic rocks (Hussey, 1979; O'Brien et al., 1986b), and diabase dykes that cut the plutons which are absent in nearby Lower Cambrian rocks. If the intervening Gander Zone is the opposing margin of the Iapetus Ocean, examples of similar plutons are expected be there.

Twillingate plutons (10-13) are restricted to the newly-defined Notre Dame Subzone of the Dunnage Zone, which requires further clarification in southwestern Newfoundland (Williams et al., 1988). These are large bodies of tonalite, trondhjemite and lesser granodiorite and granite, commonly associated with mafic volcanic rocks and affected by Ordovician deformation. Trondhjemites associated with ophiolite suites fall in this category; most are too small in size to be included on Figure 1. They are found in both the Notre Dame and Exploits subzones of the Dunnage Zone and occur as highly allochthonous examples above rocks of the Humber Zone. The Twillingate granite (10) cuts mafic pillow lavas, and mylonitic Twillingate granite is cut by mafic dykes dated at 470 Ma (Williams et al., 1976). Several other examples are overlain nonconformably by Silurian and Devonian rocks. Plagiogranites of this category are interpreted as anatectic products of subduction (Payne and Strong, 1979) and/or ophiolite differentiates. Their abundance in the Notre Dame Subzone and paucity in the Exploits Subzone supports the stratigraphic and structural evidence for the new twofold division of the Dunnage Zone (Williams et al., 1988).

The Burgeo (14-17) and Middle Ridge (18-20) plutons occur in all subzones of the Gander Zone. Locally, they transgress the Gander-Exploits boundary. The Burgeo plutons are foliated to mylonitic, medium to coarse grained, feldspar-phyric biotite granites that are pretectonic and syntectonic with respect to middle Paleozoic deformation (Dunning et al. 1988). The best examples form a virtually continuous belt extending along the eastern or outer margin of the Gander Lake Subzone and beyond to southwestern Newfoundland. They are everywhere associated with metamorphic rocks. The type example (14) is composite and includes early migmatitic and tonalite phases as well as later muscovite-bearing granites.

The Middle Ridge plutons vary from foliated to massive, medium grained equigranular to potassium-feldspar-phyric leucogranites with associated aplites and pegmatites. They contain muscovite, garnet, andalusite, sillimanite, tourmaline and locally beryl. They cut Burgeo-type plutons and are everywhere associated with high-grade metasedimentary rocks. Anatectic melts in surrounding metasedimentary rocks indicate a sedimentary source (Pickerill et al., 1978; Colman-Sadd, 1985). Two small occurrences in metasedimentary rocks of the eastern Humber Zone underscore their origin through local anatexis.

Isotopic ages, fabrics and crosscutting relationships among the Burgeo- and Middle Ridge-type plutons indicate an important middle Paleozoic structural and metamorphic event. It may be a deep expression of terrane convergence, suggested by the presence of Silurian melanges in the Exploits Subzone of Notre Dame Bay. The event is also synchronous, at least in part, with doming of the Gander subzones through their Exploits structural cover (Colman-Sadd and Swinden, 1984; Williams et al., 1988). The plutons are therefore equated with accretionary processes.

The remaining three categories of Newfoundland granitic rocks are mainly massive, undeformed equidimensional bodies (Mount Peyton, Topsails and Ackley). They cut local zone boundaries and therefore postdate Appalachian accretion.

The Mount Peyton plutons (21-23) are large composite bodies that contain early mafic phases and later granitic phases. They cut deformed Ordovician and Silurian rocks with thin hornfels aureoles and locally cut the Notre Dame-Exploits subzone boundary. Their marked preference for the northwestern portion of the Exploits Subzone is enigmatic.

The Topsails plutons (24-29) are mainly alkali granites coeval with overlying or surrounding Silurian volcanic rocks. The type example (24) consists of partly subvolcanic to peralkaline rhyolites (Whalen et al., 1987) and it is possibly related to the Silurian Springdale Caldera (Coyle and Strong, 1987). Another example of an unroofed Silurian caldera cuts the Ordovician Burlington granodiorite (12). The Devils Room granite (27) cuts Grenville gneisses of the Long Range inlier and is the westernmost middle Paleozoic intrusion in Newfoundland. The Cape Brule pluton (25) is deformed and metamorphosed with coeval volcanic rocks. The Wild Cove Pond pluton (29), tentatively included in this category, may be a Mount Peyton correlative; some of its porphyritic biotite granites resemble Ackley granites. These plutons define a north-south belt in western Newfoundland that clearly transgresses zone boundaries. The plutons follow a belt of major Silurian volcanism, possibly controlled by transcurrent faulting.

The Ackley plutons (30-36) are coarse grained, potassium-feldspar-phyric to equigranular biotite granites. Most examples are large and circular or have interlocking circular lobes. Features of high-level emplacement are common, and some examples contain molybdenite, tin, tungsten and fluorite mineralization. They cut the Gander-Avalon zone boundary and are restricted to a narrow boundary area. Paleozoic plutonism is unknown eastward across the wide Avalon Zone.

RELATIONSHIPS BETWEEN GRANITIC ROCKS AND LOWER CRUSTAL BLOCKS

Recent deep reflection studies off northeastern Newfoundland (Keen et al., 1986) and in the Gulf of St. Lawrence (Marillier et al., in press) define Grenville, Central and Avalon lower crustal blocks (Fig. 2). The Grenville block meets the Central block beneath the Dunnage Zone, roughly coincident with the Notre Dame-Exploits subzone boundary. The boundary between the Central and Avalon blocks coincides with the surface boundary between the Gander and Avalon zones.

The Long Range and Holyrood plutons are confined to their respective pre-Appalachian settings of the Grenville and Avalon lower crustal blocks on opposing sides of the orogen. The Round Pond and Cross Hills plutons also developed in these same opposing lower crustal blocks.

The Twillingate plutons occur above the Grenville block. Their ages, petrogeneses and associations suggest that most are allochthonous and confined to upper crustal levels, as are other rocks of the Dunnage Zone. Examples in southwestern Newfoundland may relate to Ordovician collisional processes.

The Burgeo and Middle Ridge plutons are entirely confined to the Central lower crustal block and extend to its western boundary in southwestern Newfoundland. Their absence in the Grenville and Avalon blocks implies a major plutonic/metamorphic episode confined to the Central block.

The post-accretionary Mount Peyton plutons are confined to western portions of the Central block, and the Topsails plutons to the Grenville block. Only the Ackley plutons clearly cross block boundaries, concentrated in the Central-Avalon boundary area.

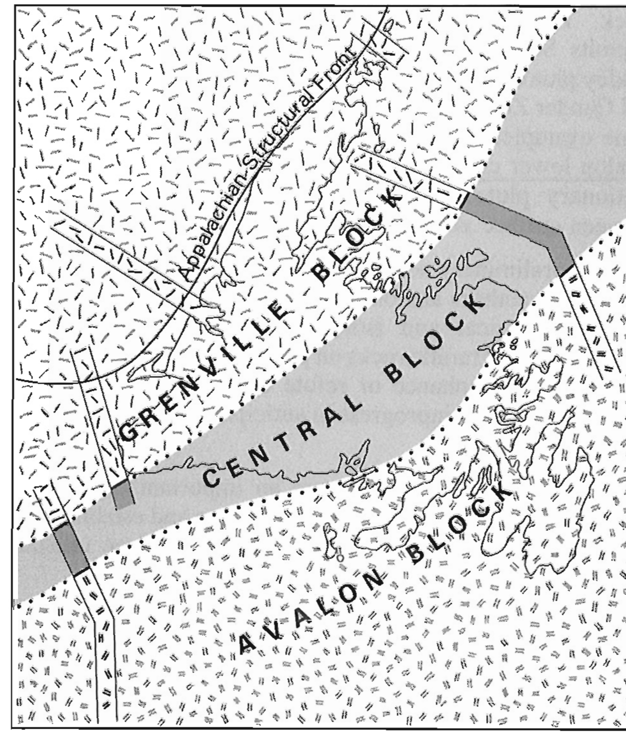


Figure 2. Deep crustal blocks of Newfoundland defined by marine seismic reflection transects (modified from Marillier et al., in press).

SUMMARY AND CONCLUDING REMARKS

Spatial relationships between the distribution of granitic rocks, tectonostratigraphic zones and lower crustal blocks imply genetic links and parallel tectonic and plutonic evolution of the orogen.

Plutons that predate Appalachian accretionary events are integral parts of zones and lower crustal blocks (e.g. Long Range, Holyrood, Round Pond-Cross Hills). Others (e.g. Twillingate) are allochthonous upon lower crustal blocks and, like ophiolite suites, their distribution provides a measure of tectonic transport within the orogen.

Plutons related to accretionary events (e.g. Burgeo and Middle Ridge), while locally transgressing zone boundaries, frequent subzones of the Gander Zone and are entirely within the confines of the Central lower crustal block. This pattern between plutons and surface subzones may reflect diapiric emplacement of plutons during doming of a lighter Gander crust through a heavier Dunnage cover (Toby Rivers, pers. comm., 1988). Furthermore, the correspondence between the plutons and the Central lower crustal block supports genetic links between the Gander Zone and Central lower crustal block.

Postaccretionary plutons (e.g. Topsails, Mount Peyton, Ackley), while transgressing zone boundaries, also show a preference for certain zones and/or lower crustal blocks. The Topsails plutons are confined to the Humber Zone and Notre Dame Subzone above the Grenville lower crustal block. The Mount Peyton plutons are confined to the Exploits Subzone and Central lower crustal block. The Ackley plutons are mainly confined to the Exploits Subzone and Gander Zone above the Central lower crustal block, but some examples cut nearby parts of the Avalon Zone and Avalon lower crustal block. A coherence between postaccretionary plutons and surface zones implies linkages between surface zones and deep crustal blocks.

This preliminary account is based mainly on field mapping by the authors and only a cursory assessment of available geochemical and isotopic data. Classification of Newfoundland granitic rocks on parameters other than those used here may enhance or refute our observations. New studies are already inprogress in anticipation of onland deep seismic experiments.

Studies of plutonic rocks are an important adjunct to unravelling the development of any orogen and establishing relationships between surface rocks and lower crustal blocks.

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Tectonic relationships along the proposed central Newfoundland Lithoprobe transect and regional correlations[1]

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Abstract

The proposed Lithoprobe transect of central Newfoundland, Meelpaeg Transect, crosses the Humber, Dunnage, Gander and part of the Avalon tectonostratigraphic zones. The transect also crosses the boundaries between Grenville, Central and Avalon lower crustal blocks. Clarification of surface zone boundaries, related kinematic studies and appraisals of metamorphic and plutonic zonal expressions provide insight in deciphering relationships between surface rocks and the geometry of lower crustal blocks.

Résumé

Le projet de transect Lithoprobe du centre de Terre-Neuve, le transect Meelpaeg, traverse les zones tectonostratigraphiques de Humber, de Dunnage et de Gander ainsi qu'une partie de celle d'Avalon. Il traverse aussi les limites séparant les blocs crustaux inférieurs de Grenville, du Centre et d'Avalon. La clarification des limites des zones en surface, des études de cinématique connexes et des évaluations des expressions zonales de nature métamorphique et plutonique permettent de déchiffrer les relations qui existent entre les roches superficielles et la géométrie des blocs crustaux inférieurs.

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INTRODUCTION

The prospect of an onland deep seismic reflection profile sponsored by the Canadian Lithoprobe Project, and the results of recent marine seismic transects (Keen et al., 1986; Marillier et al., in press) bring new and considerable interest to the Newfoundland Appalachians and relationships between surface rocks and lower crustal blocks. Deep crustal structure is being revealed for the first time and there are many ideas and hypotheses that need rethinking and testing, armed with the deep seismic results.

A project to subdivide the Dunnage Zone and clarify Dunnage-Gander relationships was commenced in 1987 (Williams et al., 1988). Field work continued in 1988 with a focus on structural, plutonic and metamorphic styles along the proposed central Newfoundland Lithoprobe route, or Meelpaeg Transect. Studies of kinematic indicators at major zone boundaries and in previously unknown shear belts within the zones (Piasecki, 1988, in press) are integrated in this project. Where exposure is poor, zone boundaries are projected and correlations are made with nearby better known areas.

The following account summarizes the results of field work conducted in 1988 and reviews current concerns and ideas that bear on tectonic relationships along the Newfoundland Meelpaeg Transect.

TECTONIC BOUNDARIES AND THE COURSE OF THE MEELPAEG TRANSECT

Zonal subdivision of Newfoundland and the course of the Meelpaeg Transect are shown in Figure 1. Burgeo Road is also shown as a possible ancillary reflection profile. The zonal subdivision is essentially unchanged from Williams et al. (1988), except that the Meelpaeg Subzone is extrapolated to the Cape Ray area. The main features and divisions along the Meelpaeg Transect are also depicted in Figure 2. From west to east, the Meelpaeg Transect crosses the Humber Zone, the Notre Dame and adjacent Exploits subzones of the Dunnage Zone, the Meelpaeg Subzone of the Gander Zone the southeastern portion of the Exploits Subzone, the Gander Lake Subzone of the Gander Zone and the western portion of the Avalon Zone. Most of the major tectonic zone/subzone boundaries are exposed within a few kilometres or less of the course of the transect. An important exception is the Baie Verte-Brompton Line (Humber-Dunnage boundary), which is hidden by Carboniferous rocks of the Deer Lake Basin. The Red Indian Line (Notre Dame-Exploits boundary) is exposed about 20 km southwest of the transect on the southeastern shoreline of Red Indian Lake. Noel Pauls Line (Exploits-Meelpaeg boundary in the northwest) is exposed in Noel Pauls Brook about 10 km southwest of the transect, and its southeastern counterpart near Meelpaeg Lake can be fixed within a few metres on the transect and observed 7 km to the south at Cold Spring Pond. Both the Exploits-Gander Lake and Gander Lake-Avalon boundaries are exposed on the transect in roadside pits and cuttings.

The Baie Verte-Brompton Line is projected onto the Meelpaeg Transect between Deer Lake and Grand Lake (Fig. 2). In its type area to the north at Baie Verte (Fig. 1), it is the Ordovician boundary between metaclastic rocks of the Fleur de Lys Supergroup of the Humber Zone (Appalachian miogeocline of Williams and Hatcher, 1982), and ophiolites such as the Birchy and Advocate complexes of the Dunnage Zone (Williams and St. Julien, 1982; Hibbard, 1983). Correlative metaclastic rocks (Mount Musgrave Group) and ophiolite complexes (Pynns Brook Complex) occur in an inlier surrounded by Carboniferous strata near the eastern side of Deer Lake (Williams et al., 1982, 1983). On the western side of Deer Lake, early Paleozoic rocks of the Humber Zone are overlain unconformably by Carboniferous strata. Green mafic pillow lavas near the southeastern shore of Grand Lake on the Hinds Lake access road are the most westerly exposures of Dunnage Zone rocks on the transect, except for those in the Humber Arm Allochthon (Fig. 1).

The Red Indian Line (Williams et al., 1988) occurs at or near the southeastern shore of Red Indian Lake. It separates the Buchans Group of the Notre Dame Subzone and Victoria Lake Group of the Exploits Subzone. The line is marked by isolated occurrences of red Silurian sandstones. Where well exposed south of the transect on the southeastern shore of Red Indian Lake, a steep brittle fault separates Silurian sandstones on the Notre Dame side and Ordovician volcanic rocks on the Exploits side. According to this definition, the line is fixed to within a few metres on the transect where similar red sandstones abut Ordovician volcanic rocks.

Noel Pauls Line separates Cambro-Ordovician sedimentary and volcanic rocks of the Victoria Lake Group from a monotonous sequence of interbedded shale and quartzite (Spruce Brook Formation), intruded by pretectonic mafic dykes and biotite granite (Colman-Sadd, 1987). Contrasts in the intensity of deformation, metamorphism and plutonism are all roughly coincident with Noel Pauls Line. On Noel Pauls Brook, a few kilometres to the southwest of the transect, pillow lavas of the Victoria Lake Group are represented by mafic bands in a wide zone of tectonically banded rocks that marks the line on the Exploits side. These abut thick bedded quartzites on the Meelpaeg side. Southwestward, sedimentary rocks of the Spruce Brook Formation are absent and biotite granite containing refolded mylonitic foliations and lineations lies adjacent to Victoria Lake Group pillow lava near Rodeross Lake (Colman-Sadd, 1988) and sedimentary rocks along the southeastern shoreline of Victoria Lake (Williams et al., 1988). In contrast to some later granite intrusions, this deformed granite is an integral part of the Meelpaeg Subzone and its distinctive lithology facilitates definition of the Exploits-Meelpaeg boundary.

The eastern boundary of the Meelpaeg Subzone with rocks of the Exploits Subzone (Great Burnt Lake Volcanic Belt) can be traced for over 50 km in an approximately north-south direction. If the proposed hydro road from the North Salmon dam to Island Pond is completed by the time the Vibroseis survey is done, the transect will cross the boundary north of Great Burnt Lake. Intensely deformed granite forms the eastern boundary unit of the Meelpaeg Subzone there, in contact with moderately deformed volcanic and sedimentary rocks of the Exploits Subzone along a fault that is locally marked by slices of ultramafic rocks (Swinden and Collins, 1982). An alternative route south of Great Burnt Lake crosses similar rocks, but some of the internal geology of the Meelpaeg Subzone is obscured by posttectonic granites and slices of indeterminate metasediment (Colman-Sadd, 1985). Farther south at Cold Spring Pond, there is almost continuous exposure from migmatized Spruce Brook Formation, across mylonitic granite of the Meelpaeg Subzone, into peridotite and sedimentary and volcanic rocks of the Exploits Subzone (Colman-Sadd, 1984). A sharp spur of Exploits rocks extends westward between two lobes of the Meelpaeg Subzone in this area, suggesting an original horizontal tectonic boundary.

From Great Burnt Lake, the line of the transect passes southward almost parallel to the Meelpaeg-Exploits boundary and then swings southeast across the Exploits Subzone toward the Gander Lake Subzone at Bay d'Espoir. In the central part of the Exploits Subzone it skirts the edge of an enigmatic body of migmatite at White Hill (Colman-Sadd, 1984). This may be a Gander Zone diapir penetrating an Exploits Subzone cover.

The Exploits-Gander Lake boundary has been studied in some detail by Piasecki (1988) along the well-exposed shoreline of Bay d'Espoir. Here it is a major ductile shear zone, represented in its northern part by the Day Cove Thrust (Colman-Sadd, 1980), which dips moderately to steeply toward the north-northwest. The Day Cove Thrust marks the contact between an assemblage of marine sedimentary and bimodal volcanic rocks metamorphosed in the greenschist and lower amphibolite facies (Exploits), and sheared upper amphibolite facies clastic sedimentary rocks, migmatites and granites (Gander Lake). The Meelpaeg Transect crosses the continuation of the shear zone about 20 km northeast of the coast (Colman-Sadd, 1976), where it is exposed in several roadside pits. In this area the Day Cove Thrust is steep, but there is no reason to believe that it is otherwise different from the exposures on the coast. Farther to the northeast there is a progressive decrease in metamorphic grade on both sides of the boundary that tends to reduce the contrast across it (Dickson, 1988), but the boundary still separates rocks characterized by volcanic lithologies on the Exploits side from monotonous clastic sedimentary rocks on the Gander Lake side.

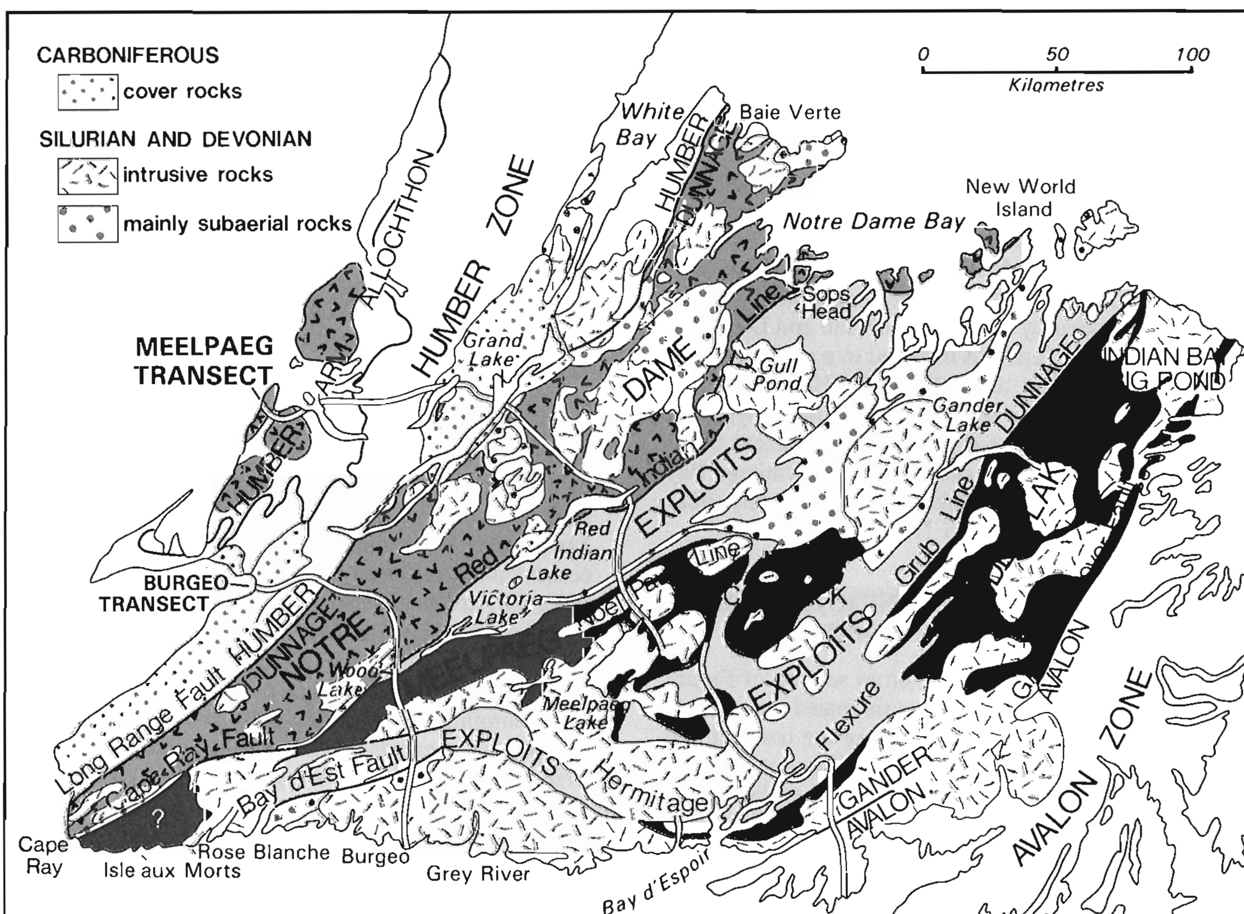


Figure 1. Zonal subdivision of central Newfoundland and the courses of possible Lithoprobe transects. Modified after Williams et al., 1988.

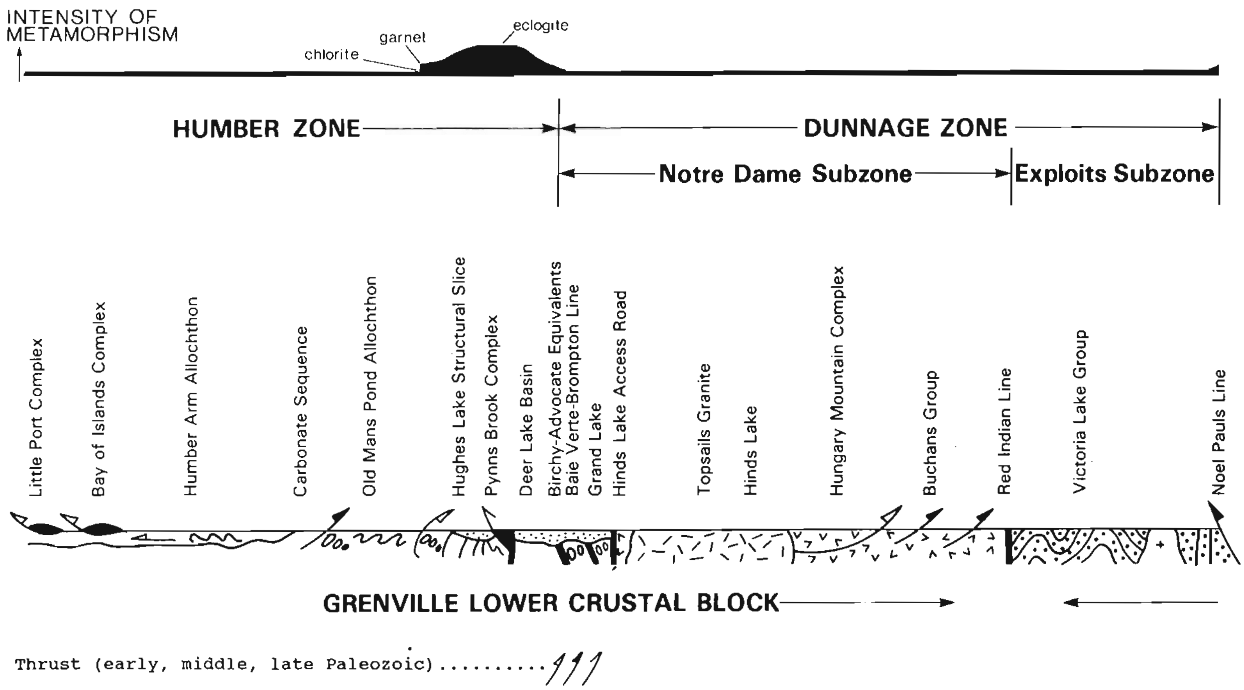


Figure 2. Zonal subdivisions and schematic geological relationships along the Lithoprobe Meelpaeg Transect.

The Gander Lake-Avalon boundary is the Hermitage Bay Fault, which is marked by a spectacular topographic lineament and separates Siluro-Devonian granites and high-grade metamorphic rocks of the Gander Lake Subzone from low-grade late Precambrian sedimentary and volcanic rocks that have been intruded by both Precambrian and Devonian granites in the Avalon Zone. As exposed in a roadcut on the transect (Colman-Sadd et al., 1979), the Hermitage Bay Fault is marked by a brittle breccia zone. Green mylonite, similar to that seen on the Dover Fault in northeastern Newfoundland is present along the course of the Hermitage Bay Fault about 20 km northeast of the transect (Dickson, 1988) and along a splay of the fault in Hermitage Bay.

STRUCTURAL STYLES AND KINEMATICS OF SOME ZONE/SUBZONE BOUNDARIES

Structural styles are depicted in the cross section of Figure 2 with polarities of tectonic transport indicated on some of the boundaries. Structural polarities along the transect and elsewhere in Newfoundland are shown in Figure 3. These are based on kinematic indicators such as S-C foliations, shear bands and rotated porphyroclasts (Berthe et al., 1979; Simpson and Schmid, 1983). Apparent conflicts in transport directions reflect changing kinematics as orogenesis progressed. Much remains to be done in establishing sequence and correlation among kinematic indicators, essential to understanding of the kinematic history.

Humber Zone

Across the Humber Zone, the transect is entirely within transported rocks of the Humber Arm Allochthon, the Old Mans Pond Allochthon and the Hughes Lake Structural Slice (Figs. 1 and 2). Locally, in the vicinity of Hughes Brook, parautochthonous carbonates form a narrow intervening area between the Humber Arm and Old Mans Pond allochthons. Toward the west, the Humber Arm Allochthon is believed to be a relatively thin (less than 2 km) tectonic cover above the carbonate sequence with a structural polarity toward the west (Williams et al.,1983). Farther east, the surface structures are more complex and varied. The Old Mans Pond Allochthon has a penetrative cleavage that dips consistently northwest and a regional order of stratigraphic units that are progressively younger to the southeast. The boundary between the Old Mans Pond Allochthon and the Hughes Lake Structural Slice dips gently northwest and within the Hughes Lake slice a well established sequence of east-younging stratigraphic units are steep to overturned. The pattern suggests that original upright sequences, which first moved west, were subsequently backfolded or thrust eastward. Mylonitic alkali granite and coeval volcanic rocks at the stratigraphic bottom of the Hughes Lake sequence have a lineation and S-C fabric indicating east over west polarity (overthrusting to the west) after dip restoration. Structural styles and their controls in this area are still debatable. One model suggests overturning of conventional west-directed thrusts into shapes resembling huge sled runners (Williams et al., 1982; 1983); another appeals to basement

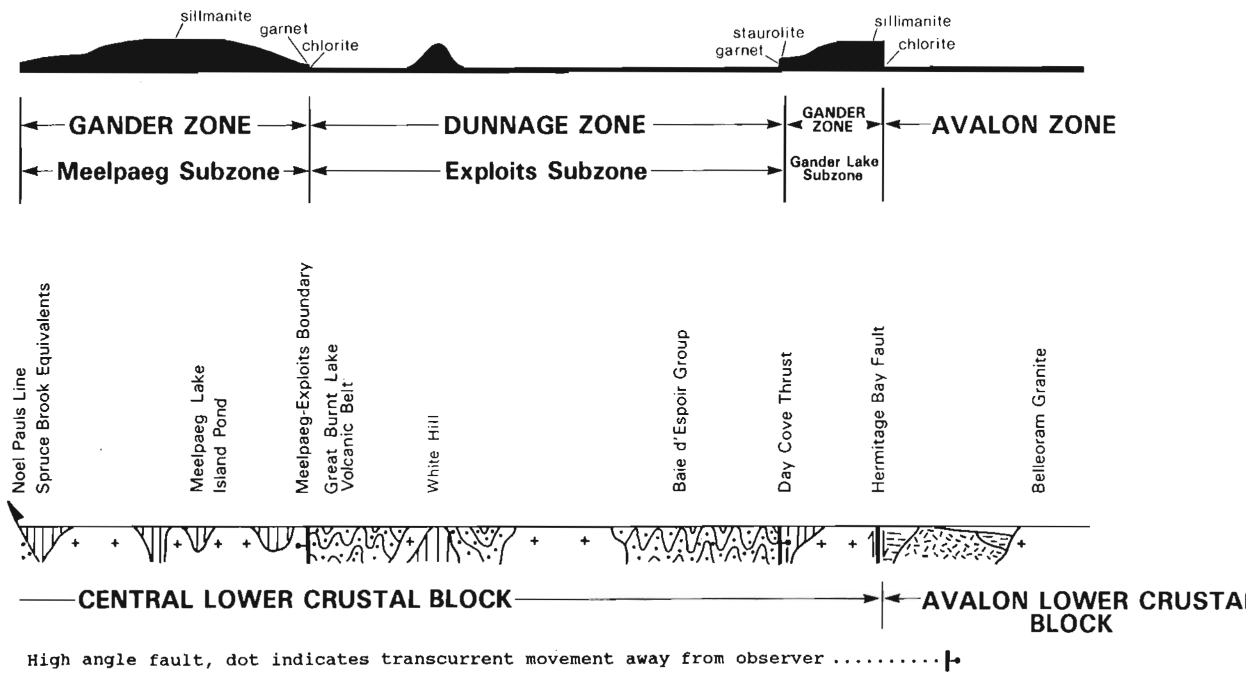


Figure 2 (cont.)

wedging beneath the overturned sections with surface shortening taking the form of eastward thrusting (Cawood and Williams, 1988); still another suggests that the Old Mans Pond Allochthon is a window beneath a structural cover that contains the carbonate sequence as well as the Long Range basement inlier (Waldron, 1988).

The Baie Verte-Brompton Line has a prolonged history of movement. Earliest movements of ophiolite complexes across the miogeocline have long been interpreted as east to west, and this is now supported by our observations of kinematic indicators in mylonitic rocks within the Birchy Complex. Later movements are west over east (overthrusting to the east) as best demonstrated by the Scrape Thrust and by rock relationships and mylonitic fabrics along the Baie Verte Road (Hibbard, 1983; Piasecki, in press). Kinematic indicators (S-C foliations and shear bands) in the Kidney Pond Conglomerate at Flatwater Pond and farther northward also confirm this later polarity of movement. Important Carboniferous faults trend northeast-southwest in the Deer Lake-Grand Lake region. Some are thrusts, such as the western boundary of the Pynns Brook inlier; some are high-angle faults downthrown on their western sides; others are interpreted as right-lateral transcurrent faults. The Long Range Fault, affecting Carboniferous rocks south of the transect, has a complex movement history in which early transcurrent movements (left-lateral) have been followed by brittle faulting downthrowing its western side.

Dunnage Zone

Both the Notre Dame and Exploits subzones of the Dunnage Zone are crossed by the Meelpaeg Transect. The Notre Dame Subzone was affected by early Paleozoic deformation and sub-Silurian or sub-Devonian unconformities abound. Nearby parts of the Exploits Subzone were apparently unaffected by early Paleozoic deformation in northeastern Newfoundland; Ordovician-Silurian sections are uninterrupted and all deformation episodes affect Lower and Middle Silurian rocks (Karlstrom et al., 1982).

Notre Dame Subzone

In the Notre Dame Subzone, structural styles across ophiolite belts of the Baie Verte Peninsula indicate original westward imbrication of east-facing units (Williams and St. Julien, 1982). Later structures are east directed (Hibbard, 1983; Piasecki, in press). Along the Meelpaeg Transect, eastward thrusting of the Hungry Mountain Complex onto the Buchans Group and eastward imbrication of the Buchans Group are the dominant structural styles. These east-directed structures extend southeast to the Red Indian Line (Calon and Green, 1987; Thurlow and Swanson, 1987). The structures are truncated by middle Paleozoic granites (Topsails) toward the north.

Southwestward of the transect at the south end of Grand Lake, the weakly metamorphosed clastics and ophiolite-volcanics of the Notre Dame Subzone are thrust southward over gneisses and granitoids locally metamorphosed in granulite facies and yielding Ordovician ages (Central Gneiss Terrane of van Berkel and Currie, 1988 and K.L. Currie, pers. comm., 1988). Whether or not rocks of the Central Gneiss Terrane are Notre Dame or Humber equivalents is debatable. The Central Gneiss Terrane itself appears to comprise smaller miogeoclinal(?) units assembled along ductile shear zones containing ophiolitic remnants (Fox and van Berkel, 1988; Piasecki, in press). If miogeoclinal rocks there were structurally comingled with ophiolite occurrences, then all intruded, metamorphosed and polydeformed together, a Humber-Dunnage distinction is impractical.

Exploits Subzone

Structural styles across the Exploits Subzone between the Red Indian Line and Noel Pauls Line are poorly known and exposure is scant. Northeast of the transect where the Exploits Subzone is wider, map patterns indicate relatively open upright folds about axes that plunge northeast (Kean et al., 1981). Between the Meelpaeg Subzone and the Gander Lake Subzone, sedimentary and volcanic rocks of the Exploits Subzone (Baie d'Espoir Group) display tight to isoclinal upright folds refolded by later, generally open recumbent structures (Colman-Sadd, 1980).

The Red Indian Line has surprisingly little structural expression in adjacent rocks and is almost everywhere a steep brittle rectilinear fault.

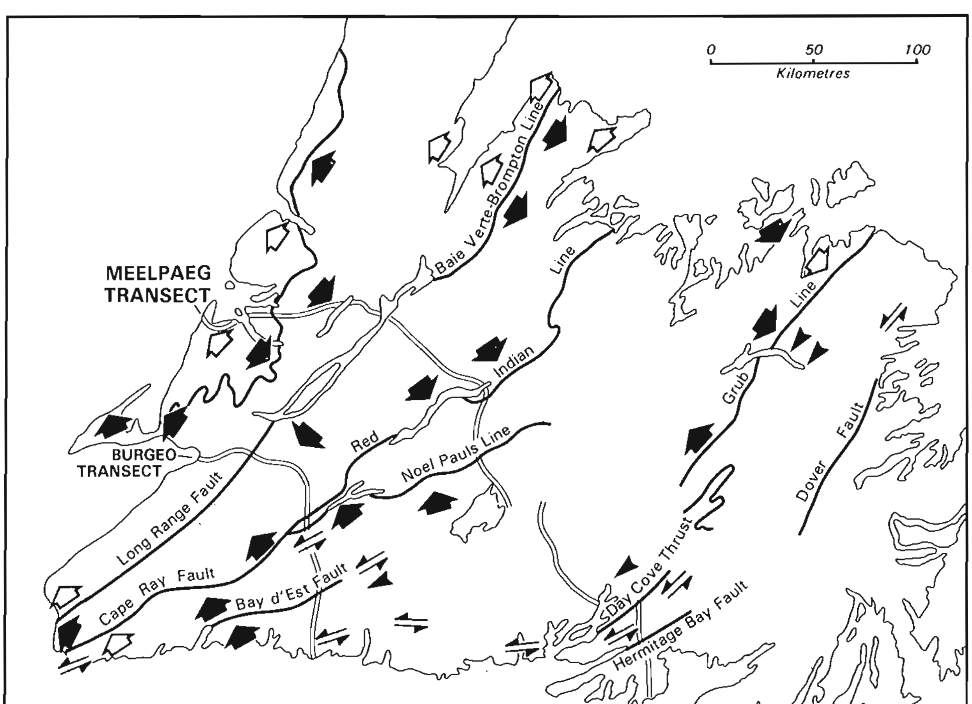
South of the Meelpaeg Transect on the Burgeo Road, the Red Indian Line is absent where gabbros cut welded tuffs of the Victoria Lake Group. Some of these rocks, where banded and highly indurated, were interpreted as mylonites (Williams et al., 1988). Southwestward along strike at the Cape Ray Fault, there are sequential ductile movements: southeast over northwest thrusting in mylonitic granites on the Notre Dame side that predate middle Paleozoic volcanic rocks of the Windsor Point Group, and later east over west oblique thrusting that produced mylonites in the Windsor Point Group. The polarity of transport corresponds to that along Noel Pauls Line (Fig. 3), and also corresponds to northwest overthrusting in the Silurian La Poile Group about 60 km to the east (O'Brien, 1988).

Gander Zone

Both the Meelpaeg and Gander Lake subzones of the Gander Zone are crossed by the Meelpaeg Transect (Figs. 1 and 2). Of prime importance here is the interpretation of the Meelpaeg Subzone as a structural inlier, or window of Gander Zone rocks exposed through an Exploits Subzone structural cover (Colman-Sadd and Swinden, 1984; Williams et al., 1988).

The Meelpaeg Subzone consists mainly of sillimanite-grade rocks, but deformation is similar in style and intensity to that in lower grade rocks of the Exploits Subzone. The earliest fabric in metasedimentary rocks is commonly steep and subparallel to bedding, since associated folds are tight to isoclinal (Colman-Sadd, 1985). This fabric is also developed in biotite granite that intrudes the sediments. Second generation folds vary from open to isoclinal with the local development of crenulation cleavage (Colman-Sadd, 1987).

Deformation in the Meelpaeg Subzone increases in intensity toward its boundaries with the Exploits Subzone, even though the metamorphic grade decreases in places, such as along the transect at Noel Pauls Brook. Within about 5 km of Noel Pauls Line, the first fabric contains an east-west lineation (allowing for unfolding of later structures) and is commonly mylonitic, especially in pretectonic biotite granite. Second deformation folds of the first fabric in the biotite granite plunge moderately to the northeast and have wavelengths of several kilometres. The same folds affect metasedimentary rocks and they are isoclinal adjacent to Noel Pauls Line. At the line, bedding and tectonic fabrics



Thrust polarity of upper plate (early, middle Paleozic)
Sense of shearing in steep shear zone ....
Polarity of upper plate in flat shear zone ....

Figure 3. Polarity of tectonic transport in central Newfoundland.

dip moderately southeast, and the contained lineation plunges gently south (Colman-Sadd, 1987), an attitude that may result from the exposed part of the line being located on the overturned northwestern limb of a regional second deformation antiform. Kinematic indicators related to the first mylonitic fabric indicate a southeast over northwest sense of displacement, that is northwest thrusting of Meelpaeg over Exploits. However, if the effects of the second folds are removed, northwest thrusting of Exploits over Meelpaeg is indicated.

On the eastern side of the Meelpaeg Subzone, there is a progressive increase in intensity of second deformation. Second folds in metasedimentary rocks are isoclinal and extremely flattened within 5 km of the Great Burnt Lake Fault (Colman-Sadd, 1985). A third set of tight to open folds with an associated crenulation cleavage is also developed in this area. A long narrow body of megacrystic biotite granite occurs at the eastern edge of the Meelpaeg Subzone, trending north-south for some 60 km but nowhere more than 7 km wide. Its eastern boundary is the Great Burnt Lake Fault and the granite does not intrude rocks of the adjacent Exploits Subzone. The granite has a steep planar fabric that strikes north-south and a subhorizontal lineation (Colman-Sadd, 1985). The relationships imply north-south transcurrent movement, in contrast to the northwest overthrusting at Noel Pauls Line. Whereas the interpretation of northwest movement on Noel Pauls Line is based on the earliest recognized structures, the north-south transcurrent movement at Great Burnt Lake is a second deformation feature. The large granite and granodiorite plutons that underlie the central and eastern portions of the Meelpaeg Subzone are northern lobes of the North Bay batholith. These are posttectonic with respect to first and second deformations.

The equivalent Meelpaeg-Exploits boundary on the Burgeo Road (Fig. 1) is marked by steeply inclined mylonites with a subhorizontal lineation indicating a right-lateral sense of transcurrent displacement. The mylonites contrast in metamorphic grade and structural style with shales of the Bay du Nord Group on the Exploits side.

The boundary between the Exploits and Gander Lake subzones is the Day Cove Thrust, which is traceable northeast across the transect from Bay d'Espoir (Colman-Sadd, 1976). Structural analysis of this boundary in Bay d'Espoir shows it to be a zone of ductile shearing that includes metasedimentary and migmatitic rocks with syntectonic metamorphism and plutonism between the Day Cove Thrust and the Hermitage Bay Fault (Piasecki, 1988). Abundant kinematic indicators at the gently inclined contact show that rocks of the Exploits Subzone moved west-southwest, parallel to the regional strike. Along strike to the northeast and across strike to the southeast, foliations are steep, but where kinematic indicators have been examined the sinistral sense of movement is the same. Strikingly similar shear geometry and similar intimate relationships between shearing, metamorphism and plutonism also characterize a wide area of the Gander Lake Subzone in its type area to the north (Hanmer, 1981).

Avalon Zone

Only a narrow western portion of the Avalon Zone is crossed by the transect. Its rocks and structures are completely different from those of the adjacent Gander Lake Subzone. Late Precambrian granite that cuts volcanic rocks is overlain nonconformably by Silurian-Devonian sedimentary rocks, in turn cut by middle Paleozoic granites (Williams, 1971). Folds in both Precambrian and Paleozoic rocks are upright about northeast-trending axes.

The Avalon-Gander Lake boundary is the Hermitage Bay Fault (Blackwood and O'Driscoll, 1976; Kennedy et al., 1982). Although the fault is represented by a breccia zone on the line of the transect, mylonitic rocks occur elsewhere. A detailed study of the movement history of the equivalent boundary in northeastern Newfoundland, the Dover Fault, is the topic of a separate study. Preliminary results indicate a complex history of ductile transcurrent and later vertical movements followed by brittle movements (Caron and Williams, 1988).

METAMORPHISM ALONG THE MEELPAEG TRANSECT

Regional metamorphism in rocks along the transect is depicted schematically in Figure 2. Western parts of the Humber Zone are in lower greenschist facies. Eastern parts are in upper greenschist to amphibolite facies between the Hughes Lake Structural Slice and the Baie Verte-Brompton Line. The higher grade rocks are brought to the surface and juxtaposed with lower grade rocks by deep thrusts. Rocks of the Dunnage Zone are everywhere lower grade, greenschist or less, except in the aureoles of plutons and along boundaries with the miogeocline and Gander Zone. Gander Zone inliers in the Exploits Subzone contain high-grade metamorphic rocks. No equivalent inliers are known in the Notre Dame Subzone, except possibly for the Central Gneiss Terrane (van Berkel and Currie, 1988) that may represent a Humber Zone window through the Notre Dame Subzone. Amphibolite grade metamorphism in the Gander Lake Subzone terminates abruptly at the Hermitage Bay Fault. Paleozoic metamorphism in the Avalon Zone is of greenschist or lower grade.

At the Baie Verte-Brompton Line, metamorphism decreases in rocks of the Humber Zone and increases in rocks of the Dunnage Zone, although the intensity of deformation everywhere increases toward the tectonic boundary. Eclogite pods, retrograded at their margins and surrounded by psammitic schists in amphibolite facies occur west of the Humber-Dunnage boundary at Baie Verte Peninsula. At the Baie Verte-Brompton Line, psammitic rocks of the Humber Zone and ophiolites and melanges of the Dunnage Zone are of upper greenschist to lower amphibolite facies. Metamorphism decreases eastward across a series of imbricated ophiolite suites at the western margin of the Dunnage Zone (Williams and St. Julien, 1982).

In the Gander Zone as a whole, and especially in the Meelpaeg and Gander Lake subzones, amphibolite facies

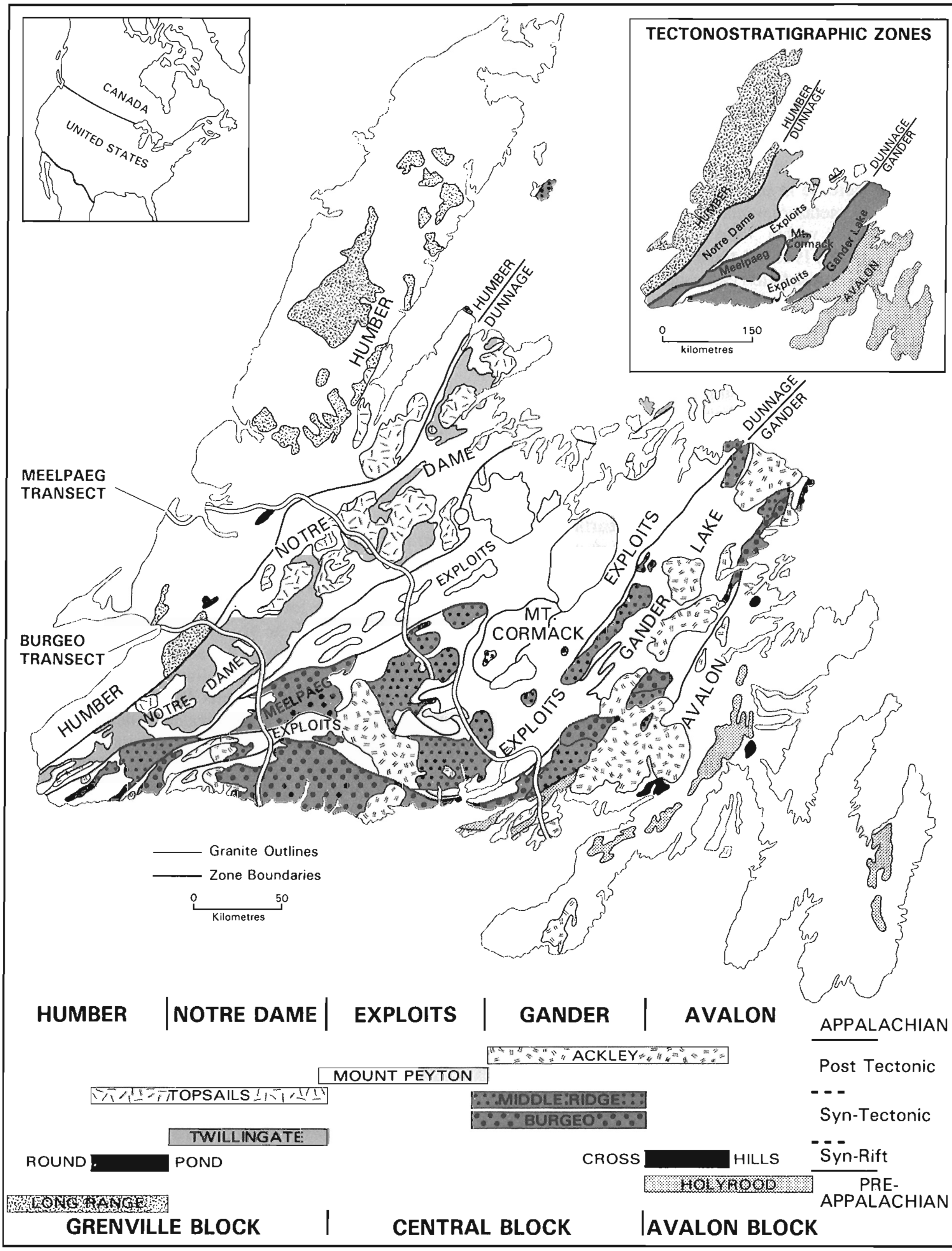


Figure 4. Distribution of plutonic rocks with respect to tectonostratigraphic zones, and location of Meelpaeg and Burgeo Lithoprobe transects (after Hayes et al., 1987; Williams et al., 1989).

metamorphism is widespread, and large areas of migmatite are unrelated spatially to granitoid plutons (Colman-Sadd, 1980, 1984, 1985, 1988; Colman-Sadd and Swinden, 1984). The wide range of metamorphic temperatures, independent of plutons, is characteristic of Gander Zone metamorphism and contrasts with the uniformity of low-grade conditions that prevailed in the Exploits Subzone. The normal progression is from andalusite to sillimanite-muscovite assemblages, with partial melting occurring at the second sillimanite isograd (Colman-Sadd and Swinden, 1984). Kyanite occurs in the Cape Ray area (Brown, 1977) and the northern part of the Gander Lake Subzone (Kennedy, 1976).

Along the line of the Meelpaeg transect, there is no sharp jump in metamorphic grade from the Exploits subzone to the Meelpaeg Subzone across Noel Pauls Line. Rocks on both sides are of greenschist facies. However this is exceptional and farther southwest at Victoria Lake, the Meelpaeg Subzone has foliated granite and garnetiferous psammites at the contact. Within the Meelpaeg Subzone, metamorphic grade rises rapidly southeastward along the transect, where muscovite-sillimanite grade metasedimentary rocks are intruded by biotite granite and granodiorite (Colman-Sadd, 1985, 1987).

On the eastern side of the Meelpaeg Subzone, there is some decrease in grade toward the contact north of Great Burnt Lake and sillimanite is generally absent. However, garnetiferous rocks of the Meelpaeg Subzone contrast with chlorite grade sedimentary and volcanic rocks of the Great Burnt Lake Volcanic Belt (Colman-Sadd, 1985). At Cold Spring Pond, some 15 km farther south, the contrast is much more marked, with migmatite of the Meelpaeg Subzone separated from greenschist facies sedimentary and volcanic rocks by a narrow band of mylonitic granite (Colman-Sadd, 1984).

Southeast from the Great Burnt Lake Fault, the transect passes through Exploits Subzone rocks mainly of greenschist facies, except for a possible Gander diapir of migmatitic sillimanite-grade metasedimentary rocks at White Hill. Elsewhere Exploits rocks only exceed greenschist facies in narrow aureoles around plutons and in a narrow zone near their southern boundary with the Gander Lake Subzone.

At the Day Cove Thrust, there is a sharp jump in metamorphism from garnet grade on the Exploits side to muscovite-sillimanite grade on the Gander Lake side. Less than 10 km northeast of the transect the contrast diminishes and greenschist facies rocks occur on both sides of the contact (Dickson, 1988). However, in all parts of the Gander Lake Subzone there is a metamorphic progression southeastward that culminates in migmatitic gneisses intruded by syntectonic granitoid plutons (Blackwood, 1978; Colman-Sadd, 1980; Dickson, 1988; Piasecki, 1988).

The juxtaposition of migmatites of the Gander Lake Subzone and greenschist or lower grade Precambrian rocks of the Avalon Zone is one of the sharpest metamorphic breaks along the transect, and emphasizes the importance of the Hermitage Bay Fault and possible precursors.

The controls and timing of metamorphism are complicated and beyond the scope of this cursory treatment. All of the metamorphic rocks are truncated by middle Paleozoic plutons. Present understanding is that the chief metamorphism in the Humber Zone is Taconic or early Paleozoic, with local strong middle Paleozoic overprinting. Metamorphism in the Meelpaeg and Gander Lake subzones may be entirely middle Paleozoic and preliminary isotopic ages of syntectonic and posttectonic granites suggest Silurian metamorphism (Dunning et al., 1988).

In the Exploits Subzone, Silurian Botwood Group sediments seem to have been affected by the same events as the Cambrian and Ordovician rocks. The Avalon Zone has a Precambrian history of deformation but isotopic dates of muscovites indicate that present metamorphic character was established in the middle Paleozoic (Dallmeyer et al., 1983).

PLUTONISM ALONG THE MEELPAEG TRANSECT

Relationships between plutons and zones along the Meelpaeg and Burgeo transects are summarized in Figure 4. Plutonic rocks are absent in the Humber Zone except for the Round Pond Granite of the Hughes Lake Structural Slice that is late Precambrian and interpreted as a rift-related pluton consanguineous with bimodal late Precambrian volcanic rocks (Williams et al., 1985).

In the Notre Dame Subzone, the transect crosses large middle Paleozoic granites of the Topsails type. These are alkali plutons, mainly of Silurian age, and are coeval with Silurian volcanic rocks (Coyle and Strong, 1987; Whalen et al., 1987; Williams et al., 1989).

Plutonic rocks are not exposed in the Exploits Subzone between the Red Indian Line and Noel Pauls Line, but nearby examples are mainly diorites and gabbros of Ordovician to Devonian age (Kean et al., 1981; Colman-Sadd, 1988). Southeast of Noel Pauls Line in the Meelpaeg Subzone, foliated biotite granite is cut by coarse grained porphyritic biotite granodiorite (Ackley type) and garnetiferous muscovite leucogranite (Middle Ridge type). Similar plutons occur in the Exploits and Gander Lake subzones (Colman-Sadd, 1980, 1984; Blackwood and Green, 1982). The Avalon Zone has a late Precambrian history of plutonism related to its volcanic rocks (O'Driscoll and Strong, 1979).

Devonian coarse grained porphyritic biotite granite of Ackley type truncates both the Dunnage-Gander and Gander-Avalon zone boundaries.

DEEP STRUCTURE AND IMPLICATIONS OF SURFACE FEATURES

The seismic definition of lower crustal blocks off northeastern Newfoundland (Keen et al., 1986) and in the Cabot Strait off southwestern Newfoundland (Marillier et al., in press) constrains deep structure beneath onland transects (Fig. 5). Off northeastern Newfoundland, a Grenville Block abuts a Central Block at deep crustal levels beneath the central portion of the Dunnage Zone. The Grenville-Central block boundary is also identified off southwestern Newfoundland, at or near the projection of the Cape Ray Fault. This and results elsewhere in the Gulf of St. Lawrence indicate an allochthonous Dunnage Zone above a deep suture between Grenville and Central lower crustal blocks. The Central-Avalon lower crustal block boundary is imaged on marine reflection profiles as steep and coinciding with the Gander-Avalon zone boundary.

The Baie Verte-Brompton Line, although unexposed on the transect, is one of the most significant boundaries in the Appalachian Orogen. It delimits the exposed edge of the miogeocline and in most places it is marked by occurrences of ophiolitic suites and melanges. Its setting on the Meelpaeg Transect is surmised to be like that in its type area at Baie Verte. There the line is underlain by miogeoclinal rocks of the Humber Zone that extend several tens of kilometres eastward in the subsurface (Keen et al.,1986).

The Grenville-Central lower crustal boundary is roughly coincident with the surface trace of the Red Indian Line. A suture between lower crustal blocks is therefore anticipated on the Meelpaeg Transect in the vicinity of Red Indian Lake. The Red Indian Line is traceable from Notre Dame Bay to Burgeo Road (Fig. 1). Farther southwest the Exploits Subzone is absent and Noel Pauls Line converges with the Red Indian Line. From there both project southward toward the Cape Ray Fault. Examination of the Cape Ray area in 1988 reveals psammitic schists characterized by a striking development of concordant amphibolite bands on all scales extending eastward from the Cape Ray Fault to a ductile shear zone at Isle aux Morts, and psammitic schists relatively devoid of amphibolite extending from Isle aux Morts eastward toward Rose Blanche. Similar rocks on the transect in the Meelpaeg Subzone contain amphibolites derived from mafic dykes, but their more easterly equivalents in the Mount Cormack Subzone are virtually devoid of such intrusions. A possible equivalence of rocks of the Meelpaeg and Mount Cormack subzones with those in southwestern Newfoundland implies that the Cape Ray Fault, the Red Indian Line and Noel Pauls Line are all colinear in southwestern Newfoundland.

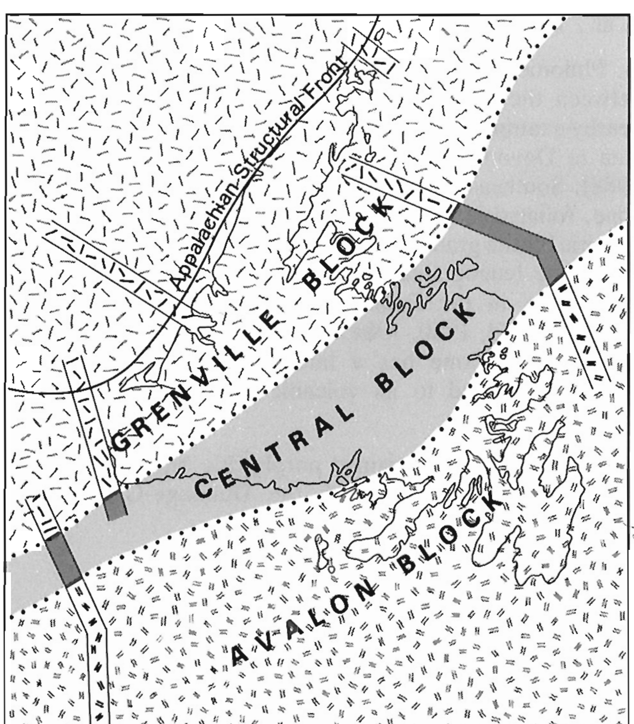


Figure 5. Lower crustal blocks of the Newfoundland Appalachians (from Marillier et al., in press).

CONCLUDING REMARKS

A present concern stemming from marine deep seismic experiments and clarification of surface zonal divisions is whether or not rocks of the Gander Zone are linked to the Central lower crustal block or are allochthonous across the Central lower crustal block. The interpretation of all Gander boundaries as tectonic divorces rocks of the Gander Zone from their surrounding Dunnage cover, permitting stratigraphic links with the Central lower crustal block.

The correspondence among tectonostratigraphic zones, plutonic rocks and lower crustal blocks suggests genetic links between all. Plutons that relate to rifting (Round Pond, Cross Hills) and subsequent constructive processes of orogenic development (Twillingate) are coincident with tectonostratigraphic zones. Younger plutons that relate to destructive processes are more likely to cross zone boundaries and reflect lower crustal blocks. Thus the localization of Topsails plutons northwest of the Red Indian Line and of the Burgeo, Middle Ridge and Ackley plutons to the southeast of the Red Indian Line suggests different underlying crustal blocks.

The timing, styles and mechanisms of accretion are clearly different across the orogen. The earliest accretionary boundaries affecting the miogeocline are marked by ophiolites and melanges. Later boundaries between outboard terranes are ductile mylonite zones or brittle faults. Kinematic studies in progress strive to determine the sense of displacement on tectonic boundaries and their relationship in time, and so provide some insight into mechanisms of accretion and subsequent deformation.

Models of the Mount Cormack and Meelpaeg subzones as Gander inliers through a Dunnage cover imply a two-layer crust southeast of the Red Indian Line. While the expectable polarity of initial Dunnage transport is east or southeast, present kinematic studies indicate left-lateral shear throughout the Gander Lake Subzone and southwestward transport of the Dunnage Zone at the Exploits-Gander Lake boundary (Fig. 3). Furthermore, relationships at Noel Pauls Line imply northwest polarity of Dunnage transport, if mylonites at the line relate to Dunnage-Gander initial superpositioning. Timing is of the essence in these analyses. The earliest mylonitic fabrics may be much later than initial Dunnage-Gander superpositioning. Sub-Caradoc or sub-Landeilo unconformities above ophiolitic rocks at the Exploits-Gander Lake boundary in northeastern Newfoundland indicate an Ordovician disturbance and first deformation of Dunnage ophiolite complexes. Isotopic ages suggest that ductile shearing and coeval plutonism at present boundaries are later (Dunning et al., 1988). In western Newfoundland, the earliest transport of ophiolitic Dunnage rocks across the miogeocline caused no penetrative deformation. A similar situation may have existed on the opposite side of the orogen.

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Near-surface seismic reflection studies of the Jeanne d'Arc Basin, northeastern Grand Banks of Newfoundland[1]

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Abstract

About 1200 km of high resolution seismic reflection profiles were obtained in the northeastern sector of Grand Bank to study the seafloor sediments. Data were collected using a multitipped 16 kJ sparker and a 400 m 24-channel streamer; a Huntec Deep Towed Seismic subbottom profiler; a 1 kJ sparker with a 7.6 m streamer; and sidescan sonars. Survey lines were chosen to tie in as many of the existing well site surveys and borehole locations as possible and constitute a regional network for the integration of multichannel shallow seismic data obtained earlier at surveys performed to study conditions at local well sites. Our objective is to use the combined data set to develop a comprehensive seismostratigraphic model. Four distinct reflection patterns have been described which characterize the shallow seismic sections from northeastern Grand Bank. These patterns are (1) parallel reflections, (2) clinoform reflections associated with a large relict progradational delta at the Hibernia site, (3) vertically displaced reflections and (4) channel structures. Recent data substantiate this subdivision.

Résumé

Dans le secteur nord-est des Grands Bancs, on a réalisé environ 1200 km de profils de sismique-réflexion, de résolution élevée, pour étudier les sédiments du fond marin. On a recueilli des données en employant un étinceleur de 16 kJ à plusieurs pointes et une flûte marine de 400 m de long à 24 voies; un profileur sismique Huntec du fond marin et des terrains proches du fond, remorqué à grande profondeur; un étinceleur de 1 kJ avec une flûte marine de 7,6 m de long; et des sonars à balayage latéral. On a choisi les lignes de levés de façon à regrouper le plus grand nombre possible des levés de puits de sondage existants et le plus grand nombre de points de forage; ces lignes constituent un réseau régional qui servira à intégrer les données sismiques obtenues à faible profondeur avec un système multivoies lors de levés accomplis pour étudier les conditions régnant dans les sites locaux de puits de sondage. On se propose d'utiliser des données combinées pour élaborer un modèle sismostratigraphique compréhensif. On a décrit quatre schémas distincts de réflexion qui caractérisent les coupes sismiques profondes du nord-est des Grands Bancs. Ces schémas sont (1) des réflexions parallèles, (2) des réflexions clinoformes associées à un grand delta progradant résiduel existant sur le site d'Hibernia, (3) des réflexions déplacées verticalement (4) des structures en forme de rigole. Les données récentes confirment le bien-fondé de cette subdivision.

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INTRODUCTION

Quarternary sediments on northeastern Grand Bank are discontinuous and thin (<5 m). The underlying Tertiary sediments constitute the foundation for petroleum development structures. The Tertiary sediments must support vertical stresses due to these structures and resist horizontal loads imposed by waves, currents and ice. A compilation of available information on the surficial and near-surface bedrock geology and the processes affecting their stability provides a regional framework for use in the interpretation of the extent and variability of features identified at site-specific surveys and at boreholes performed for the design of oil production and transportation facilities on the northeastern Grand Banks of Newfoundland.

Previously collected high resolution seismic reflection information from this area has been compiled into a digital database, and new singlechannel and multichannel seismic reflection transects were collected across the shelf edge of northeastern Grand Bank in 1988 (Fig. 1).

The deeper structure within the Tertiary section and tectonic elements of the underlying Jeanne d'Arc Basin have been studied in detail using conventional exploration seismic reflection data (Arthur et al., 1982; Grant et al., 1986). While structures and features in the deeper section are seen to propagate upward into the shallow Tertiary sediments, vertical resolution of the deep seismic data is insufficient for interpretation of details of the shallow seismostratigraphy.

The available single channel seismic reflection data have been analyzed to provide a regional interpretation of the near-surface Tertiary seismostratigraphy. The interpreted seismic sections were digitized, and major reflections were assigned a unique identifier code, as part of a digital seismostratigraphic database. With this database, individual reflections can be easily compared and correlated, allowing calculation of depths to reflections or thicknesses of units between significant reflections for the generation of maps and sections using a computer-aided drafting (CAD) system.

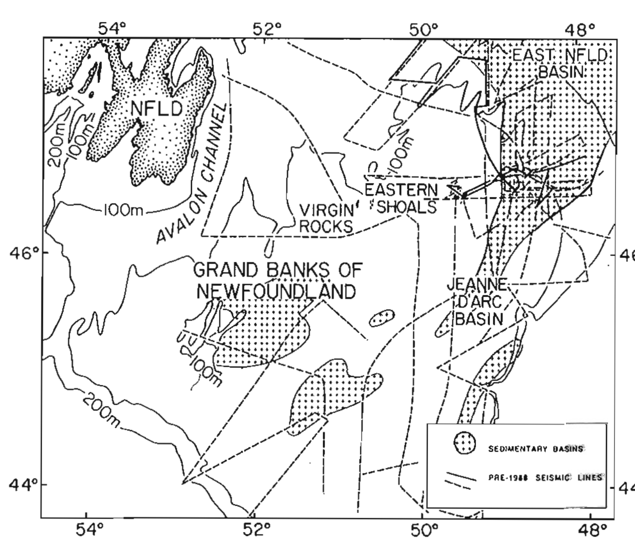

Figure 1. Map of portion of the Grand Banks of Newfoundland showing bathymetry, sedimentary basins and location of pre-1988 shallow seismic reflection lines.

In this paper we present highlights of new seismic data for northeastern Grand Bank. Selected profiles illustrate an interpretation of the stratigraphy and structure of the shallow Tertiary sediments with respect to their depositional environments and relationship to deep-seated structures based on earlier work by Lewis et al. (1987; unpublished report, 1988). The 1988 profiles constitute a regional network for the correlation of multichannel shallow seismic data obtained earlier at local well sites. Our objective is to ultimately use the combined data set or regional and local (well site) seismic sections for a comprehensive seismostratigraphic model of the northeastern sector of Grand Bank.

REGIONAL SETTING

The regional framework for the Quaternary, surficial and shallow bedrock geology of the Grand Banks of Newfoundland is known from studies by King and MacLean (1975), Fader and King (1981), Fader et al. (1982), Barrie et al. (1984, 1986), Fader et al. (1985), Fader (1986), King and Fader (1986), Fader and Miller (1986a,b), King et al. (1986), and Lewis et al. (1987). The surface of the Grand Banks slope gently from shallow water (<5 m) at Virgin Rocks and Eastern Shoals near the western flank of the bank to about 100 m at the eastern edge (Fig. 1). The surface of the northeastern sector of Grand Bank is veneered with thin discontinuous deposits of sand and gravel (Grand Banks Sand and Gravel) which directly overlie a prism of Cretaceous-Tertiary bedrock (Fader and King, 1981). The Tertiary clastics thicken seaward from the area of the Virgin Rocks to more than 4000 m in the East Newfoundland Basin.

A series of north-south fault-bounded Mesozoic basins are separated in the subsurface by basement highs (Amoco and Imperial, 1973; Jansa and Wade, 1975; Grant et al., 1986). The East Newfoundland Basin is the most northerly depocentre underlying the Grand Banks and the northeastern Newfoundland Shelf (Fig. 1), and contains sediments over 15 000 m thick. The southern extension of this basin is the Jeanne d'Arc Basin where all of the hydrocarbon discoveries to date, including Hibernia, Terra Nova, Ben Nevis and Whiterose, have been made (Arthur et al., 1982; Grant et al., 1986).

Fader et King (1981) identified reflections, gently dipping to the east, on airgun seismic reflection profiles from northeastern Grand Bank, and interpreted these as representing Tertiary sediments. The Tertiary reflections are truncated at an unconformity which is overlain by a thin discountinuous cover of Quaternary sediments. Numerous infilled channels, cutting Tertiary beds to depths of 300 m, occur on the southeastern margin of Grand Bank (Fader and Miller, 1986b). Grant (1972) showed the presence of buried channels to similar depths on the northwestern margin of Grand Bank. Only one area of infilled channels is known on northeastern Grand Bank (G. Vilks (ed.), unpublished report, 1984; Fader and Miller, 1986b; Lewis et al., 1987).

Fader and Miller (1986b) also note widespread wedge-shaped zones of clinoform reflections between major parallel reflections on seismic profiles of the Tertiary section across southeastern Grand Bank. Some of these clinoform reflections dip north.

Lewis et al. (1987; unpublished report, 1988) described four distinct reflection patterns which characterize the shallow airgun seismic reflections sections from northeastern Grand Bank. These patterns are:

(1) Parallel reflections. Seismic sections on the northeastern Grand Banks are generally characterized by a succession of medium to high amplitude continuous parallel reflections which generally dip east-northeastward at less than 0.5 degrees.

(2) Clinoform reflections. Overlapping 5-shaped or oblique clinoform reflections which occur within the general sequence of parallel reflections were interpreted as a prograding delta and termed the Hibernia delta. The clinoforms of the Hibernia delta and the equivalent parallel reflections seaward of the delta are similar to the sequences identified on the southeastern part of Grand Bank (Fader and Miller, 1986b).

(3) Vertically displaced reflections. Upturned parallel reflections on the downthrown (eastern) side of a near-vertical zone of disruption, are interpreted as the upward extension of growth faults at the basin margin.

Another type of vertical adjustment is evident at salt structures in the Jeanne d'Arc Basin. Parallel reflections in the Tertiary section were observed to be upturned against and draped or arched over deeper salt diapirs.

(4) Channel structure. At the western margin of the study area, infilled channel structures are cut into the Tertiary sequence of parallel reflections to about 150 m below the seabed. The channel structures comprise a complex of unorganized independent channels which are interpreted as glacial tunnel valleys that have been infilled or as the erosional remnants of a subaerial drainage system.

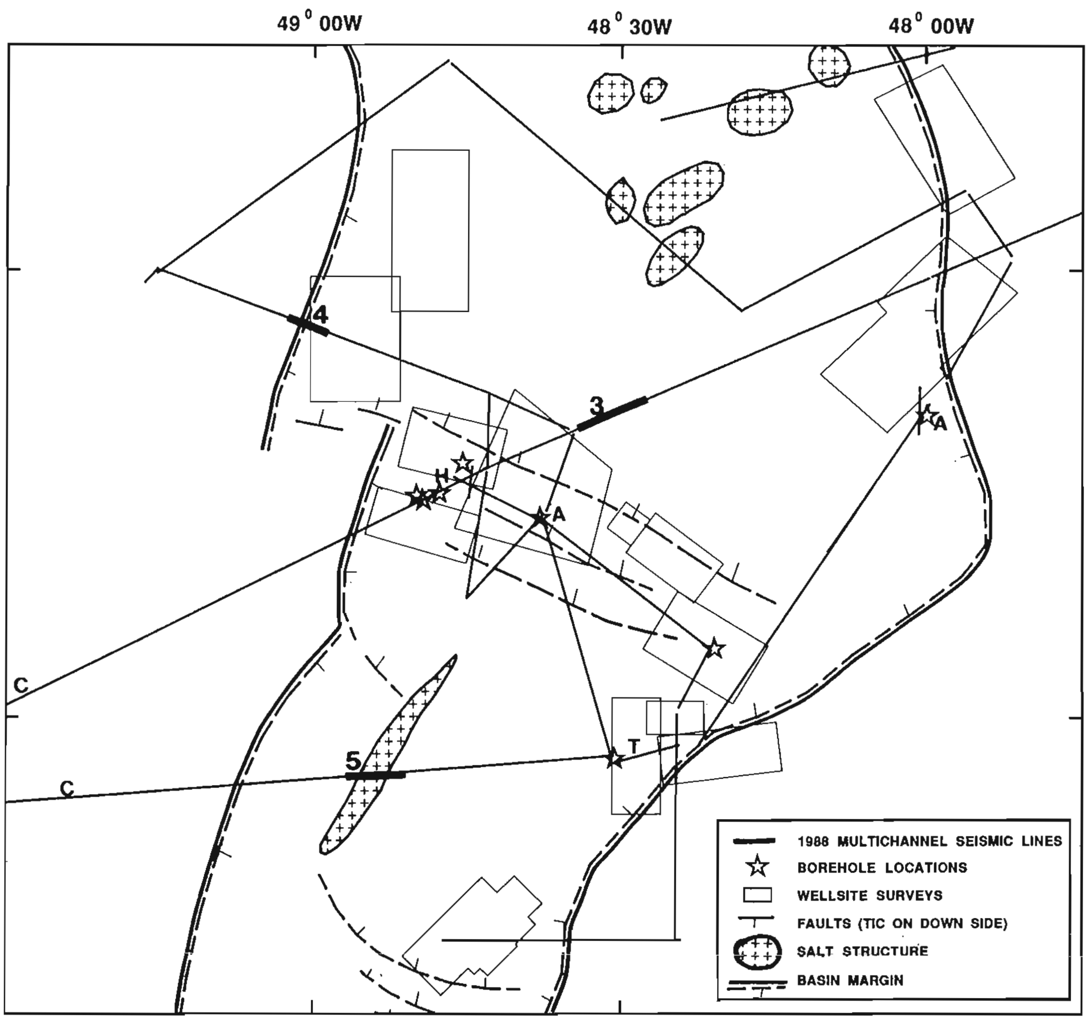

Figure 2. Map of northeastern Grand Banks area showing locations of industry well site surveys and 1988 GSC multichannel seismic lines superimposed on a structural map of part of the Jeanne d'Arc Basin (adapted from Jackson, 1985 and Grant et al., 1986). Letters indicate locations of sites or features discussed in the text. Numerals (3-5) and associated thickened line segments show locations of seismic sections discussed in the text and shown in Figures 3-5.

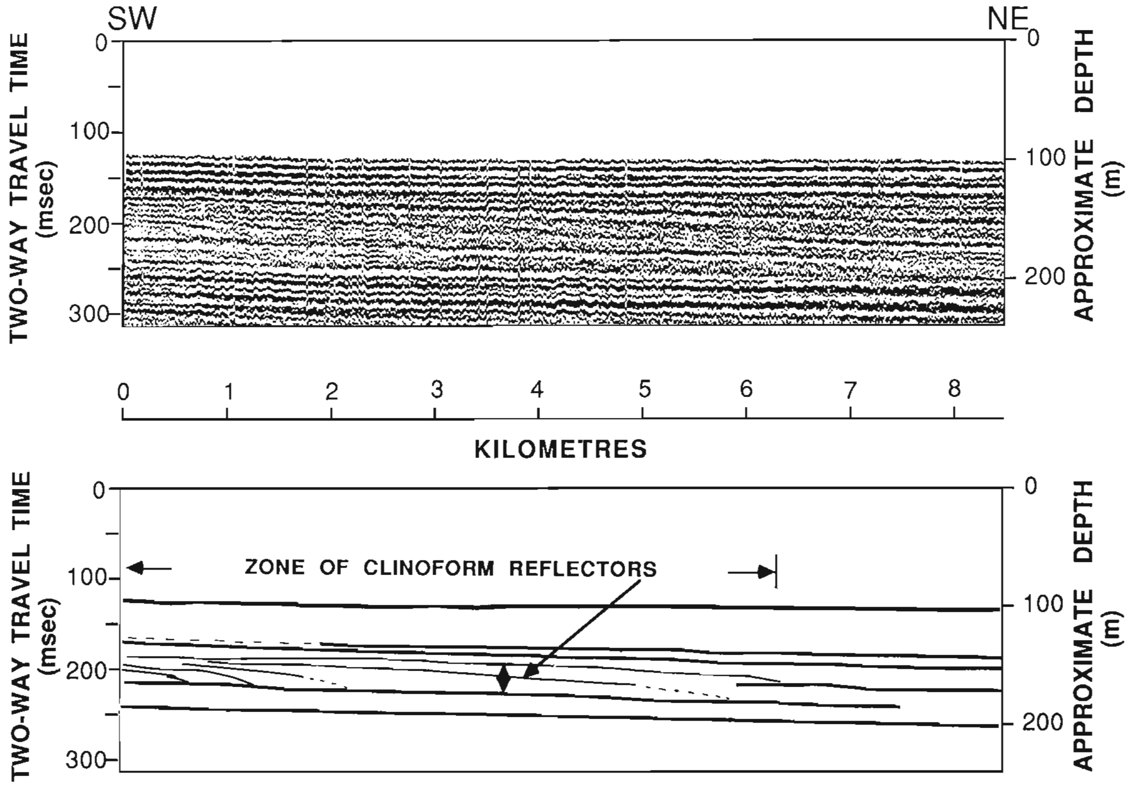

Figure 3. Near-trace record of a 1988 multichannel 16 kJ sparker profile showing a sequence of parallel reflections enclosing a zone of clinoform reflections representing the Hibernia delta seismostratigraphic unit. See Figure 2 for location. The approximate depth is estimated on the basis of a velocity of 1500 m/s.

Between 1979 and 1984, seven boreholes (Fig. 2) were completed by Mobil Oil Canada Limited and Petro-Canada Resources Limited, five in the Hibernia area and one each in the Ben Nevis and Terra Nova areas. In 1987 and 1988, the Atlantic Geoscience Centre completed boreholes at two locations (marked A on Fig. 2), and in 1988 Mobil Oil Canada Limited and Petro-Canada Resources Limited completed additional boreholes at their Terra Nova and Hibernia sites respectively (marked T and H on Fig. 2). Core descriptions, dates and physical property measurements from these boreholes will be used to provide geological data for the interpretation of the seismostratigraphy. Detailed investigations of the relationship between the acoustic and physical properties of the sediments will be performed by constructing synthetic sonograms in the area of the boreholes (Kearey and Brooks, 1984).

FIELD INVESTIGATIONS AND PRELIMINARY INTERPRETATION OF THE NEAR-SURFACE SEISMOSTRATIGRAPHY

In September 1988, about 1200 km of multichannel reflection seismic sections were obtained in the northeastern sector of the Grand Bank, using a multitipped 16 kJ sparker and a 400 m 24-channel streamer; a Huntec Deep Towed Seismic subbottom profiler (Hutchins et al., 1976); a 1 kJ sparker with a 7.6 m streamer; and sidescan sonars. Survey lines were chosen to tie in as many of the existing well site surveys and borehole locations as possible.

The 1988 data set supports the earlier seismostratigraphic analysis (Lewis et al., 1987; unpublished report, 1988) from northeastern Grand Bank. The seabed is a major unconformity and erosional surface as shown in many place by the truncation of underlying reflections. Major unconformities in the subsurface, however, are generally not evident, though irregularities between the parallel reflections seen in the near-surface Tertiary sediments occur.

The majority of seismic sections on the northeastern Grand Banks exhibit a succession of medium to high amplitude, continuous coherent parallel reflections which generally dip east-northeastward at less than 0.5 degrees (Fig. 3), and from borehole investigations (S.V. Barrie et al., C-Core unpublished report, Memorial University of Newfoundland, 1983) represent widespread units of bedded marine silt-clay sediments of Tertiary age. These sediments show a seaward increase in dip and sequence thickness. To accommodate this thickening, reflection events show offlap sequences typical of progradation. These gently dipping conformable sequences are interpreted to have been deposited in prodeltaic environments, based on the presence of delta structures in the area.

The seaward limit of progradational clinoform reflections is shown at about 6.3 km along the profile of Figure 3 within the general sequence of parallel reflections. These reflections were interpreted by Lewis et al. (1987; unpublished report, 1988) as a delta. This delta was tentatively attributed a Plio-Pleistocene age on the basis of

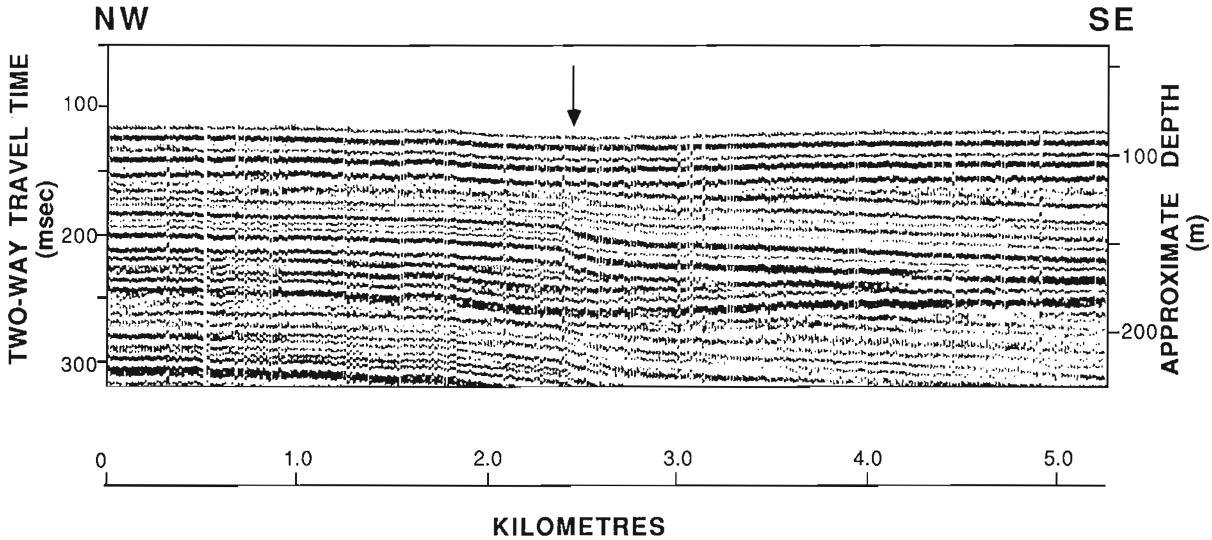

Figure 4. Near-trace record of a 1988 multichannel 16 kJ sparker profile. Arrow indicates vertical adjustment in a sequence of parallel reflections just below the seabed at the western margin of the Jeanne d'Arc Basin. See Figure 2 for location. The approximate depth scale is estimated on the basis of a velocity of 1500 m/s.

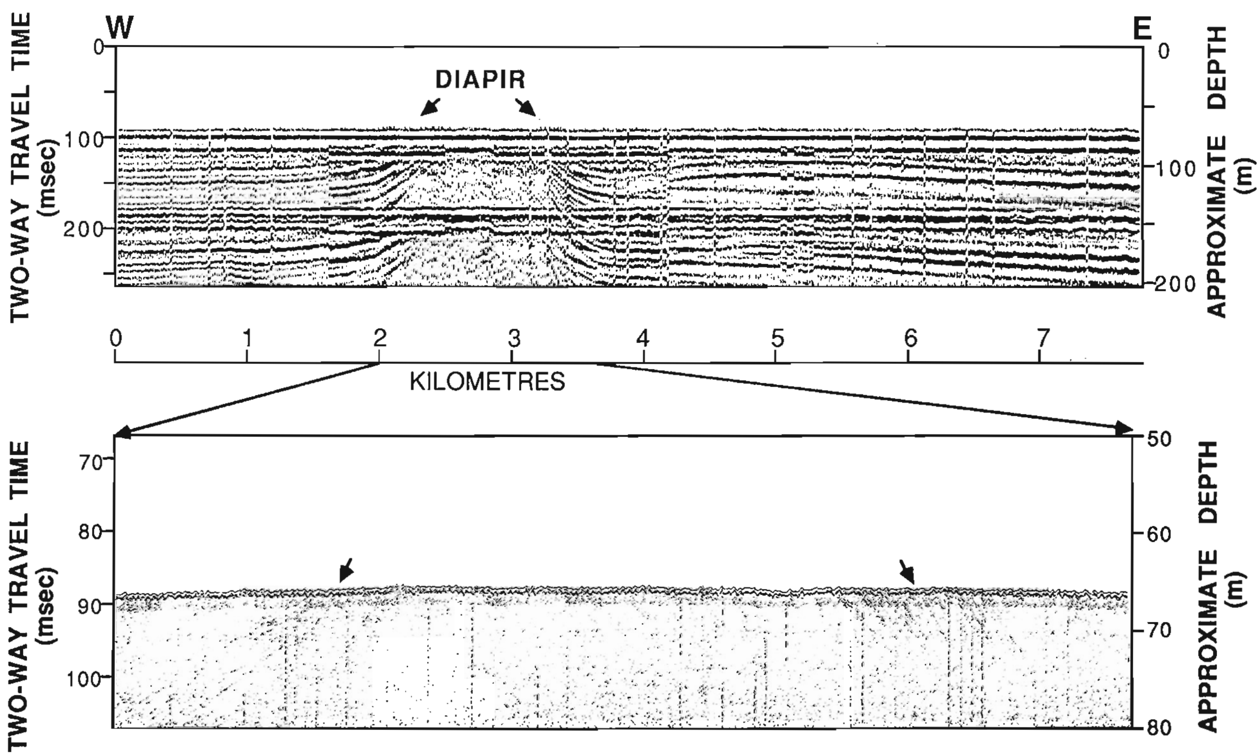

Figure 5. Seismic reflection profiles showing a salt diapir and upturned Tertiary beds of the Jeanne d'Arc Basin truncated at the seabed. Upper panel is a near-trace record of a 1988 multichannel 16 kJ sparker reflection section. Lower panel illustrates a high resolution boomer section of the seabed across the same truncated diapir. Arrows above each section point to upturned parallel reflections against the acoustically amorphous diapiric salt. Note the occurrence of a low amplitude deflection in a parallel reflection near the diapir between 4 and 7 km. See Figure 2 for location. The approximate depth scale is estimated on the basis of a velocity of 1500 m/s.

limited foraminifera and palynomorph data from the 1979-1984 boreholes. The delta, which underlies the proposed Hibernia production site, is dominantly composed of sand, and overlies fine grained, mid to late Tertiary sediments. The Hibernia delta is truncated at the present seabed and may be a significant source of sand for the formation and maintenance of bedforms on the present ocean floor. The existence of a Plio-Pleistocene delta at Hibernia is in broad agreement also with inferences of late Cenozoic sea levels based on global seismostratigraphy (Haq et al., 1987) and the stratigraphy of continental slope sediments on the eastern Canadian continental margin (Piper and Sparkes, 1986; Piper et al., 1987).

At the faulted western margin of Jeanne d'Arc Basin, parallel reflections are discontinuous or show abrupt changes in dip, as shown, for example, in Figure 4. Though the near-seabed reflections are bent slightly, the seabed reflection itself is apparently not deformed, suggesting stability at present. Other records, illustrated in Lewis et al. (1987; unpublished report, 1988), show upturned parallel reflections on the downthrown (eastern) side of a near-vertical zone of disruption.

Seismic reflection profiles over diapirs which reach the seabed (Fig. 5) display ruptured and upturned parallel reflections on the flank of the acoustically amorphous diapir material. Structural disturbances such as growth faults, flexures, salt diapirs and anomalous reflector bulging (shown in Lewis et al., 1987; unpublished report, 1988) over trans-basin faults are present in the shallow subseabed section and appear to be related to deeper seated phenomena. The data set will be analyzed to determine a potential relationship between deep and shallow features and the roles such structures may play in the development of seabed morphology and the timing and extent of tectonic activity identified on the basis of flexures over deep-seated faults.

Channels are generally rare. However, one major complex of infilled channels with independent thalwegs exists on northeastern Grand Bank, and is interpreted to have originated as glacial tunnel valleys beneath a Quaternary ice cap(s) or as earlier preglacial fluvial remnants infilled with Quaternary sediments. Additional evidence of channel structures in the same area were observed in 1988 and their location is shown by the letter "C" on Figure 2.

SUMMARY

Knowledge of regional geological conditions aids the understanding of variations in physical properties, and increases confidence in their determination. Our present knowledge of the upper few hundred metres of the Tertiary section indicates geological features that may influence sediment strength. The distribution of these features could form criteria for a zonation of the potential variability of physical properties affecting stability in the foundation zone of northeastern Grand Bank.

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Petrology, structural geology, and gold mineralization of the Elmtree mafic body, northern New Brunswick[1]

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Abstract

Significant gold mineralization is present in a small lenticular mafic body in the Ordovician Elmtree Group in northeastern New Brunswick. Margins are fine grained ophitic gabbros that grade into medium grained subophitic gabbros and coarse grained anorthositic cumulates toward the centre of the body. The mafic body was deformed and metamorphosed under greenschist facies conditions during the Taconic Orogeny. East-west-trending dominant shear fabric cuts across an earlier fabric. Shearing is brittle in the interior coarse grained cumulate portion of the body. Fine grained ophitic marginal zones display anastomosing foliation. A weakly penetrative, closely spaced cleavage intersecting the shear fabric at shallow angles indicates a component of dextral movement within the shear zone. The intrusion has been strongly altered during the regional greenschist metamorphism and during the hydrothermal alteration associated with the shear zone. Gold mineralization appears to be intimately associated with this secondary hydrothermal alteration. Timing of the shearing is believed to be Devonian, corresponding to the Acadian Orogeny.

Résumé

On rencontre dans un corps mafique lenticulaire de petite taille, situé dans le groupe ordovicien d'Elmtree de la partie nord-est du Nouveau-Brunswick, une minéralisation importante en or. Les marges se composent de gabbros ophitiques à grain fin, qui passent progressivement à des gabbros subophitiques à grain moyen et à des cumulats anorthositiques à grain grossier à mesure que l'on se rapproche du centre de ce corps mafique. Ce dernier a été déformé et métamorphisé dans les conditions du faciès des schistes verts durant l'orogénèse taconique. Une fabrique dominante, créée par cisaillement et de direction est-ouest, recoupe une fabrique plus ancienne. Le cisaillement est fragile dans la portion intérieure du corps mafique composée d'un cumulat à grain grossier. Des zones marginales ophitiques à grain fin montrent une foliation de type anastomosé. Une schistosité faiblement pénétrante et très serrée qui recoupe la fabrique de cisaillement à des angles très faibles, indique l'existence d'une composante de mouvement dextre à l'intérieur de la zone de cisaillement. L'intrusion a été fortement altérée durant le métamorphisme régional au degré du faciès des schistes verts et durant l'altération hydrothermale associée à la zone de cisaillement. La minéralisation en or semble être intimement associée à cette altération hydrothermale secondaire. On estime que le cisaillement a eu lieu au Dénovien, c'est-à-dire durant l'orogénèse acadienne.

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INTRODUCTION

Detailed mapping of the gold-bearing Elmtree mafic body was carried out during the summer of 1987 as part of a general suvey of the mafic and ultramafic rocks in New Brunswick. The petrology of the mafic body and the internal structure of the shear zone have been studied in detail in order to understand relationships amongst the lithologies, structures and the gold mineralization.

Gold mineralization was first discovered in 1984 by the Lacana Corporation (now Corona Resources). Following the discovery, detailed geophysical and geochemical work and diamond drilling were carried out by the company.

GEOLOGICAL SETTING

The study area is located about 20 km northwest of Bathurst near the Alcida settlement. The area is underlain by the Elmtree Group and the Chaleur Bay Group sedimentary rocks (Fig. 1). The Elmtree Group consists dominantly of sandstone and siltstone, with minor limestone, conglomerate and mafic volcanic rocks. These rocks were strongly deformed during the Taconic Orogeny. The Silurian Chaleur Bay Group is composed of siltstone, sandstone, conglomerate and limestone, and their metamorphosed equivalents (Davies et al., 1969). The Chaleur Bay Group occurs in a broad northeasterly-trending synclinorium (Fig. 1) and unconformably overlies the Elmtree Group (Ruitenberg et al., 1977). The Rocky Brook-Millstream fault zone defines the southern contact of the Chaleur Bay Group with the Tetagouche Group of rocks.

The Elmtree and Tetagouche groups are intruded by mafic to felsic rocks ranging in age from Ordovician to Devonian. Numerous sill-like bodies of mafic rocks in the Tetagouche Group appear to have intruded prior to the

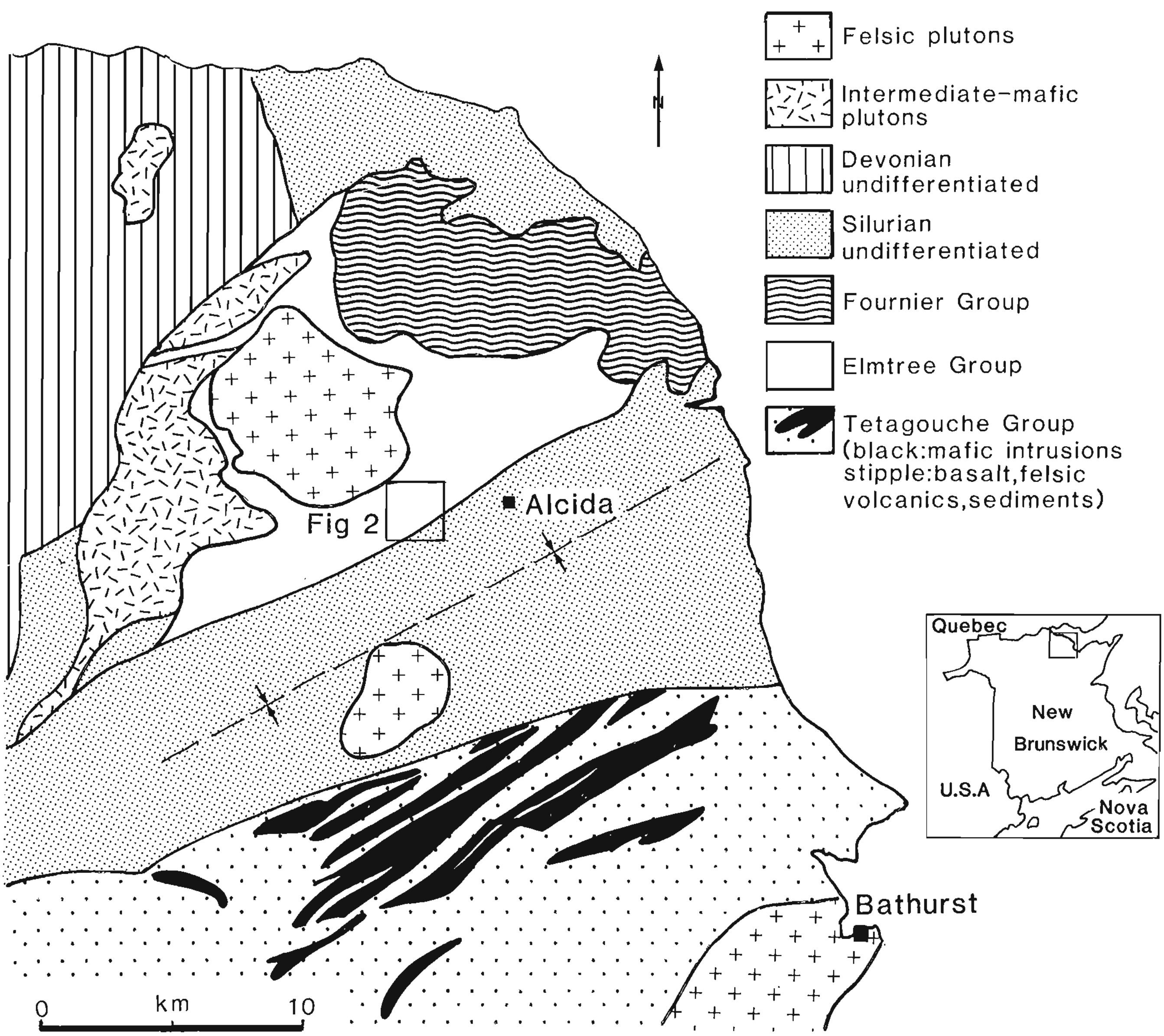

Figure 1. General geology of the area (after Davies, 1979 with some simplifications).

Taconic Orogeny (Davies et al., 1969). Rocks intruding the Chaleur Bay Group are mostly diorite and diabasic gabbro, and are generally less mafic in character than those occurring in the Tetagouche Group. Ultramafic rocks occur along the Rocky Brook-Millstream fault zone for a distance of approximately 3 km. Two other lensoid-shaped ultramafic bodies occur to the north of the Elmtree mafic body, which is situated within Elmtree Group rocks, just north of the Elmtree - Chaleur Bay boundary.(Fig. 2). There are two granitic stocks of Devonian age in the area (Fig. 1). One is located to the north of the Elmtree mafic body. The other one occurs in the Chaleur Bay Group rocks to the south.

PETROLOGY

The Elmtree mafic body can be traced about 600 m along strike (Fig. 2). The width of the body reaches a maximum of 40 m. It dips vertically or steeply to the north (McCutcheon et al., 1988), and appears to be conformable with the dominant structural trend of the surrounding rocks. The regional strike of the foliation is approximately northeast (060°).

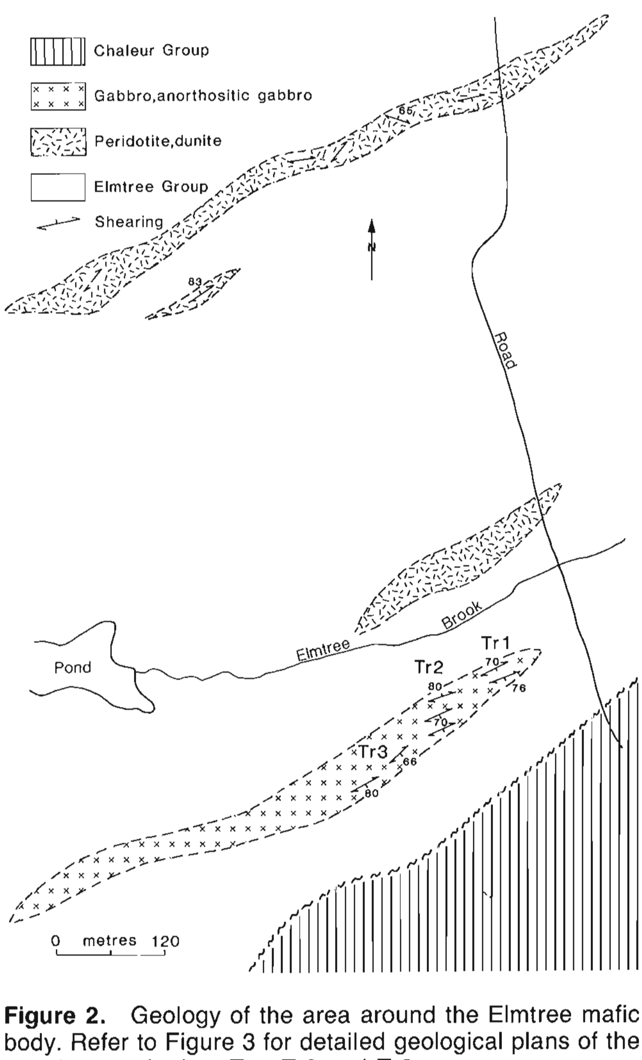

Figure 2. Geology of the area around the Elmtree mafic body. Refer to Figure 3 for detailed geological plans of the trenches marked as Tr1, Tr2 and Tr3.

The lens-shaped Elmtree mafic body is zoned. Fine grained ophitic gabbros occur at the margins (Fig. 3). These marginal rocks grade into medium grained subophitic gabbros and coarse grained anorthositic rocks toward the centre of the body. Ophitic to subophitic marginal rocks account for 60 percent of the total width of the body. Pseudomorphs of plagioclase laths ranging in size from 0.3 to 1 mm occur in a groundmass of clinopyroxene in these rocks. The interior medium to coarse grained portion of the body is partly subophitic and partly cumulus in nature. Cumulus habit is especially apparent in the anorthositic portions. Other rocks that are texturally transitional from ophitic rocks into cumulates would properly be called accumulates following Irvine's (1982) terminology. The grain size of cumulus plagioclase pseudomorphs reaches 7 mm in length. Most of the large plagioclase laths display deformational textures. Kinked, bent and broken crystals of plagioclase are common (Fig. 4). Largely pseudomorphic clinopyroxene appears to be interstitial in nature. Clinopyroxene partly becomes a cumulus phase where it forms large subhedral and euhedral grains (Fig. 5). Some of these coarse grains are strained. Apatite is an accessory phase and appears to be more abundant in the anorthositic rocks. Its grain size is about 60-80 mm. Primary ilmenite in the form of disseminated subhedral to anhedral grains is especially abundant in the ophitic gabbros. Grain sizes range up to 400 mm.

Secondary minerals consisting of albite, epidote (clinozoisite) and chlorite (clinochlore) are widespread in the intrusion. Minor hornblende and biotite are present in some rocks. Talc, quartz and carbonate locally become abundant. A pervasive albite-epidote-chlorite assemblage is indicative of greenschist facies conditions. Quartz-carbonate alteration in the form of narrow veins becomes abundant only in the sheared coarse grained anorthositic portion of the body and clearly postdates the albite+epidote +chlorite assemblage. Some of the strongly strained quartz occurring as an interstitial phase among the plagioclase laths appears to be earlier than the development of the quartz-carbonate assemblage.

Although the textures are preserved, alteration has obliterated most of the original igneous minerals with the exception of local relics of augite. Plagioclase has been completely altered to albite and clinozoisite. Alteration of plagioclase appears to have proceeded faster than the alteration of clinopyroxene. Augite relics display a narrow range in composition from $En_{45}Fs_{11}Wo_{45}$ to $En_{35}Fs_{22}Wo_{43}$ (Fig. 6).

GEOCHEMISTRY

Whole rock chemical analyses of the intrusion rocks collected for the purpose of defining the original chemistry are given in Table 1. These analyses represent unweathered and fresh rocks of the intrusion. Rocks collected for whole rock geochemical purposes were analyzed at X-Ray Assay Laboratories Limited, Toronto. Major elements were analyzed by X-Ray fluorescence using fused disks and by wet chemical techniques. Trace elements were analyzed by ICP-MS and XRF using both fused disks and pressed pellets.

Based on duplicates and analyses of standard MRG-1 (gabbro, GSC) and W-2 (diabase, USGS) (Abbey, 1983), the precision and accuracy of the analyses are estimated as follows (relative %): SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, Na_2O, K_2O and P_2O_5: ±1 %; FeO, MnO, Rb: ±2 %; CO_2: ±3 %; H_2O, S: ±5 %; Ba, Co, Cr, Cu, Ni, Pb, Zn, Nb, Sr, Y, Zr: ±10 %.

Except for several samples, most of the elements do not display erratic variations amongst the rocks belonging to a specific rock type. In general, anorthositic gabbro contains lower concentrations of CaO, MgO, Cr, Ni and Co and higher concentrations of Na_2O, P_2O_5, Zr and Y than the gabbro. The Al_2O_3 contents are uniform. SiO_2 is uniform in gabbro whereas it is variable within 3 weight percent in anorthositic gabbro. The CO_2 contents are variable. Gabbro contains higher concentrations of CO_2 than the anorthositic gabbro. Anorthositic gabbro samples with high CO_2 are considerably richer in CaO as compared to those with low CO_2. This indicates strong mobility of Ca in CO_2-rich fluids. The H_2O abundances are uniform around 3 weight percent. These high H_2O and CO_2 concentrations are too high for primary compositions. In addition to this, taking the pervasive albitization and chloritization into account, it would be unrealistic not to consider the mobilization of major elements. Studies on metabasalts containing albite, epidote and chlorite as alteration products have reported substantial Ca losses and Mg gains (Cann, 1969; Miyashiro et al., 1971; Spooner and Fyfe, 1973; Humphris and Thompson, 1978; Gelinas et al., 1982). Al released during the albitization of plagioclase can be fixed in chlorite. Epidote might have accommodated some of the Ca released from plagioclase and clinopyroxene during albitization and chloritization. Such an alteration also requires oxidation of FeO. The rocks with higher Fe_2O_3/FeO ratios contain abundant epidote. This ratio is relatively uniform in gabbro but highly variable in anorthositic gabbro. Original concentrations of TiO_2, P_2O_5, Cr, Ni, Co, Zr and Y appear to have been unchanged. A positive correlation exists between P_2O_5 and Zr. Anorthositic gabbro is richer in Zr and P_2O_5 than the gabbro.

The detailed sampling and analysis of rocks along a diamond-drill intersection of the body demonstrated some interesting patterns (McCutcheon et al., 1988). In general, some elements appear to be unaffected by metamorphism and hydrothermal alteration; therefore, the well-defined trends are very likely to have resulted from igneous processes. MgO, CaO, V, Co and Ni display depletions inward from the fine grained gabbro into the coarse grained anorthositic gabbro zone. Na_2O and P_2O_5 display antipathetic variations to those of above. These variations are very similar to the variations described above and are compatible with the mineralogical and petrographical observations. Decrease in the MgO, CaO, V, Co and Ni can be easily explained by the low modal abundance of clinopyroxene in the anorthositic rocks.

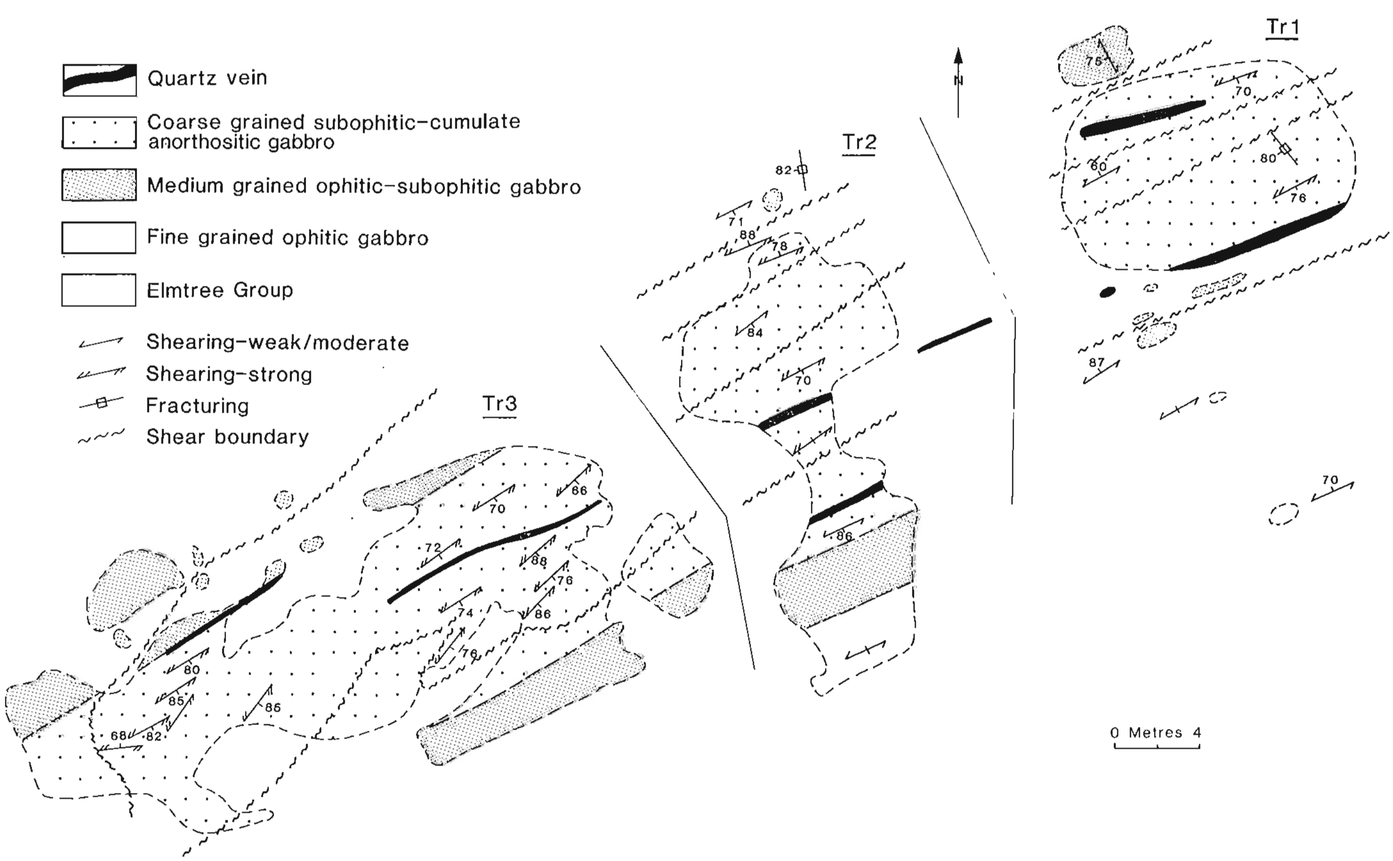

Figure 3. Detailed geological plans of the trenches. See Figure 2 for the location of the trenches.

The chemistry of the ophitic portions of the body is very similar to the chemistry of the Tetagouche basalts and mafic subvolcanic intrusives (A.D. Paktunc, unpublished report, 1988) and in general indicates an ocean floor setting (Figs. 7 and 8).

STRUCTURAL GEOLOGY

Steeply dipping shear fabric within the Elmtree mafic intrusion strikes approximately 055°, parallel to the general trend of the body. Locally, it encloses and anastomoses around infrequent metre-scale lenses of rock lacking strong deformation features (Fig. 9). The shear fabric appears to have preferentially developed within the central coarse grained portion of the mafic body. A few distinct zones of intense, closely spaced shear fabric up to one metre in width are evident in this portion. These zones are characterized by pervasive rusty weathering and by the presence of 5-8 percent sulphide mineralization. A number of narrow discontinuous quartz veins trend subparallel to the foliation.

Outside the coarse grained portion of the body where the dominant shear foliation is weakly to moderately developed, a weakly penetrative, closely spaced cleavage striking 020-030° and dipping steeply to the northwest is preserved (Fig. 9). This foliation, which intersects the shear fabric at shallow angles, has been deformed by shearing, and represents either an earlier formed cleavage or an S-fabric to the shear foliation. Insufficient field evidence hinders a conclusion as to which possibility is more likely. In either case, the orientation of the cleavage relative to the shear fabric and its clockwise rotation into this fabric indicates a component of dextral movement within the shear zone.

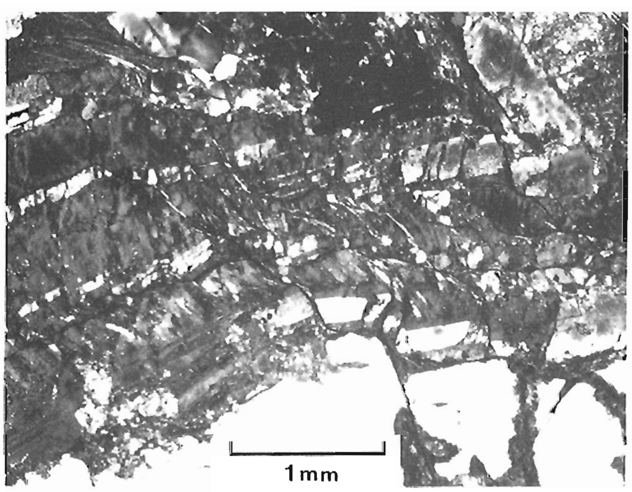

Figure 4. Photomicrograph showing deformed plagioclase laths. Transmitted light; crossed polarizers.

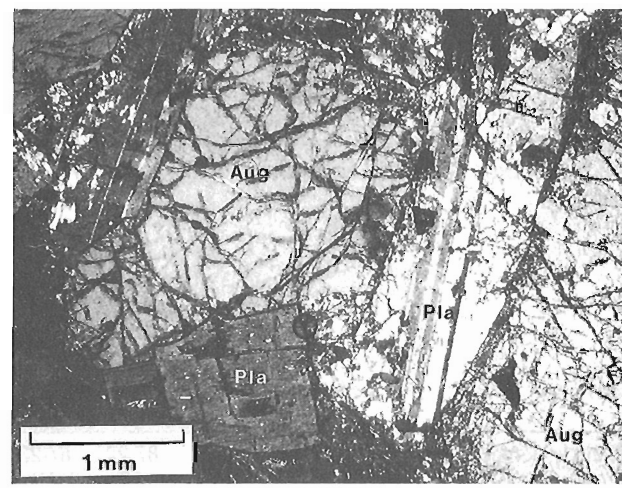

Figure 5. Photomicrograph showing subhedral nature of augite grains (Aug). Pla: plagioclase. Transmitted light; crossed polarizers.

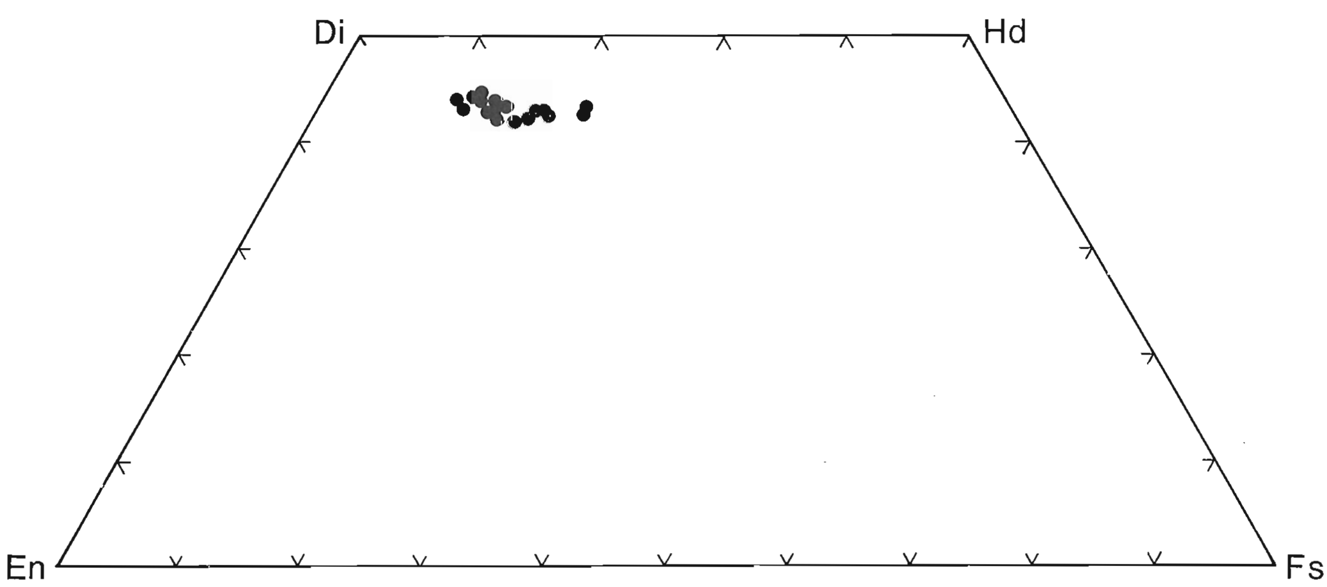

Figure 6. Clinopyroxene compositions in the pyroxene quadrilateral. Analyses by Cameca-CAMEBAX (WDS) and MAC (EDS) electron microprobes at the Geological Survey of Canada. Analytical conditions: 15 kV, 30 nA and 10 sec for the Cameca probe and 20 kV, 10 nA and 50 sec for the MAC probe.

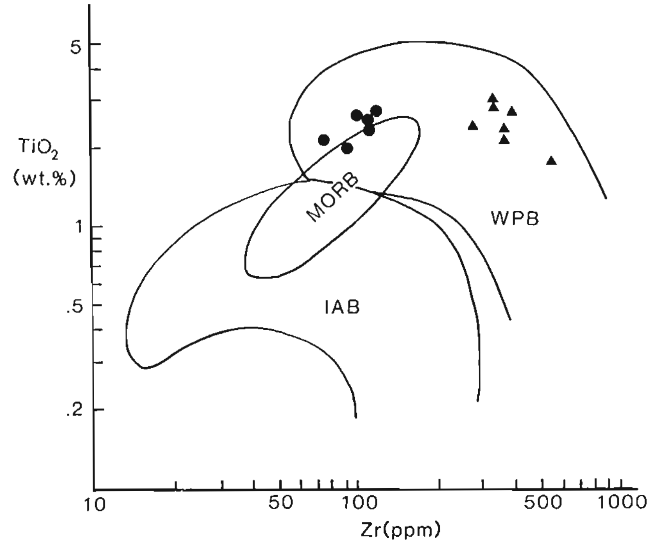

Figure 7. Discriminant diagram of TiO_2 versus Zr showing the distribution of the Elmtree samples with regard to the fields of mid-ocean ridge basalts (MORB), within plate basalts (WPB) and island arc basalts (IAB) (after Pearce 1980). Gabbros shown by dots; anorthositic gabbro by triangles.

A steeply dipping fracture cleavage with spacings on the centimetre scale overprints the shear fabric and is present in both the coarse grained central and the fine grained marginal rocks. This cleavage strikes at high angles to the shear foliation and is best developed within the cumulate zone. Outside of this zone, the cleavage becomes less penetrative and more closely spaced. The exact relationship of this fabric to the other structural elements is not known, but it likely formed in response to a relatively late deformational event.

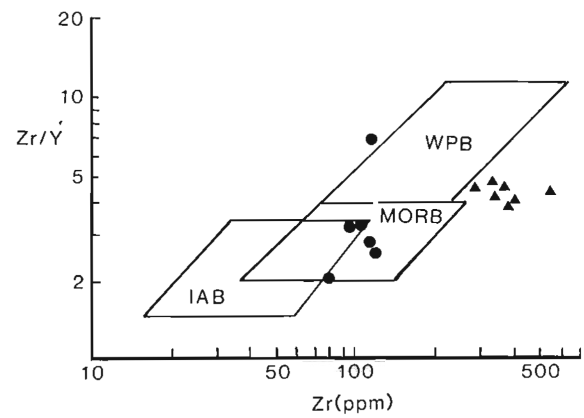

Figure 8. Discriminant diagram of Zr/Y ratio versus Zr (after Pearce, 1980). Refer to Figure 7 for the abbreviations and symbols.

Table 1. Whole rock chemical analyses

Sample number	87-14	87-16	87-17	87-22	87-23	87-24	87-25	87-26	87-32	87-33	87-19	87-20	87-21
rock	AG	AG	AG	AG	AG	AG	AG	G	G	G	G	G	G
SiO_2	58.00	55.10	51.40	53.50	54.00	48.00	50.40	48.70	46.20	48.00	46.90	48.40	47.80
TiO_2	1.71	2.66	2.11	2.37	3.01	2.26	2.87	2.15	2.57	2.73	2.01	2.37	2.53
Al_2O_3	13.60	13.40	12.90	14.00	13.50	13.30	12.30	14.50	12.40	13.20	13.50	13.70	13.30
Fe_2O_3	7.03	5.02	11.23	2.33	5.79	12.94	3.40	1.82	1.65	3.05	2.07	3.30	3.90
FeO	5.10	6.90	2.40	9.50	9.90	2.30	12.50	10.50	11.10	11.10	10.10	9.80	10.70
MnO	0.23	0.35	0.21	0.28	0.37	0.20	0.35	0.24	0.22	0.25	0.22	0.25	0.26
MgO	1.52	2.29	2.48	2.04	2.71	2.49	2.41	6.74	5.38	5.72	6.07	5.68	5.78
CaO	0.92	0.88	1.52	4.64	1.03	1.16	4.90	3.55	6.92	7.81	7.57	6.01	6.87
Na_2O	5.43	5.27	3.93	4.86	3.91	5.13	3.75	3.07	2.85	3.81	2.79	3.95	3.87
K_2O	0.41	0.24	1.35	0.40	0.35	0.57	0.24	0.39	0.13	0.12	0.28	0.22	0.20
P_2O_5	0.66	0.65	0.67	0.32	0.52	0.70	0.64	0.17	0.18	0.21	0.19	0.23	0.22
H_2O	2.90	2.90	3.50	2.50	3.70	3.30	3.10	4.60	4.20	3.20	4.20	3.40	3.20
CO_2	0.02	0.03	0.02	2.11	0.06	0.05	3.15	2.28	5.38	0.17	4.74	0.40	0.26
S	0.40	3.02	1.60	0.01	0.11	1.40	0.02	0.05	0.09	0.09	0.03	0.01	0.01
Total	97.93	98.71	95.32	98.86	98.96	93.8	100.0	98.76	99.27	99.46	100.6	97.72	98.9
ppm													
Ba	247	180	414	667	268	362	196	348	120	101	152	267	236
Co	2	2	3	2	11	12	6	27	17	14	16	18	21
Cr	94	89	104	15	20	26	11	113	24	24	51	58	26
Cu	17	14	35	13	10	37	13	100	78	73	62	69	72
Ni	7	10	6	7	4	6	2	56	27	20	45	27	30
Pb	14	15	11	25	14	21	9	9	10	8	11	12	5
Zn	68	75	165	120	118	118	139	99	97	102	83	116	113
Nb	20	37	37	26	37	39	24	10	18	14	15	23	15
Rb	21	12	49	30	24	30	27	28	25	10	29	38	17
Sr	179	168	83	327	191	154	224	273	194	171	330	298	303
Y	120	102	82	61	79	96	69	38	32	46	27	39	16
Zr	522	392	363	271	322	366	325	74	102	116	93	109	109

AG: anortositic gabbro; G: gabbro

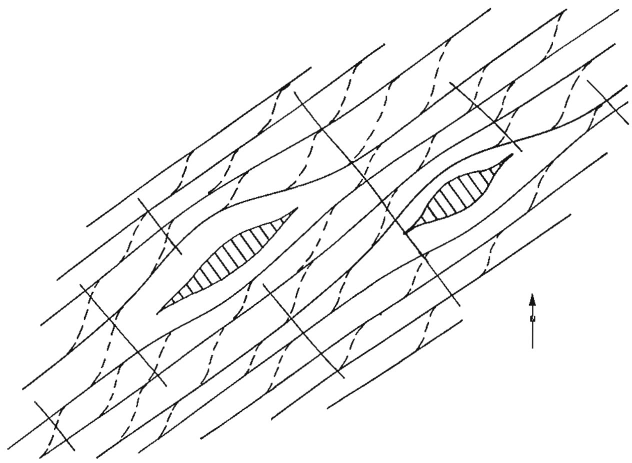

Figure 9. Sketch illustrating the relationship of structural fabrics present in the Elmtree mafic body. A strong shear fabric (055°) anastomoses around infrequent, relatively undeformed lenses of rock (hatched). A fabric (dashed lines) striking 020-030° represents either a deformed earlier cleavage or an S-fabric to the shear foliation. Both imply dextral shear motion. A spaced fracture cleavage striking 320° overprints the shear fabric.

DISCUSSION AND IMPLICATIONS FOR THE GOLD MINERALIZATION

The chemistry of the fine grained ophitic marginal rocks is similar to the Bathurst mafic intrusives (A.D. Paktunc, unpublished report, 1988) and basalts (van Staal, 1987), and in general suggests an ocean floor setting. Zoning of the Elmtree mafic body may have resulted from flow differentiation. High shear strength in a flowing medium near the margins (Komar, 1972) causes the phenocrysts to move away toward the centre. Accumulation of the plagioclase phenocrysts in the centre of the body, or zoning in general might have resulted from such a process.

The age of the body appears to be Ordovician and the Elmtree fault postdates the emplacement of the mafic body. This is in contrast with the views of McCutcheon et al. (1988) who believe that the age of the body is Devonian and that the body was emplaced along the Elmtree fault zone. The Elmtree mafic body is lithologically and chemically similar to Ordovician intrusive rocks in the Bathurst Camp. It appears that the timing of the shearing is post-Silurian, perhaps related to an Acadian orogenic event. The dextral sense of movement within the shear zone is similar to the sense of movement noted in deformed Silurian Chaleur Group sediments lying to the south of the study area (C.R. van Staal, pers. comm., 1987). Presence of an early structural fabric (S_1), predating the shearing suggests an Ordovician age. The quartz+carbonate assemblage postdating the pervasively developed chlorite+epidote+albite assemblage appears to be penecontemporaneous with the shearing. This places some constraints on timing of the hydrothermal alteration with which the gold mineralization is associated.

Gold mineralization is associated with sulphide-rich zones localized around zones of intense shear fabric and especially in the near vicinity of quartz veins. Along the shear planes, characterized by pervasive rusty weathering, abundant disseminated sulphides (up to 8 percent) consisting of euhedral arsenopyrite, pyrite and pyrrhotite are present. Arsenopyrite is unique to these structures as it is scarce to absent within the surrounding rocks. These narrow intensely sheared and altered zones are reported to carry the highest concentrations of gold mineralization within the mafic body. A ten metre thick altered anorthositic gabbro zone is reported to carry 6 ppm Au on average (D. Hoy, pers. comm., 1988). Hydrothermal fluids that carried the gold may be associated with the Middle Devonian felsic magmatism, specifically the emplacement of the granitic body to the north. Heat required for mobilization of the hydrothermal fluids might have been supplied by the emplacement of the granitic body.

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Fluid inclusion studies and the origin of cupriferous calc-silicate hornfelses at Patapedia, Gaspésie, Quebec[1]

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Abstract

The factors responsible for the formation of cupriferous calc-silicate hornfelses in the Patapedia thermal zone are re-evaluated after study of fluid inclusions in quartz phenocrysts of synchronous felsic dykes.

Three distinct fluid inclusion types are recognized: a low to moderate salinity, high density aqueous fluid (Type I); a low density CO_2-rich fluid (Type II); and a high salinity, high density aqueous fluid (Type III). Types I and II predominate, whereas Type III inclusions form <10% of the population. All three types are interpreted to have been present during prograde metamorphism. Estimated temperatures and pressures of metamorphism are 450° to 500°C and 700-1000 bars, respectively.

A model is proposed in which the metamorphism was caused by heat transferred from a low to moderate salinity fluid of partly orthomagmatic origin. On cooling and migrating to higher levels this fluid evolved, first by dissolving H_2O/CO_2 produced by metamorphic reaction and later, by exsolving a high salinity aqueous fluid. The bulk of the metamorphism occurred at pressure-temperature conditions at which aqueous and CO_2-rich fluids were largely immiscible. Copper mineralization accompanied retrograde metamorphism at temperatures below 400°C and resulted from falling temperature and/or increased pH from CO_2 effervescence.

Résumé

Les facteurs responsables de la formation des cornéennes calco-silicatées cuprifères de la zone thermométamorphique de la Patapédia ont été réévalués suite à une étude des inclusions fluides contenues dans des phénocristaux quartzeux provenant de dykes felsiques synchrones.

On a identifié trois types distincts d'inclusions fluides: un fluide aqueux de salinité faible à modérée et de densité élevée (type I); un fluide de faible densité, riche en CO_2 (type II); et un fluide aqueux de salinité élevée et de forte densité (type III). On rencontre le plus souvent les types I et II, tandis que les inclusions de type III représentent < 10% de la population d'inclusions. Il semble que les trois types auraient existé durant le métamorphisme prograde. On estime les températures auxquelles le métamorphisme s'est produit dans l'intervalle de 450° à 500°C, et dans le cas des pressions, de 700 à 1 000 bars.

On propose un modèle dans lequel le métamorphisme résulte de l'échange de chaleur provenant d'un fluide de salinité faible à moyenne, d'origine partiellement orthomagmatique. En se refroidissant et en migrant vers des niveaux plus élevés, ce fluide a évolué, tout d'abord en dissolvant du H_2O/CO_2 produits par les réactions métamorphiques, et ensuite en exsolvant un fluide aqueux de salinité élevée. La majeure partie du métamorphisme a eu lieu dans des conditions de pression et température dans lesquelles les fluides aqueux et riches en CO_2 étaient largement immiscibles. Une minéralisation en cuivre a accompagné le métamorphisme rétrograde aux températures inférieures à 400°C, et résultait d'une chute de température ou d'un accroissement du pH produit par effervescence du CO_2, ou les deux.

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INTRODUCTION

The Patapedia thermal zone, Gaspésie, Quebec, is an unusual grouping of calc-silicate metamorphic bodies which have outwardly telescoping isograds and display extensive retrograde alteration with associated copper sulphide mineralization (Fig. 1). These features and the concentration of altered dyke rocks in the metamorphic bodies led Williams-Jones (1982) to postulate that the metamorphism of the original calcareous sedimentary rocks was caused by heat transferred from an intrusive-related hydrothermal system. Such systems are typically quite saline (e.g. Roedder, 1971; Wilson et al., 1980; Reynolds and Beane, 1985), and, by interacting with carbonate rocks, may produce immiscible H_2O-CO_2 fluids. This possibility, which was not considered by Williams-Jones (1982) may have had profound implications for both the metamorphism (cf. Bowers and Helgeson, 1983a; Skippen and Trommsdorff, 1985) and the copper mineralization (cf. Drummond and Ohmoto, 1985).

Secondary fluid inclusions in quartz phenocrysts in the dyke rocks provide an excellent opportunity to establish the nature of the hydrothermal fluid and the fluid resulting from metamorphism. In particular, they provide a means of examining the possible role of fluid immiscibility in the development of the thermal zone and in concentrating the copper. This paper presents new estimates of pressure and temperature and re-evaluates the metamorphic and copper mineralization history of the Patapedia thermal zone in the light of results obtained from a detailed study of quartz phenocryst-hosted fluid inclusions.

GEOLOGICAL SETTING

The Patapedia thermal metamorphic zone occurs in argillaceous limestones/limey shales and massive limestones of the Late Ordovician to Early Silurian Matapedia Group and, to a minor extent, in noncalcareous siltstones of the overlying Devonian Fortin Group which outcrop in the Aroostock-Matapedia anticlinorium and Gaspé-Connecticut synclinorium, respectively. These rocks were isoclinally folded due to compressional deformation during the Acadian Orogeny, and were regionally metamorphosed to subgreenschist facies (Duba and Williams-Jones, 1983). The thermally metamorphosed zone is located near the intersection of the Patapedia fault, parallel to the river shown in Figure 1 and the Matapedia fault (not shown), a northeast-striking structure that runs the entire length of the Gaspé peninsula. A swarm of northeast-trending felsic porphyritic dyke rocks of calc-alkaline affinity have intruded the area. There are also a number of breccia dykes in the zone. Although no

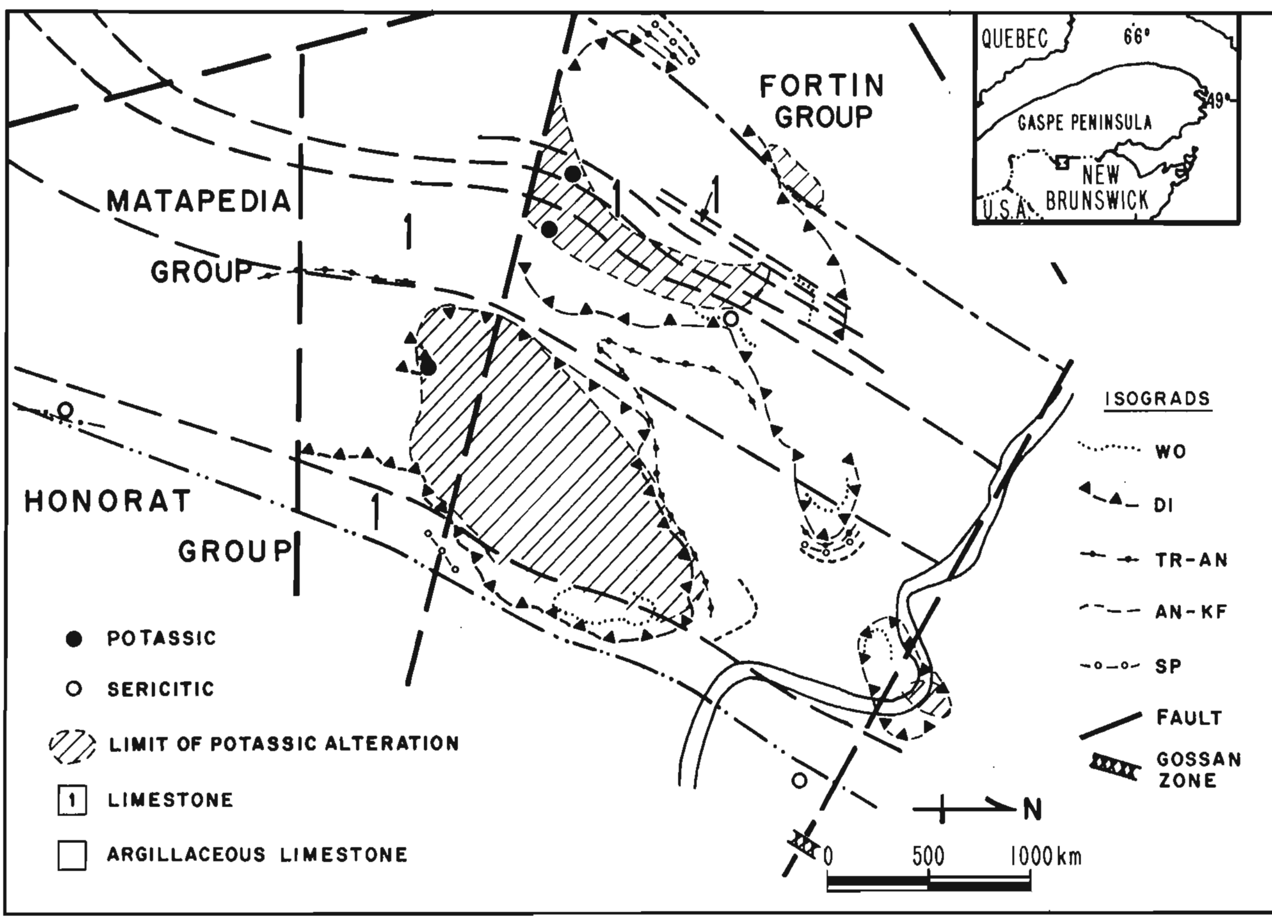

Figure 1. Map showing the location of the Patapedia thermal zone, the distribution of isograds and the locations of samples used in the study. AN = anorthite, DI = diopside, KF = potassium-feldspar, SP = titanite, TR = tremolite and WO = wollastonite. The copper mineralization is concentrated in the potassically altered zones.

larger igneous body is exposed, it is believed that such a body or bodies must exist at depth and that the dyke swarm represents its (or their) present surface extension.

The Patapedia metamorphic zone comprises a large central mass and several small satellites, the forms and distribution of which were apparently fault controlled (Fig. 1). Contacts with the unmetamorphosed sedimentary rocks are remarkably sharp and in consequence isograds are tightly telescoped along the margins of the metamorphic bodies. Most of these bodies are characterized by a central core of hornfelses and marbles containing the assemblage diopside $\pm$ tremolite $\pm$ calcite $\pm$ biotite and a transitional zone less than 10 percent of the width of the body in which up to 5 isograds can be mapped based on the following assemblages: anorthite-potassium-feldspar, titanite, tremolite-anorthite, diopside, and wollastonite. The most sensitive marker for the outer limit of thermal metamorphism is plagioclase composition which changes abruptly from $< AN_{10}$ in regionally metamorphosed sedimentary rocks to $> An_{50}$ in the hornfelses.

Retrograde metamorphism of the calc-silcate hornfelses and marbles was extensive and is evident as veins and coarse grained segregations of one or more of the following minerals: garnet, epidote, chlorite, calcite and quartz. Williams-Jones (1982) has shown that this event can be roughly subdivided into a higher temperature core zone characterized by the presence of garnet and a lower temperature outer zone containing epidote and tremolite-bearing assemblages. Sulphide mineralization, largely pyrrhotite and chalcopyrite, commonly forms part of the retrograde assemblage and, locally, is sufficiently concentrated, particularly in the garnet zone, to provide economic grades of copper.

Most of the dyke rocks within thermally metamorphosed zones show signs of hydrothermal alteration. Potassic alteration is prevalent within the central parts of the metamorphic zone and consists of potassium-feldspar flooding of the groundmass and rimming and/or veining of plagioclase phenocrysts. This alteration is overprinted by the assemblage sericite $\pm$ calcite $\pm$ pyrite which, in the outer parts of the metamorphic zone, completely obliterates evidence of earlier textures and assemblages. Propylitic alteration is found in some dyke rocks in the outer part of the metamorphic zone, but is more typical of dykes outside the mapped limit of metamorphism.

The features that have been described above, and evidence of magmatic signatures in the oxygen isotopic compositions of the hornfelses and marbles, led Williams-Jones (1982) to conclude that the Patapedia thermal metamorphic zone was formed as a result of heat transferred by an orthomagmatic-dominated hydrothermal system. The metamorphism of the argillaceous limestones produced a largely calcite-free rock which was fractured, altered and mineralized during later lower temperature stages of evolution of the hydrothermal system to produce a porphyry-style copper deposit.

The development of the thermal aureole and the related alteration of the dyke rocks represent the last hydrothermal activity in the Patapedia area.

FLUID INCLUSION STUDY

Introduction

The Patapedia thermal zone, as indicated above, is thought to have formed as a result of the transfer of heat from an intrusive-related hydrothermal system. It is believed that the dyke swarm in the zone is a manifestation of an unexposed intrusive body which was the source of the heat, and that the alteration of the dyke rocks records the evolution of the hydrothermal system. Quartz phenocrysts are common in most of these dyke rocks and, more importantly, contain abundant fluid inclusions. It is thus reasonable to expect that the inclusions would represent samples of fluid trapped at various stages in the life of the hydrothermal system and that their study could yield valuable information about the evolution of the system and the related pressure-temperature conditions of metamorphism. Some inclusions are present in diopside and tremolite in the calc silicate hornfelses, in garnet and diopside associated with mineralization and in calcite in the marbles. They are, however, too rare and small, to usefully study.

Fluid Inclusion Petrography

Six samples of quartz phenocryst-bearing dyke rocks were studied, of which three are potassically altered and the others phyllically altered. The locations of these samples are shown in Figure 1.

The majority of fluid inclusions lie along healed fractures which crisscross the quartz host and are thus considered to be secondary (Roedder, 1984). Many, however, are irregularly distributed in such a manner that their relationships to healed fractures or primary features of the phenocrysts are obscure. The origin of these inclusions is thus indeterminate. Most of the inclusions are less than 10 μm in diameter, although some range up to 30 μm in diameter, and have regular equant to subequant shapes (negative crystals are common, Fig. 2a). Three types of inclusions can be distinguished in the quartz phenocrysts:

Type I — inclusions containing liquid and a vapour bubble making up 10 to 30 percent of the total inclusion volume and occurring either in clusters or along healed fractures (Fig. 2a).

Type II — inclusions containing liquid and a large vapour bubble representing between 50 and 75 percent of the total volume of the inclusion (Fig. 2b). These inclusions commonly occur in planes with Type I inclusions (Fig. 2c). They may occur in clusters or be isolated.

Type III — halite-bearing inclusions containing a small vapour bubble occupying 10 to 20 percent of the total inclusion volume. A small proportion of these inclusions also contain a second smaller cubic phase with lower relief and rounded corners, tentatively identified as sylvite (Fig. 2d). Most Type III inclusions are isolated.

If it is assumed that isolated inclusions or inclusions in clusters were earlier than those obviously aligned along healed fractures, then it can be concluded that Type III inclusions were trapped early whereas Type I and II inclusions were trapped during early as well as later stages in the formation of the aureole.

Fluid inclusions that are related, i.e. that occur within the same planes or clusters, generally show fairly consistent phase ratios, indicating that any post-entrapment modification that they might have undergone was largely complete before the nucleation of additional phases. However, some clearly "necked down" inclusions were observed. A significant proportion of inclusions are grouped in domains where phase ratios are inconsistent. This is particularly common with Type I and Type II inclusions and could be the result of "necking down" or, alternatively, reflect mixed entrapment of a boiling or effervescing fluid. The relative proportions of Types I, II and III inclusions are 40 percent, 50 percent and 10 percent respectively.

Crushing tests

Several samples were tested for the presence of incondensible gases by crushing them in the presence of an immersion oil and observing the effect under the microscope. These tests confirmed that Type II inclusions consist largely of incondensible gases, and that Type I inclusions contain small volumes of these gases. The released gas bubbles are soluble in barium chloride indicating that they consist mainly of CO_2. No observations were made on the comparatively uncommon Type III inclusions.

Microthermometry

Fluid inclusion microthermometry was carried out using a gas-flow S.G.E. Model III heating/freezing stage that had been standardized from -90 to 600°C using procedures described in Roedder (1984). Temperatures were measured using a Chromel-Alumel thermocouple with an operating range of -133 to 1400°C that was placed on the sample and read from a digital trendicator.

Samples were prepared as doubly-polished sections 100-200 μm thick, and were examined under a microscope in order to select inclusions suitable for microthermometry. Inclusion leakage was tested by repeating runs. Phase change temperatures were repeated at least twice and were reproducible within ± 0.2°C at low temperature and ± 2°C at high temperature. All stage-obtained measurements have a precision of 1 percent.

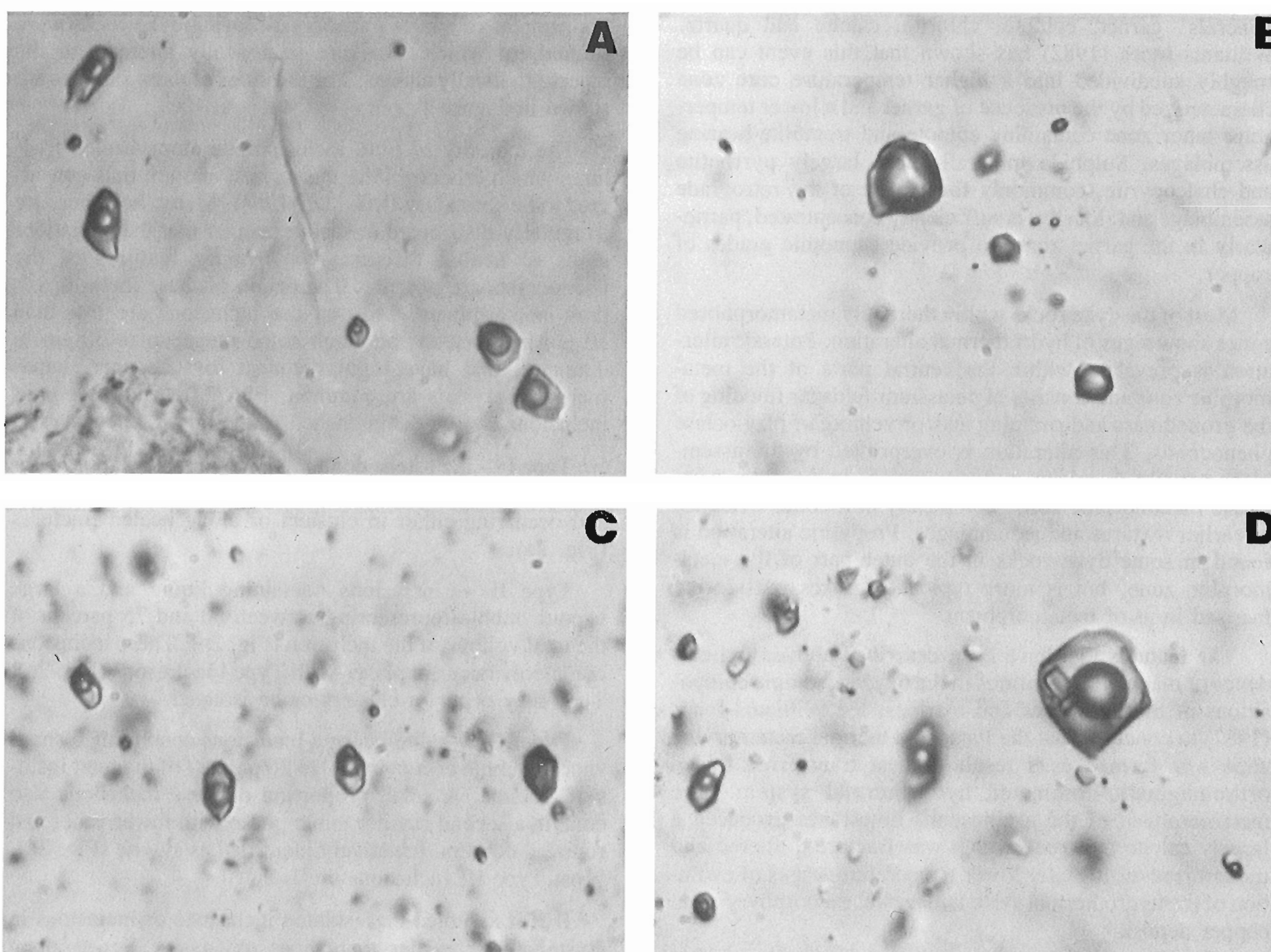

Figure 2. Typical Type I fluid inclusions exhibiting similar phase ratios **(a)**; a cluster of Type II fluid inclusions **(b)**; a plane containing Type I fluid inclusions and a single Type II fluid inclusion (dark), all with negative crystal forms **(c)**; a Type III fluid inclusion containing both halite (large crystal) and sylvite **(d)**. All photomicrographs are 60 μm wide.

Type I inclusions

Type I inclusions homogenize over a broad temperature range from 180 to 440°C with a broad peak centred at 350°C. Most inclusions, however, homogenize between 275 and 400°C (Fig. 3). Inclusions from potassically altered samples display a narrow well-defined peak of homogenization temperatures between 350 and 400°C, compared with a broader lower temperature peak (275 -375°C) for inclusions from phyllically altered samples (Fig. 4). Salinities calculated using freezing point depressions and a regression equation developed by Potter et al., (1978) range from 1.0 to 24.7 equivalent weight percent NaCl. Most inclusions, however, contain between 2 and 12 equivalent weight percent NaCl. (Fig. 5). Estimates of the initial melting temperature ranged from -42 to -10°C, but, because of the comparatively low salinities are undoubtedly high. They nevertheless indicate that some of the inclusions contained salts other than, or in addition to, NaCl and KCl: the systems NaCl-H_2O and NaCl-KCl-H_2O have eutectic temperatures of −20.8 and −23.2°C, respectively (Crawford, 1981). The most likely explanation for the low initial melting temperatures is the presence of dissolved calcium chloride which lowers the eutectic temperature of the system NaCl-H_2O to −52°C (Crawford, 1981). A very small proportion of Type I inclusions developed clathrates which melted at temperatures between 6 and 11°C. Salinities for these inclusions were calculated using the equation of Bozzo et al., (1973) and ranged up to 7.5 equivalent weight percent NaCl.

Type II Inclusions

All Type II inclusions homogenized to vapour at temperatures ranging from 300 to 500°C, with a well-defined peak between 350 and 400°C (Fig. 3). Homogenization was hard to estimate accurately because of the difficulty of observing the final disappearance of the liquid, and although this was made easier by choosing inclusions with re-entrants into which the last remnants of liquid could concentrate, it is likely that most of the estimates are somewhat low. It is, however, worth noting that the peak homogenization temperature for Type II inclusions is coincident with that for Type I inclusions from potassically altered samples (cf. Fig. 4). Owing to the small proportion of liquid, melting phenomena could only be observed in a small number of Type II inclusions. Final ice melting occurred between 0 and −7°C corresponding to an aqueous liquid with salinities between 0 and 10.5 equivalent weight percent NaCl (Potter et al., 1978). As noted earlier, crushing tests indicate that the Type II inclusions consist mainly of CO_2. This was confirmed by rapid cooling of the inclusions which produced double freezing at temperatures below −35 and −70°C corresponding to ice and in all probability CO_2 solid, respectively. However, it was not possible to observe the melting of the low temperature solid or clathrate. Owing to the paucity of information on the vapour phase, compositions of the Type II inclusions can only be qualitatively estimated. On the basis of liquid-volume ratios and phase densities (Potter and Brown, 1977; Parry, 1986), it is estimated that the Type II inclusions contain between 10 and 40 mole percent CO_2.

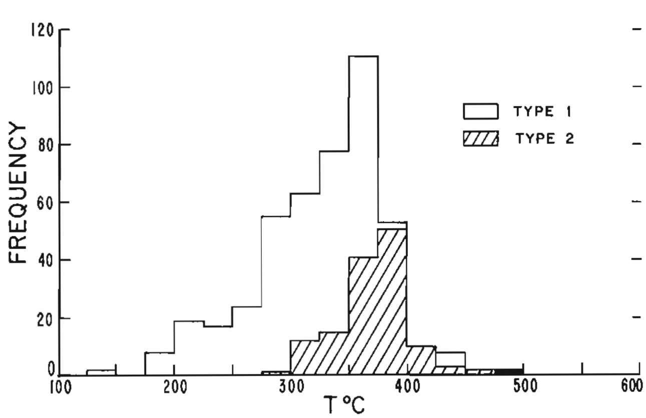

Figure 3. Histograms showing the distributions of homogenization temperatures of Type I and Type II fluid inclusions.

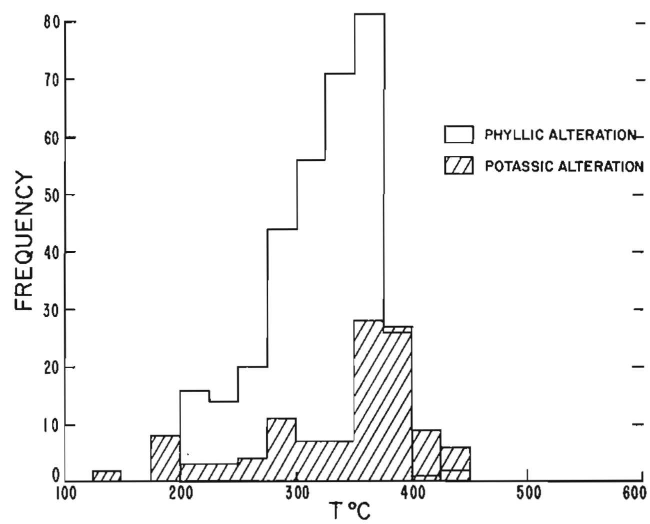

Figure 4. Histograms showing the distributions of homogenization temperatures of Type I fluid inclusions in potassically altered and phyllically altered samples.

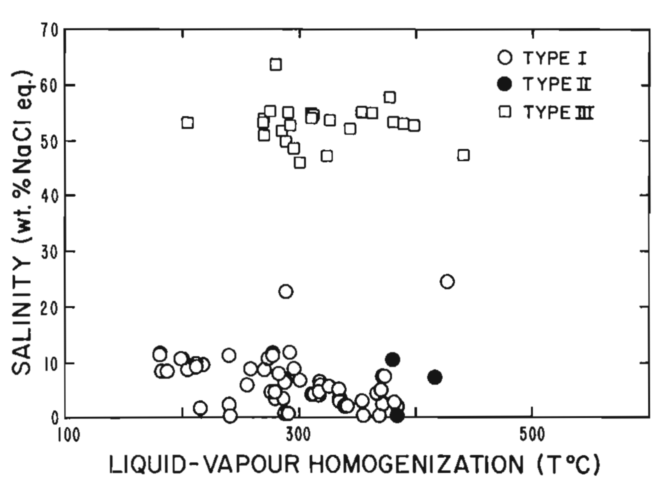

Figure 5. The salinities of Type I, Type II and Type III fluid inclusions versus liquid-vapour homogenization temperatures.

Type III Inclusions

In the Type III inclusions the vapour bubble is generally the first phase to disappear on heating, typically at temperatures between 275 and 400°C (Fig. 5), followed by halite at temperatures between 410 and 500°C. In the few sylvite-bearing Type III inclusions, sylvite was the first phase to dissolve, at temperatures between 130 and 170°C. Salinities of nonsylvite-bearing Type III inclusions were calculated using the halite dissolution temperatures and the regression equation of Potter et al., (1977). They range between 46 and 63 equivalent weight percent NaCl with a mean of 53 equivalent weight percent NaCl (Fig. 5). The salinities of the sylvite-bearing inclusions were determined using data for the ternary system NaCl-KCl-H_2O compiled in Roedder (1984). The salinities for these inclusions range from 60 to 64 weight percent NaCl $\pm$ KCl or, on average, 42 weight percent NaCl and 20 weight percent KCl. The regression equation of Potter et al., (1977) gives much lower salinities for these inclusions, ranging from 49 to 54 equivalent weight percent NaCl, suggesting that the salinities of the nonsylvite-bearing inclusions are probably somewhat underestimated.

DISCUSSION

Pressure-temperature conditions

Williams-Jones (1982) inferred, from estimates of the thicknesses of cover rocks, that lithostatic pressures, at the time of the formation of the Patapedia metamorphic zone, were of the order of 750-1150 bars. He used these values and thermal equilibrium data for various mineral assemblages in the metamorphic zone to calculate temperatures of prograde metamorphism ranging from 430°C at the margin to >480°C in the central parts, and retrograde metamorphism at temperatures above 350°C. The fluid inclusion data presented above, notably those for the Type III inclusions, provide an excellent opportunity to independently check these estimates (as the inclusions are trapped in quartz phenocrysts it is reasonable to assume that the corresponding fluid pressures will have been near lithostatic).

The homogenization of Type III inclusions by halite dissolution at temperatures much higher than those of the disappearance of the vapour bubble raises the question of whether or not the corresponding hydrothermal fluid was saturated in halite at the time of entrapment. If the fluid was saturated then the homogenization temperature provides a direct estimate of the entrapment temperature and a means to calculate the entrapment pressure. One indication of halite saturation is the presence of a "halite" trend amongst halite-sylvite-bearing inclusions (Cloke and Kesler, 1979).

The compositions of Type III inclusions containing halite and sylvite are shown on the NaCl-KCl-H_2O diagram presented in Figure 6 together with "halite trends" for Granisle-Bell and Panguna (Wilson et al., 1980). Although there are only 5 data points they do nevertheless appear to generate a trend similar to and intermediate between those of Granisle-Bell and Panguna. This suggests that the sylvite-halite-bearing Type III inclusions and possibly most Type III at Patapedia were saturated in halite at the time of their entrapment.

Entrapment pressures of Type III inclusions were determined using the method of Roedder and Bodnar (1980), the density expression of Bodnar (1983) and a P-V-T equation of state developed by Bowers and Helgeson (1985). The resulting pressures and corresponding homogenization temperatures are plotted on Figure 7 and concentrate in a field ranging from 700 to 1000 bars and 450 to 500°C.

Also plotted on Figure 7 are the bounding isochores, determined using the data of Potter and Brown (1977), for the high temperature Type I inclusions from potassically altered samples (see Fig. 4). It should be noted that these bounding isochores are approximate, since, as has previously been shown, Type I inclusions do contain some CO_2. The interpreted conditions for the Type III inclusions occur within the field for this set of Type I isochores. This suggests that the high temperature Type I inclusions from potassically altered samples and the Type III inclusions were trapped at similar conditions of 700-1000 bars and 450-500°C.

The Type II inclusions, as noted earlier, are estimated to consist of a low salinity aqueous liquid <7 equivalent weight percent NaCl and 10 to 40 mole percent CO_2 vapour. Since these inclusions homogenize to vapour it can be concluded that they were entrapped on the dew point curve, i.e. at pressures below the critical pressure for the corresponding system. If the inclusions are modeled by a H_2O-CO_2-NaCl fluid containing 6 weight percent NaCl in the aqueous phase, for which P-V-T data have been determined experimentally by Gehrig (1980), then it can be concluded that pressures were <1000 bars and <700 bars assuming 40 mole percent CO_2, and 10 mole percent CO_2, respectively. The corresponding critical temperatures are 420 and 500°C, respectively, which are within the range of estimated homogenization temperatures for the Type II

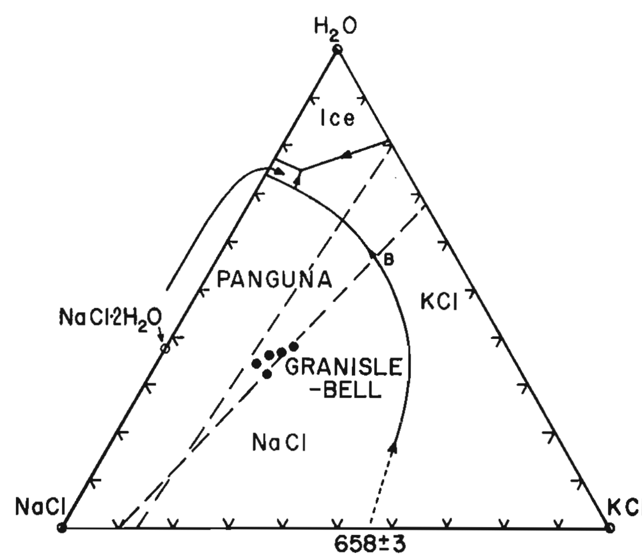

Figure 6. A ternary H_2O-NaCl-KCl diagram showing the compositions of sylvite-bearing Type III fluid inclusions. Also shown on the diagram are the trends of compositions of similar fluid inclusions in the Granisle-Bell, and Panguna porphyry copper deposits (taken from Wilson et al., 1980). The phase boundaries are reproduced from Roedder (1984).

inclusions. It would thus appear that Type II inclusions were largely trapped at conditions similar to those of Type III inclusions and the higher temperature Type I inclusions.

These estimates of the entrapment conditions for the three fluid types determined above are similar to those previously made for the conditions of prograde metamorphism in the Patapedia thermal zone.

Nature and origin of the fluids

This study has shown that three distinct fluid types participated in the evolution of the Patapedia thermal metamorphic zone, namely a low salinity, high density aqueous fluid, a high salinity, high density aqueous fluid and a low density carbonic fluid. Furthermore, the petrographic data suggest that Type III fluids were trapped early whereas Type I and II fluids were trapped during early as well as later stages in the history of the thermal zone.

We have already summarized the evidence presented by Williams-Jones (1982) in support of a hydrothermal origin for the prograde metamorphism by a porphyry style system. The widely accepted model for the evolution of such systems (Burnham, 1979) is that calc-alkaline magmas, intruded at high crustal levels, become saturated with water and, during crystallization, evolve a separate aqueous fluid. The reaction forming this fluid phase has a large ΔV, which produces the energy necessary to fracture the overlying rocks and release the fluid upward. The nature of this fluid is in question. Many authors (e.g. Hein and Tistl, 1987), noting the common occurrence of high salinity inclusions in porphyry deposits, have postulated high salinity orthomagmatic fluids. Others, notably Burnham (1979, 1981), from theoretical considerations of initial chloride and water contents of granodiorite magmas and chloride melt-vapour partition factors (cf. Kilinc and Burnham, 1972), have argued in favour of a low salinity fluid (10 equivalent weight percent NaCl), which is subsequently modified by boiling or condensation (cf Henley and McNabb, 1978; Burnham, 1979) to produce the commonly observed high salinity inclusions. During later stages this system collapses and is replaced by a meteoric water-dominated hydrothermal system that overprints and, to varying degrees, destroys evidence of the earlier system.

The Type I inclusions form the majority of the fluid inclusion population at all temperatures and can therefore be assumed to represent the fluid mainly responsible for the various metamorphic and alteration effects in the Patapedia thermal zone. They have compositions which, as can be seen from the above discussion, are consistent with an orthomagmatic origin. However, their estimated maximum entrapment temperatures are much lower than fluid inclusions of demonstratable orthomagmatic origin (cf. Roedder, 1971; Wilson et al., 1980, Reynolds and Beane, 1985). This suggests that the high temperature Type I inclusions (potassic alteration) either represent a cooled orthomagmatic fluid or alternatively that they originated from heated meteoric fluids, or mixed meteoric-orthomagmatic fluids. The lower temperature Type I inclusions, which are associated with phyllic alteration, probably represent later heated meteoric fluids.

The heating of the host argillaceous limestones by the above aqueous fluids and the consequent metamorphism of the former would have produced large quantities of CO_2, which, except during the early very high temperature stage, would have been largely immiscible in the aqueous fluids (Bowers and Helgeson, 1983b). We believe that samples of this immiscible CO_2-rich phase (vapour) form the very common Type II inclusions (50 percent of the inclusion population). This fluid type would have been produced and trapped throughout the prograde metamorphic event, and also during retrograde metamorphism in rocks containing unreacted calcite.

The Type III inclusions, which constitute only 10 percent of the Patapedia fluid inclusion population, would, according to traditional "porphyry" models, represent either the original orthomagmatic fluid or its condensate. An alternative explanation, which we favour, is that the Type III fluid inclusions represent liquids that separated from low to moderate salinity CO_2-bearing fluids formed during the early orthomagmatic-dominated stages of the hydrothermal system. At high temperatures and pressures CO_2 is completely miscible in low to intermediate salinity fluids (Bowers and Helgeson, 1983a). On cooling to the temperature-pressure conditions inferred for the bulk of the metamorphism (450-500°C and 700-1000 bars) such fluids could have exsolved liquids containing >50 weight percent NaCl (cf. Bowers and Helgeson, 1983b), i.e. similar in composition to the Type III fluid inclusions.

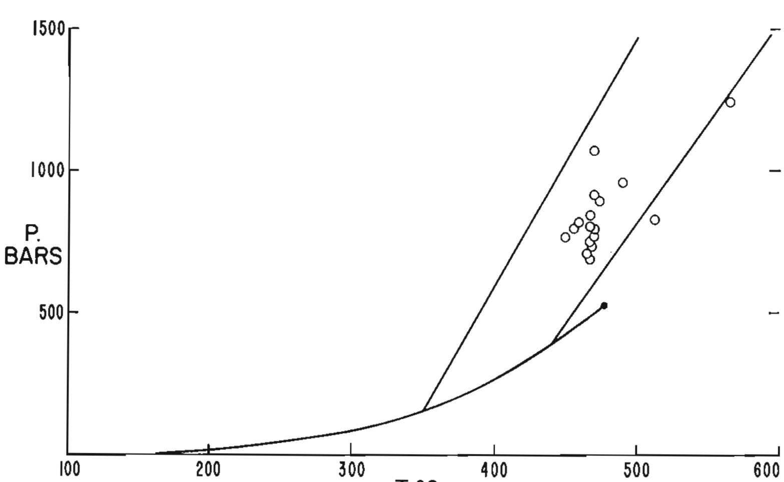

Figure 7. A pressure-temperature diagram showing the estimated conditions of entrapment of Type III inclusions and the bounding isochores for high temperature Type I fluid inclusions from potassically altered samples (see text for an explanation of how these were determined). Also shown on the diagram is the liquid-vapour curve for a 10 weight percent NaCl aqueous solution.

Schematic T-XCO_2 diagrams showing the development of the above fluid types is given in Figure 8.

The relationships between alteration in the dyke rocks and prograde and retrograde metamorphism of the sedimentary rocks are unclear. We suggest that prograde metamorphism coincided with potassic alteration, that early retrograde metamorphism was also probably accompanied by potassic alteration (association of mineralization with retrograde metamorphism in zones of potassic alteration), but that the bulk of retrograde metamorphism was associated with phyllic alteration. If this is correct then our interpretation is that the Type III fluids were restricted to the early part of the prograde metamorphic event, that Type II fluids formed throughout prograde metamorphism and to a lesser extent during retrograde metamorphism and that Type I fluids were present throughout the evolution of the thermal zone.

Metamorphic model

Williams-Jones (1982) modeled stability relationships for mineral assemblages identified in the Patapedia thermal zone on the assumption that there was a single homogeneous fluid phase consisting only of CO_2 and H_2O. However, as has already been discussed, such an assumption is not justified.

We have shown that low to moderate salinity fluids were probably present in large volumes during the metamorphism and have accordingly re-evaluated the stability relationships to take into account the effects of NaCl. The phase diagram shown in Figure 9 is for a system containing 12 weight percent NaCl (data from Bowers and Helgeson (1983b)) and shows that there is a large immiscibility field that extends to temperatures above those corresponding to the diopside-phlogopite-in reaction boundary. This is consistent with the conclusion drawn earlier, that much of the metamorphism occurred in the presence of two fluids. If, as has been argued, an aqueous liquid was the principal agent of heat transfer then it follows that the temperature at each isograd was determined by the intersection of the corresponding reaction boundary with the low XCO_2 edge of the solvus. These temperatures range from 410°C (Reactions 1 and 2 at the edge of the zone) to 465°C (Reaction 7).

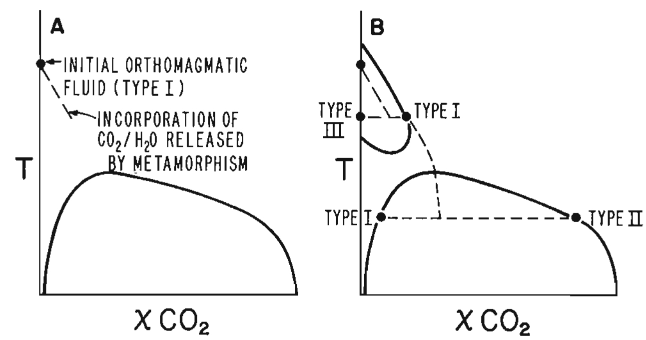

Figure 8. Schematic T-XCO_2 diagrams showing the compositional evolution of a moderate salinityorthomagmatic fluid as it rises, interacts with overlying calcareous sedimentary rocks and cools. Diagram A depicts the fluid path at high pressure (shown by the dashed line). At this pressure there is only one field of immiscibility in the system H_2O-CO_2-NaCl. However, as pressure decreases a second, high temperature, field of fluid immiscibility forms. Diagram B shows the successive intersection of the evolving fluid with these two fields of fluid immiscibility. At high temperature, a high salinity, low XCO_2 fluid (Type III) is exsolved and at lower temperature, a high XCO_2 fluid (Type II). At temperatures below the second solvus Type I and II fluids are also produced by continuing metamorphism and ultimately the influx of heated meteoric fluids.

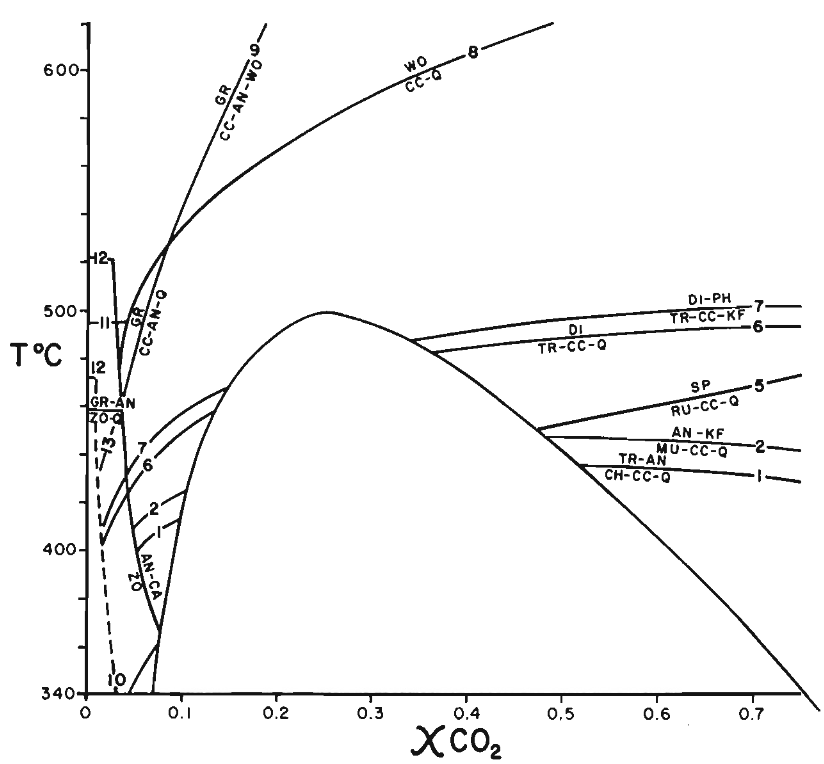

Figure 9. A T-XCO_2 phase diagram at 1 Kb redrawn from Williams-Jones (1982) showing stability relationships in the one phase H_2O-CO_2 region for the principal mineral assemblages observed in the Patapedia thermal zone. Several less important reactions have been omitted for the sake of clarity. The solvus marking the boundary of the two phase region for a NaCl(12 weight percent) -H_2O-CO_2 solution (Bowers and Helgeson, 1983a) is also shown.

At the highest grade conditions, corresponding to the wollastonite isograd, the fluid would have been homogeneous for all XCO_2 and the XCO_2 of the fluid on the reaction boundary would thus have been determined by the amount of aqueous fluid in the rock, its composition and the amount of CO_2 produced by reaction. The temperature and XCO_2 corresponding to the wollastonite-in reaction therefore cannot, unfortunately, be determined. Temperatures as high as 650°C and XCO_2 of 1 are possible although unlikely.

The subsequent retrograde metamorphism was associated with a much lower XCO_2 fluid constrained to be on the low XCO_2 side of Reactions 9 and 10. It is likely, however, that this fluid was heterogeneous or effervescing, since, as argued earlier, pressures are likely to have been much lower than during the prograde event, and as has been demonstrated by Bowers and Helgeson (1983b), the immiscibility field for $NaCl-H_2O-CO_2$ fluids increases sharply with decreasing pressure.

Controls of mineralization

As has already been noted, copper mineralization occurs in garnet-bearing calc-silicate (altered) hornfels. It can, accordingly, be concluded that copper mineralization took place after the peak of metamorphism. However, the concentration of this copper in zones where the dyke rocks are potassically altered, and the presence of sphalerite stars in the chalcopyrite (Williams-Jones, 1982), suggests that the temperature of mineralization was comparatively high. From Figure 9 it can be seen that grossular-anorthite (garnet and plagioclase [An_{30}] are very commonly associated with copper mineralization) breaks down to zoisite-quartz at temperatures below 460°C. This temperature is lowered by the activity of sodium in the plagioclase, and, for a composition of An30, decreases to 400°C. Such a temperature is consistent with that predicted by the lower part of the range for the high temperature Type I fluid inclusions from potassically altered samples particularly, if, as seems likely from the widespread fracturing, fluid pressures were somewhat lower than the near lithostatic conditions that presumably prevailed at the peak of metamorphism.

The two factors that are considered most likely to have been responsible for mineral deposition are decreasing temperature and CO_2 effervescence. As shown by Crerar and Barnes (1976), a 100°C drop in temperature in the mesothermal range will result in a two orders of magnitude decrease in the solubility of chloride complexed copper (the most likely form of aqueous copper in the Patapedia hydrothermal fluids). Thus, if copper was near saturation at 400°C, it would all effectively have been deposited by the time that the fluid had cooled to 300°C. Such temperature decreases are readily achieved by adiabatic expansion of the hydrothermal fluids during fracture-related pressure releases and by mixing of the fluids with cooler meteoric waters, both of which probably occurred at Patapedia. The importance of volatile effervescence as a major control of sulphide deposition has been underlined by the recent work of Drummond and Ohmoto (1985). They have demonstrated that as little as 5 percent boiling or effervescence of volatiles (notably CO_2) can result in a pH increase sufficient to cause deposition of 99 percent of the metals carried in a saturated solution. The evidence of this study suggests that CO_2 effervescence was an essential part of the hydrothermal history of the Patapedia thermal zone. We thus conclude that the subeconomic copper mineralization found at Patapedia was deposited in response to falling temperature and/or CO_2 effervescence.

SUMMARY AND CONCLUSIONS

This study has shown that the formation of the Patapedia thermal metamorphic zone was the result of the transfer of heat from a porphyry-style hydrothermal system dominated by fluids of low to intermediate salinity (Type I inclusions). The early, probably quite localized, metamorphic effects were caused by orthomagmatic or mixed orthomagmatic/meteoric fluids at temperatures >500°C and possibly >600°C (wollastonite formation). Carbon dioxide was produced by devolatilization of the calcareous sediments and mixed with the orthomagmatic fluids to form homogeneous H_2O-CO_2 fluids. On cooling these latter fluids separated into high salinity aqueous liquids (Type III inclusions) and CO_2 vapours. The main prograde metamorphic event was caused by cooled orthomagmatic, mixed meteoric-magmatic or meteoric fluids at temperatures between 450 and 500°C. This resulted in large scale production of CO_2 which was largely immmiscible with the hydrothermal fluid and therefore formed a separate vapour phase (Type II inclusions). The result of the metamorphism of the calcareous sediments, excluding pure limestones which were largely unaffected, was the formation of a diopside-bearing hornfels essentially devoid of calcite. Subsequent metamorphism (retrograde) of this rock took place in a regime of declining temperatures and in the presence of a fluid with an XCO_2 low enough to form garnet and epidote-bearing assemblages. The associated copper sulphide mineralization occurred in response to these falling temperatures and/or the pH increase caused by continued CO_2 effervescence.

ACKNOWLEDGMENTS

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Geology and petrology of the Handcamp gold prospect, Robert's Arm Group, Newfoundland[1]

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Hudson, K.A. and Swinden, H.S., Geology and petrology of the Handcamp gold prospect, Robert's Arm Group, Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 93-105, 1989.

Abstract

The small polymetallic Handcamp prospect is hosted by Lower Ordovician Robert's Arm Group volcanic and sedimentary rocks in north-central Newfoundland. The gold-sulphide mineralization is restricted to a 50 m wide structurally complex zone dominated by volcanic and volcaniclastic rocks, red and ferruginous chert and argillaceous rocks, which have been variably altered to a quartz-sericite-pyrite-magnetite assemblage. Alteration and sulphide mineralization appear to be both synkinematic and postkinematic. The association of late Ca-Fe-Mg carbonates and silicates with sulphides and gold suggests that CO_2-rich fluids were instrumental in the mineralizing process. Preliminary geochemical data indicate that gold enrichment correlates with enrichment of Sb, As and Mo.

The above evidence suggests that the Handcamp prospect may have been generated through much later shear-related hydrothermal deposition, rather than by exhalative volcanogenic processes, as proposed by earlier workers. If true, this has important implications for exploration for additional gold mineralization in the Robert's Arm Group.

Résumé

Les roches volcaniques et sédimentaires du groupe de Robert's Arm, datant de l'Ordovicien inférieur, dans la partie nord-centrale de Terre-Neuve, renferme la petite zone de production polymétallique possible de Handcamp. La minéralisation en or et sulfures se limite à une zone structurale complexe de 50 m de large, principalement occupée par des roches volcaniques et volcanoclastiques, des cherts rouges et ferrugineux et des roches argileuses qui ont été diversement altérées en une association de quartz, séricite, pyrite et magnétite. Il semble que l'altération et la minéralisation sulfurée soient à la fois syntectoniques et post-tectoniques. L'association des carbonates et silicates tardives de Ca, Fe Mg avec les sulfures et l'or semble indiquer que des fluides riches en CO_2 ont contribué au processus de minéralisation. Des données géochimiques préliminaires indiquent que l'enrichissement en or correspond à l'enrichissement en Sb, As et Mo.

Les indices ci-dessus semblent indiquer que la zone de production possible de Handcamp a peut-être été créée par une sédimentation hydrothermale beaucoup plus tardive liée à une phase de cisaillement, et non par des processus exhalatifs d'origine volcanique, hypothèse auparavant proposée par d'autres chercheurs. Si tel est le cas, ceci aura d'importantes conséquences pour l'exploration d'autres minéralisations en or du groupe de Robert's Arm.

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

The Handcamp prospect is a small base and precious metal occurrence located in north-central Newfoundland (Fig. 1). It is hosted by Lower Ordovician volcanic and sedimentary rocks of the Robert's Arm Group, a well-known and prolific host of volcanogenic sulphide mineralization, including former producers at Gullbridge and Pilley's Island. Most volcanogenic sulphide occurrences in this area contain significant concentrations of base metals but little gold (Swinden and Sacks, 1986; Swinden, 1987, 1988a). The Handcamp prospect is anomalous in that it is the only showing in the Robert's Arm Group for which gold is the principal metal of value. Its characteristics, origin, setting and relationships to other showings are therefore of particular interest for interpretations of regional metallogeny and for guiding exploration in the area.

The origin of the gold and accompanying sulphide mineralization at Handcamp has been the subject of controversy since its discovery in 1928. A wide range of genetic processes have previously been proposed, including epigenetic granodiorite-related mineralization (A.V. Corlett, unpublished report, 1930), low temperature replacement deposition (Anonymous, unpublished report, Newfoundland Exploration Company Limited, 1940), shear zone-related mineralization (J.M. Neilson, unpublished report, Dames and Moore Consultants, 1972), stratabound volcanogenic mineralization (J.R. DeGrace, unpublished report, Newfoundland Exploration Company Limited, 1976; A. Franc de Ferriere, unpublished report, Falconbridge Nickel Mines Limited, 1978; B.J. Fryer, unpublished report for Lew Murphy, 1980) and hot spring deposition (W.B.Burton, unpublished report, US Borax, 1983). However, few of these models were based on detailed, systematic geological studies, and many relied heavily on optimistic exploration strategies rather than substantive data.

This paper presents the preliminary results of detailed mapping carried out in 1985, 1986 and 1988, from relogging of drill core from the Handcamp prospect (1986), and from observations of sixty thin and polished sections (1988). With these data, we attempt to assess the applicability of various genetic models to the prospect. Further detailed geochemical, isotopic and petrological studies directed at this genetic question are in progress.

This work forms part of a larger metallogenic study of the Robert's Arm-Buchans Group funded by the Geological Survey of Canada under the 1984-1989 Canada-Newfoundland Mineral Development Agreement.

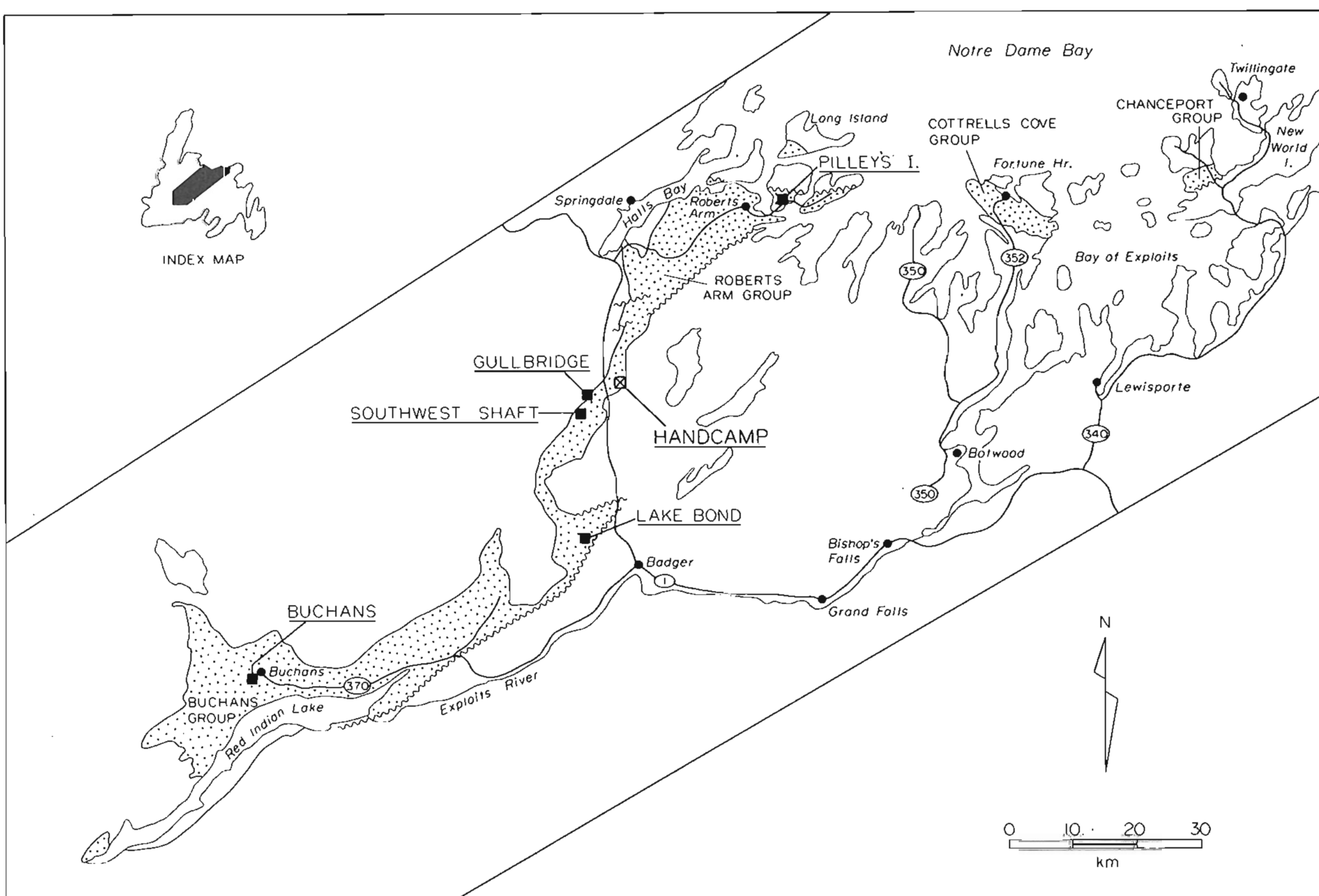


Figure 1. Location of the Robert's Arm-Buchans Group (stippled). Volcanogenic sulphide deposits indicated by filled squares.

REGIONAL GEOLOGY

The Robert's Arm Group of central Newfoundland is an Early Ordovician sequence of marine volcanic, subvolcanic and volcaniclastic rocks which extend from Notre Dame Bay southward to Lake Bond (Fig. 1). Rocks of the Robert's Arm Group are continuous with and stratigraphically equivalent to those of the Buchans Group to the south. Swinden and Sacks (1986) in their mapping of the Roberts Arm Group tentatively subdivided the stratigraphic sequence in the southern part of the area (Fig. 2) into five major units: the Baker Brook basalt, Gull Hill sediments, Burnt Island basalt, Gullbridge felsic volcanics and South Brook basalt. Copious evidence for shearing and faulting in the area suggests that the contacts between these units may be structural rather than stratigraphic, and our preliminary unpublished geochemical data for the volcanic rocks suggest that even the mapped units may be structurally composite. Host rocks to the Handcamp prospect have been correlated with the Gullbridge felsic volcanics by Swinden and Sacks (1986). The southern part of the Robert's Arm Group, including the area of the Handcamp prospect, is considerably attenuated and deformed, and contains a thermal metamorphic aureole due to the intrusion of major posttectonic granitoid bodies both to the east (Twin Lakes Complex) and west (Topsails Intrusive Suite) (Fig. 2).

The Robert's Arm Group hosts three volcanogenic deposits (Gullbridge, Lake Bond and Pilley's Island) and numerous other relatively minor Cu $\pm$ Zn showings, many of which have been interpreted to be volcanogenic stockworks by Swinden and Sacks (1986) and Swinden (1987). Pb and Ag are rare in these showings, except for certain massive sulphide lenses at Pilley's Island (Tuach, 1988); Au is seldom above background values, even in the massive sulphide bodies.

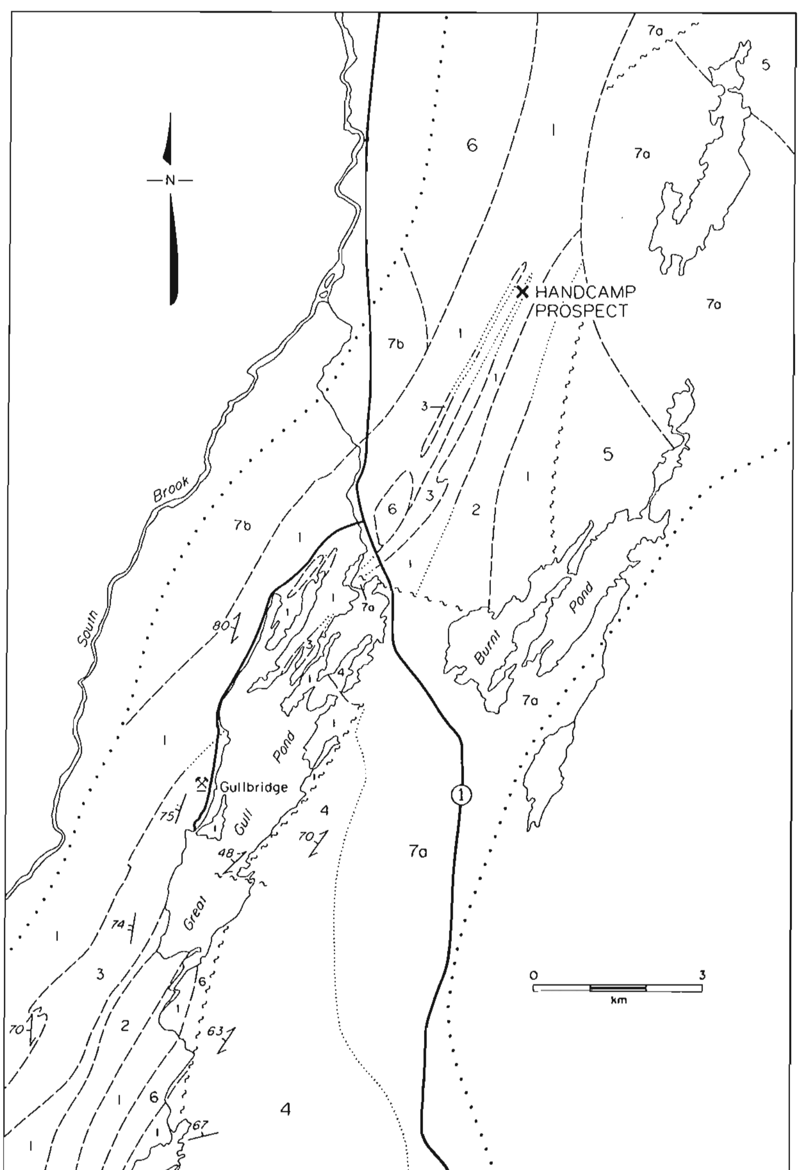


Figure 2. Geology of the southern part of the Robert's Arm Group, showing the location of the Handcamp prospect. Heavy dots outline the extent of mapping carried out for the Robert's Arm metallogenic study.

LEGEND

Devonian and Younger

7 — Granitoid intrusive rocks: 7a — Twin Lakes Complex; 7b — Topsails Intrusive Suite

6 — Mafic intrusive rocks; gabbro, diorite, diabase, pyroxenite

Ordovician — Silurian

5 — Sansom Greywacke; greywacke, fine grained conglomerate with a volcanic provenance

Ordovician — Silurian?

4 — Metamorphic rocks; quartzite, andalusite and staurolite-bearing pelite, carbonaceous shale of uncertain origin

Ordovician

ROBERT'S ARM GROUP

3 — Felsic volcanic rocks — rhyolitic ash-flow tuff, lesser rhyolite flows, minor mafic schist (informally named Gullbridge felsic volcanics)

2 — Sedimentary rocks — volcaniclastic chert, ferruginous chert, sandy and fine conglomerates (informally named Gull Hill sediments)

1 — Undivided mafic volcanic rocks — pillow basalt, massive basalt, mafic tuff, chloritic schist (informally named Baker Brook basalt, Burnt Island basalt and South Brook basalt)

SYMBOLS

Geological contact; approximate, assumed

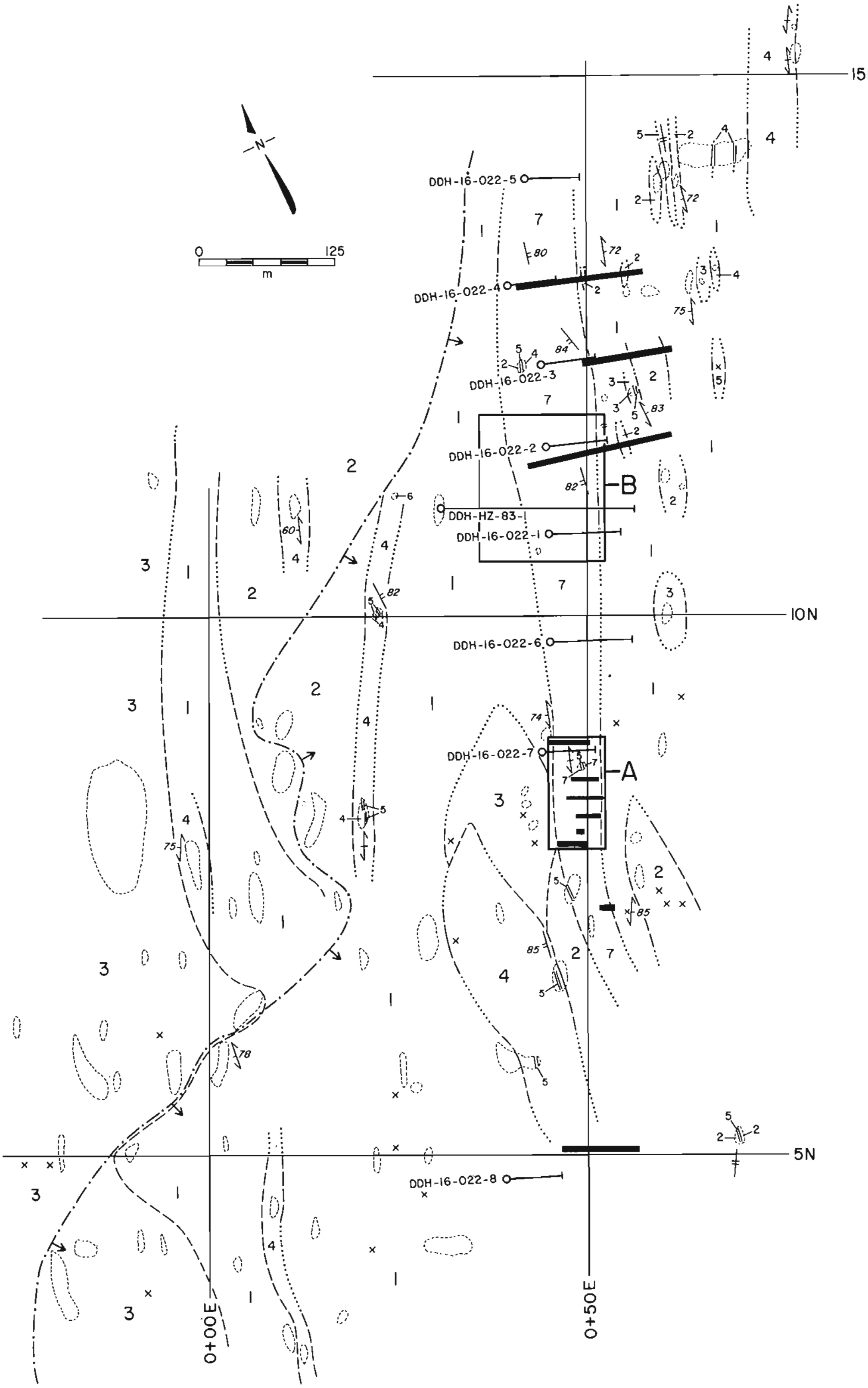
Strike and dip of bedding; tops known

Strike and dip of bedding; tops unknown

Strike and dip of principal foliation

Fault; assumed

Mine; abandoned



LEGEND

ROBERT'S ARM GROUP

8 — Quartz-sericite-pyrite rocks

7 — Mineralized zone — felsic volcanic and volcaniclastic rocks, red chert and ferruginous cherts and quartz-sericite-pyrite rocks

6 — Black argillites

5 — Red to black chert

4 — Felsic volcanic rocks — lapilli and crystal tuffs, and ash flows

3 — Massive dark to medium green mafic rock; may be unpillowed flows;

3a — intrusive mafic rock with primary amphibole

2 — Black to dark green, schistose mafic tuffs

1 — Pillow basalt

SYMBOLS

Geological contact; approximate, assumed

Strike and dip of principal foliation

Strike and dip of bedding; tops unknown

Mylonite or shear zone

Outcrop

Trench

Grid lines and markers

Diamond-drill hole

Limit of epidote alteration

Figure 3. Detailed geology of the Handcamp prospect
Outlined areas: A: Figure 3a
B: Figure 4

LOCAL GEOLOGY

The Robert's Arm Group in the vicinity of the Handcamp prospect is dominated by dark greenish-grey to reddish-brown and black pillow basalt, pillow breccia, tuffs and massive flows and lesser felsic volcanic rocks. Chert and argillite are locally interbedded in the sequence. Intermediate to mafic plutonic rocks, probably related to the Devonian (?) Twin Lakes Complex, have intruded the extreme eastern and western portions of the Handcamp area (Fig. 2). A detailed geological map of the prospect is presented in Figure 3.

Pillows in unit 1 basalt (Fig. 3) range in size from 15-25 cm, and generally have well-defined, sometimes epidote-lined margins. They are usually moderately to strongly flattened. No original minerals remain; all have been altered to a spilitic assemblage of albite-chlorite-biotite-epidote.

The pillowed basalts pass eastward into massive dark to medium green mafic rock (unit 3, Fig. 3) which may be unpillowed flows. However, these rocks locally contain primary amphibole and may be in part intrusive, as suggested by Burton (unpublished report, US Borax, 1982). Black to dark green, generally schistose mafic tuffs are also present (unit 2, Fig. 3). They contain flattened spherules which in some cases have pyrite-rich cores, enveloped by quartz, sericite and blocky epidote crystals. The latter are in turn rimmed by a layer of quartz and disseminated pyrite. The rocks are variably epidotized and chloritized.

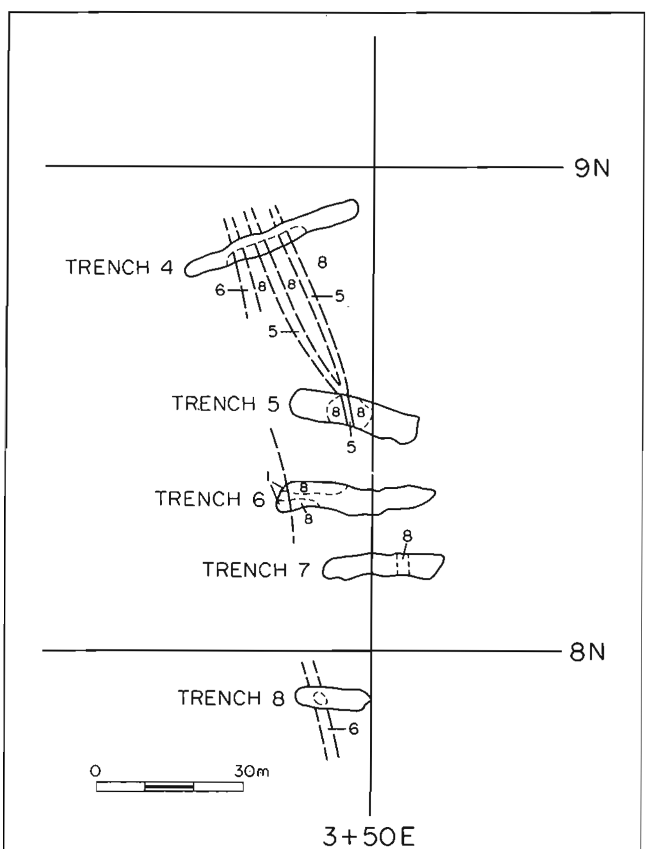


Figure 3a. Geology of trenches showing discovery trench (Trench 4).
Legend as in Figure 3.

Thin bands and lenses of felsic volcanic rocks occur throughout the Handcamp area (unit 4, Fig. 3). They range from coarse white lapilli tuffs interbedded with jasper and white chert, through pink-weathering quartz-feldspar crystal tuffs with large feldspar crystals, to ash flows with feldspar fragments and laminations which may be relict flow banding. The tuffaceous lithologies commonly contain intercalated beds and lenses of red chert up to 0.5 m thick (unit 5, Fig. 3).

Dark grey to black argillites, grey-green chert and cherty sediments are erratically distributed throughout the sequence, but are particularly common in the stratigraphic interval in which mineralization occurs. The mineralized and altered rocks of the mineralized zone form a distinctive roughly strata-parallel zone at the scale of mapping and are indicated as unit 7 on Figure 3.

All rocks in the Handcamp area are variably deformed. A prominent penetrative schistosity affecting all rocks trends roughly east-northeast. Although no folds related to this fabric were seen in the Handcamp area, a similar fabric in the Gullbridge area to the south is axial planar to isoclinal D_1 folds (Swinden and Sacks, 1986). Deformation is locally very intense, particularly in the mineralized zone (see below), with evidence of shearing and related formation of tight to isoclinal folds and the local development of mylonite in major shear zones.

MINERALIZATION AND RELATED ALTERATION

Since its discovery in 1928 by the Central Mineral Belt Syndicate, the Handcamp prospect has been investigated in a number of exploration programs by mapping, trenching, electrical and magnetic geophysical methods, soil geochemistry and shallow drilling. The most comprehensive investigation of the mineralized zone was carried out by Falconbridge Nickel Mines Limited between 1977 and 1979 (A. Franc de Ferriere, unpublished report, 1978; J. Hinchey, unpublished report, 1980). The mineralized zone was extensively trenched along its strike extent, a large area was stripped and washed near where the best Au values were obtained (Fig. 4) and 12 holes were drilled in an attempt to define the extent and character of the mineralization. Though assays as high as 587 g/t Ag and 74 g/t Au have been recorded from single samples, concentrations are generally very much lower and it has not proved possible to outline a significant tonnage of consistently mineralized rock.

Mineralization in the Handcamp area consists of pyrite, chalcopyrite, sphalerite and minor galena with erratically distributed gold. The base and precious metal values are restricted to a zone 50 m wide, characterized by felsic volcanic and volcaniclastic rocks, abundant red chert and ferruginous chert, which has been traced in outcrop and diamond-drill holes for more than 1200 m along strike (and to a depth of <50 m). Within this zone, sulphide and gold mineralization occur erratically in patches, lenses, veinlets and disseminations hosted by altered argillite, red and white chert and

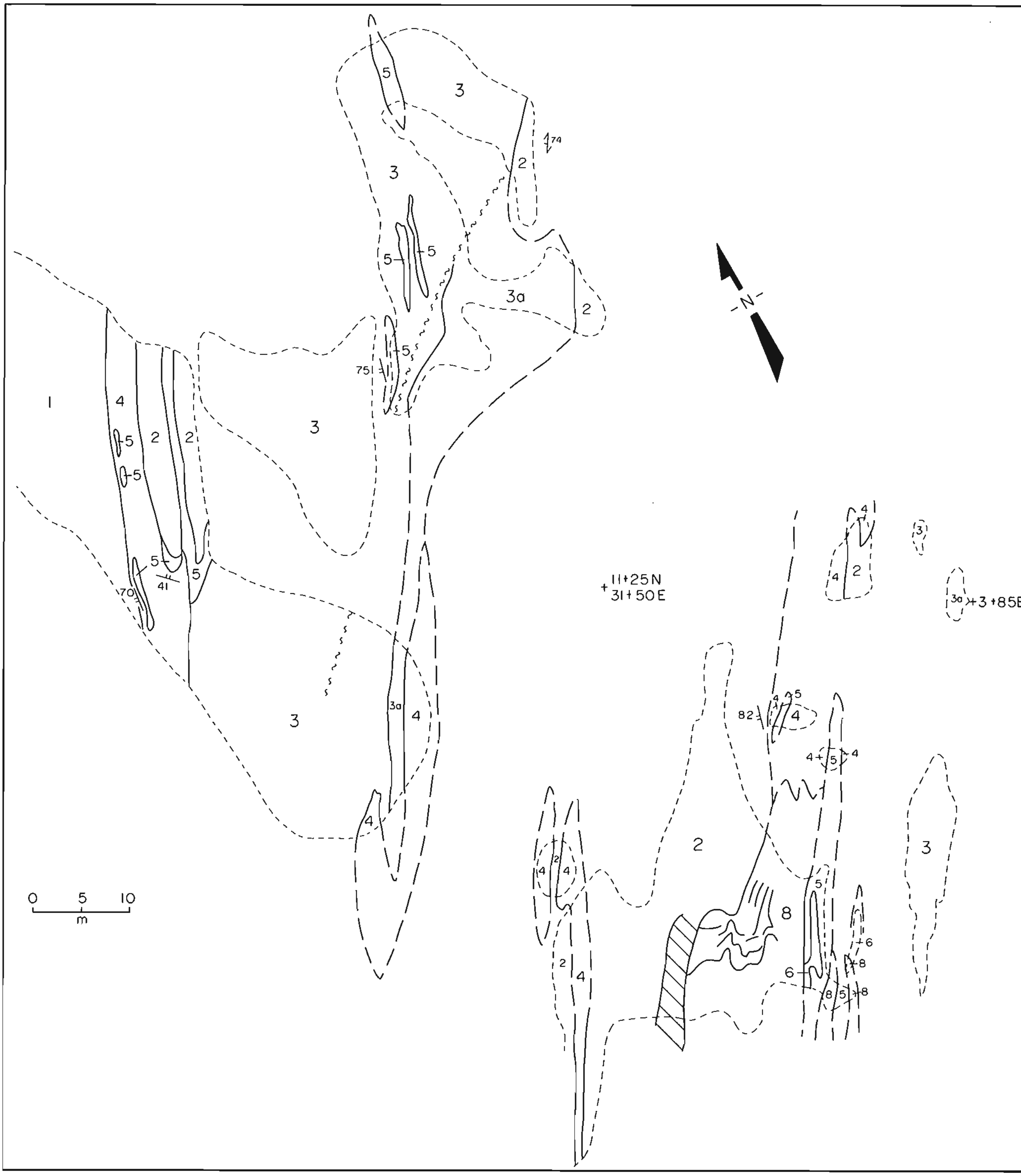


Figure 4. Detailed geology of the "cleared area" within the Handcamp map area. Legend and symbols as in Figure 3.

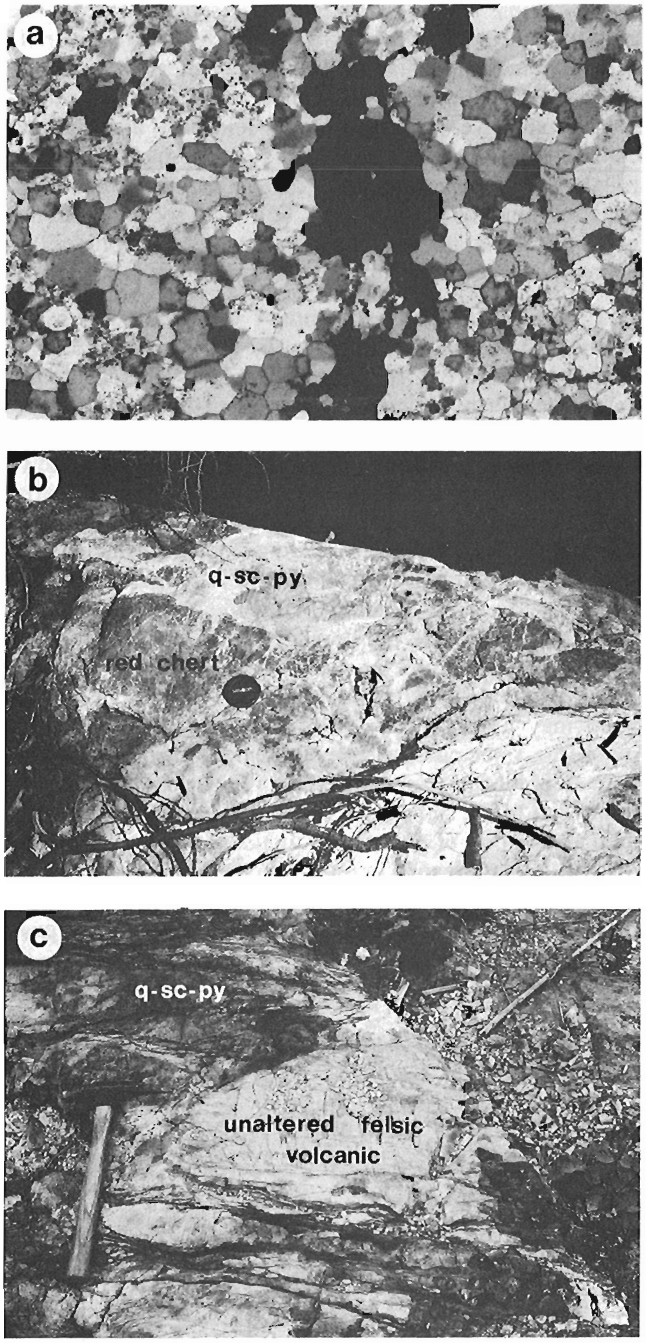


Figure 5. Photomicrographs and photographs of hydrothermally altered rocks at Handcamp.

a) Typical quartz-sericite-pyrite alteration, cut by a sphalerite veinlet. Field of view 2.2 mm, partially crossed polars (85S698D).

b) Relict fragments of red chert in quartz-sericite-pyrite altered rock. Width of camera lens cap is 5 cm.

c) Sharp contact between unaltered felsic volcanic rocks (white) and rusty quartz-sericite-pyrite altered rocks.

altered felsic and mafic tuffs. The apparent stratabound nature of the mineralization, at least on a map scale, has led many workers to suggest that the mineralization is stratabound and that its genesis is in some way tied to the deposition of the felsic volcanic and sedimentary rocks that characterize this zone (A. Franc de Ferriere, unpublished report, for Lee Murphy, 1978; B.J. Fryer, unpublished report, Falconbridge Nickel Mines Limited, 1980).

The nature of the alteration intimately associated with the mineralization is best revealed in the "cleared area" that was stripped and washed during the Falconbridge exploration program (Fig. 4). Within and near the mineralized zone, a pervasive and intense alteration, dominated in outcrop by variable silicification, sericitization and pyritization, with less common development of epidote and magnetite, has been superimposed upon the regional greenschist facies assemblage. Within the mineralized zone, rocks have been pervasively transformed to a quartz-sericite-pyrite assemblage (Fig. 5a) with superimposed quartz-pyrite and, more rarely, magnetite veins. The original lithologies have generally been completely replaced, although relict patches reveal felsic volcanic, black argillite (now biotite schist) and red chert (Fig. 5b) protoliths. Contacts between the completely altered rocks and their less altered equivalents are always sharp and appear to be structural. This is illustrated by Figure 5c, in which quartz-pyrite-sericite rocks are seen to be in sharp contact with buff white, unaltered felsic volcanic rocks. Lens-shaped fragments of these altered rocks are locally observed in mylonite and shear zones, where they are enveloped by foliation traces.

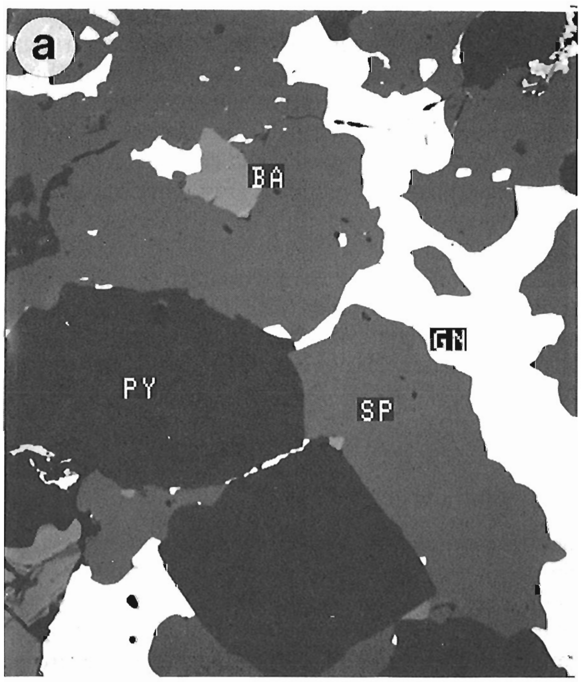
Petrography and paragenesis

Mineralization consists of veins and disseminations of pyrite, with lesser sphalerite, galena, magnetite and patchy chalcopyrite. Sulphide contents are as great as 40 % in some patches, but are commonly less than 5 %. Petrographic observations indicate that pyrite and magnetite are commonly the earliest phases, and are overgrown by sphalerite and galena (Fig. 6a). Locally, pyrite is brecciated and cemented by chalcopyrite. Sphalerite both overgrows and exhibits replacement by exsolved chalcopyrite. Quartz and barite, the principal gangue minerals, commonly overgrow or are intergrown with sulphides, although in a few samples cracks in barite are healed by galena.

Deformation is commonly most intense within the altered and mineralized zones. Sulphides in the veins are banded and these banded sulphides have been tightly folded where significant deformation occurred, features that may have been mistaken for slump textures by previous workers. In biotite schists, rotated pyrite grains display recrystallized tails and pressure shadows, are boudinaged, and record a dextral sense of shear (Fig. 6b). Rodding and elongation of pyrite grains and rock fragments is common.

No gold grains have been microscopically detected to date, although Franc de Ferriere (unpublished report, Falconbridge Nickel Mines Limited, 1978) identified gold in crosscutting veinlets containing pyrite, sphalerite, galena and rare chalcopyrite. A grain of silver sulphide (possibly argentite-acanthite), 12 microns (μ) in diameter, that is associated with magnetite and barite in a vein, was detected using a scanning electron microscope (SEM) (Fig. 7).

Though not abundant, fine (0.1 mm) grains of euhedral grossular garnet are fairly commonly found in association with sphalerite, pyrite and galena in veins in silicified rock. The garnets are hexagonal and have opaque cores with faintly birefringent rims (Fig. 8a). A vein, 1.2 cm wide, in a magnetite-rich host rock is lined with pseudo-hexagonal zoned grossular crystals. Other vein-filling minerals in highly silicified rock include wollastonite and tremolite (Fig. 8b). The latter exhibits pale blue to green pleochroism, contains sphene and quartz inclusions and is partially replaced in one instance by carbonate.



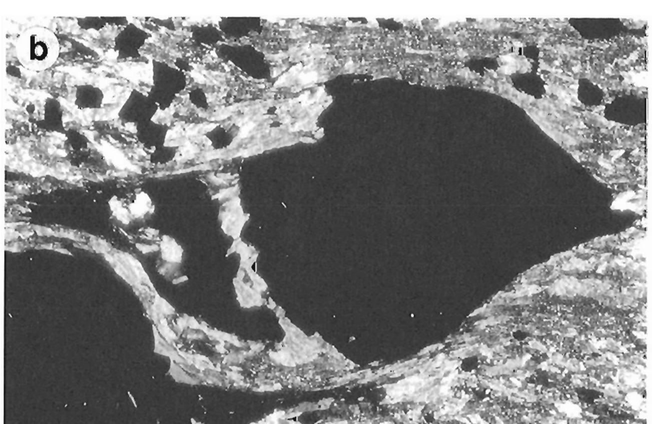


Figure 6. Sulphide textures in Handcamp samples.

a) SEM photograph illustrating overgrowth of early pyrite (PY) by sphalerite(SP), which is in turn overgrown by galena (GN) and barite (BA) (sample 85S698C(A)).

b) Recrystallized tail on pyrite grain, indicating a dextral sense of shear. Field of view 2.2 mm, crossed polars (sample HC-1-12).

Alteration extends into mafic volcanic rocks on the peripheries of the mineralized zone, but with generally decreased intensity. These less intensely altered rocks show extensive silica and lesser carbonate replacement of amphiboles, extensive sericitization of feldspars and massive epidote-chlorite replacement of the groundmass. Veins are filled by carbonate, epidote, magnetite and clinozoisite, which is commonly zoned (Fig. 8c). Host rock alteration is generally most intense around veins, and sulphides are uncommon except for minor disseminated pyrite.

Epidote alteration is common to all lithologies. It extends beyond the limits of the mineralized zone in an area depicted in Figure 3. In this area, epidote, with calcite and magnetite, forms stringers, boxwork patterns and patches mainly in mafic rocks (Fig. 8d), but also in felsic volcanics and red chert. Some of the epidote alteration appears to be later than the silicification and sulphide veining. Quartz-sericite-pyrite rocks and sulphide veins have been folded and sheared, whereas many epidote-calcite-magnetite veinlets form dendritic and boxwork networks and are undeformed. Some epidote veins, however, have been strung out parallel to the dominant foliation in both mafic and felsic volcanic rocks, especially near mylonite zones, suggesting that this type of calcium-rich hydrothermal alteration may have been in part syndeformational. Furthermore, the association of garnet, tremolite and wollastonite with sulphides in veins implies that they may have formed as a result of the same hydrothermal event. Given the common association of gold with sulphides, it is suggested that the gold mineralization is related to the same hydrothermal event that produced these sulphides and calcium metasomatism, and that this event was synkinematic to postkinematic. This is discussed in more detail below.

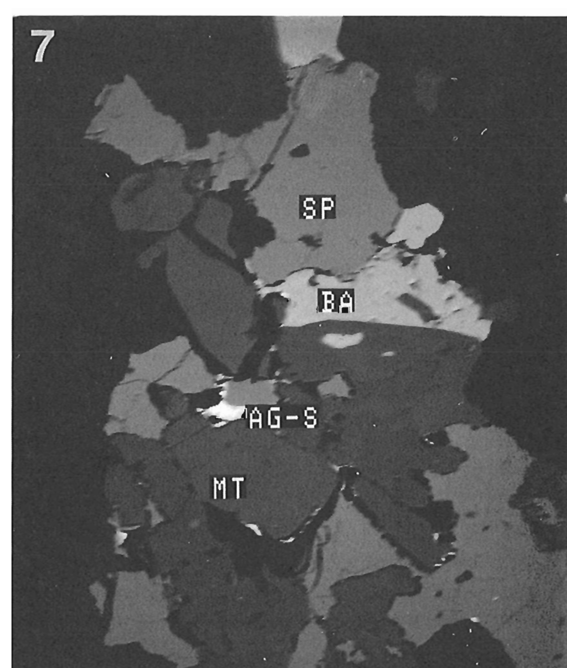


Figure 7. SEM photograph showing silver sulphide (AG-S) associated with magnetite (MT), barite (BA) and sphalerite (SP) (85S698F).

Late quartz and calcite veinlets commonly cut earlier veins and altered material, as well as foliation in some locations. They appear to record the last hydrothermal activity in the Handcamp area. It is not clear whether these are the products of the waning mineralization event or the result of later activity.

Table 1 summarizes the paragenesis of alteration and mineralization at Handcamp.

Metal contents

Twenty-one samples collected during the course of this study were assayed for gold and other base metals (Table 2). Gold values are highly variable, ranging from 1 to >10,000 ppb. Quartz-sericite-pyrite altered rocks that show the highest gold values (85S698C, 711D, 711E, 711F, 700B, 698E) are completely altered, and the only protolith that can be identified is an altered argillite (sample 85S698E).

Some relationships between gold and other metals in the deposit can be seen from inspection of Table 2 and from the X-Y plots in Figure 9. Samples with anomalous Cu, Zn and Pb are generally also high in Au, although this is not always the case. There is not a good one-to-one relationship between Au and any of the base metals. Au and Ag show a better correlation (Fig. 9a) and Au/Ag ratios are generally higher than is typical of volcanogenic sulphides in central Newfoundland. As and Sb show the best correlation with Au (Fig. 9b,c), and some anomalously high Mo values also show an admittedly diffuse positive relationship with Au (Fig. 9d).

DISCUSSION

The field, petrographical and geochemical data presented in the preceding sections may offer some clues concerning the genesis of the Handcamp prospect.

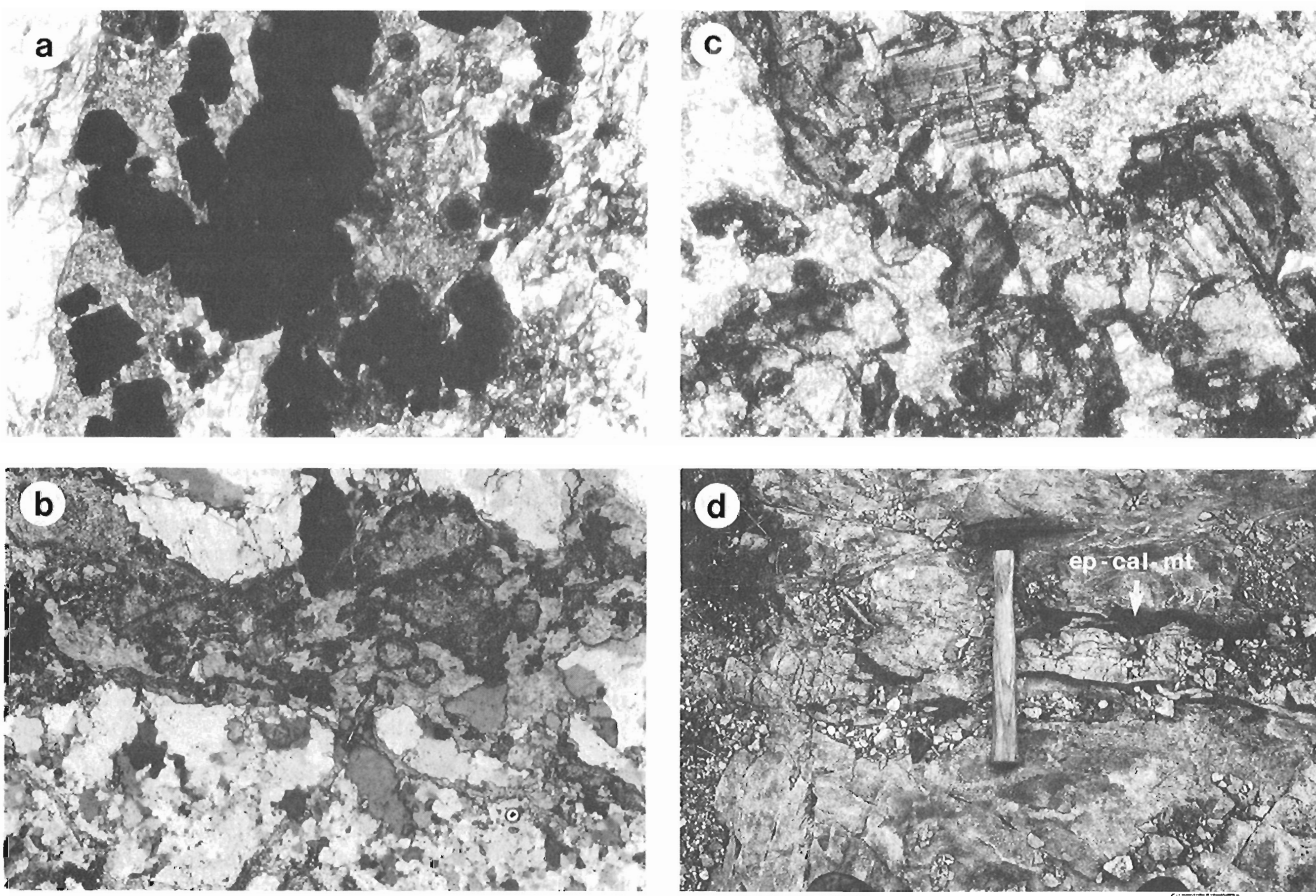


Figure 8. Photomicrographs and photograph of Ca-Fe-Mg minerals in Handcamp samples.

a) Grossular garnets associated with sulphides in a vein in a quartz-sericite -pyrite rock. Garnets are hexagonal, and have opaque cores. Field of view 2.2 mm, partially crossed polars (85S711E-1).

b) Tremolite-chlorite-epidote-garnet veinlet in quartz-sericite rock. Field of view 2.2 mm, partially crossed polars (HC-4-1).

c) Zoned clinozoisite crystals in a vein with calcite. Field of view 2.2 mm, partially crossed polars (HC-2-1).

d) Epidote-calcite-magnetite vein in pillow basalt.

Table 1. Paragenetic sequence of hydrothermal events at Handcamp.

Mineral	Stage 1	Stage 2	d	Stage 3	Stage 4
q	———	———			———
py	———	———			
sc	———				
mt	——	———		———	
sp		———			
gn		———			
cpy		———			
ba		———			
Au-Ag		???????		???????	
ep		——		———	
cal				———	———
gar		——		——	
cz				———	
trem		——		——	
woll					———

Symbols/Abbreviations: q: quartz: py: pyrite: sc: sericite: mt: magnetite; sp: sphalerite; gn: galena; cpy: chalcopyrite; ba: barite; Au-Ag: gold-silver; ep: epidote; cal: calcite; gar: grossular garnet; cz: clinozoisite; trem: tremolite; woll: wollastonite; d: deformation.

Metal contents

Although based on limited data, the metal associations in the Handcamp prospect suggest a nonvolcanogenic origin. Kerrich (1983) has pointed out that in Archean terranes, the association of gold with a suite of rare metals that includes As and Sb is characteristic of lode gold deposits as opposed to massive sulphides. The Handcamp prospect's positive correlation of Au with As and Sb (and to a lesser extent with Mo), coupled with a less evident Au-base metal correlation, invites a comparison with structurally-controlled Au environments elsewhere in Newfoundland. For example, in the Little River area north of Bay d'Espoir, significant Au values are associated with stibnite and arsenopyrite (Tuach et al., 1988) and in the Baie Verte Peninsula and western White Bay, auriferous mineralization associated with major Silurian structures is reflected geochemically in lake sediments by strong linear As and Sb anomalies (Davenport and Nolan, 1987; Tuach et al., 1988).

Structural controls

The "mineralized horizon" was thought by previous workers to be a stratigraphic interval of mainly exhalative mineralization hosted by cherts and felsic and mafic tuffs. However, field relationships suggest that it is better described as a strata-parallel zone of intense deformation and shearing, and good arguments can be made that the alteration and mineralization are related to the structural events, rather than to a previous volcanogenic mineralization episode. Close examination of this zone reveals that many of the quartz-sericite-pyrite rocks are fragments in extremely sharp contact with unaltered lithologies and are enveloped by the foliation in polylithic shear and mylonite zones. Although some of the altered rock appears to have escaped the deformation, pyrite grains are deformed and sulphide-barite-grossular garnet-epidote-magnetite veinlets are tightly folded and have been subjected to dextral shear. The implication is that the alteration and mineralization were synkinematic to postkinematic.

The age of the deformation (and by implication, possibly the mineralization) is not well constrained. The main schistosity is known to affect fossiliferous Lower Silurian rocks of the Sansom Greywacke elsewhere in the area and therefore is probably at least this young. Fifteen kilometres to the south of the Gullbridge area, shearing, which is probably of the same generation, is overgrown by metamorphic minerals related to the intrusion of the Twin Lakes Complex (Upadhyay and Smitheringale, 1972). Although not dated, this complex cuts probable Middle to Upper Silurian rocks of the Charles Lake sequence north of Grand Falls (Swinden, 1988b) and is therefore probably Upper Silurian to Devonian. It is suggested on this basis that the deformation, and possibly the mineralization, is post-Middle Silurian.

Ca-rich minerals and gold mineralization

Ca-Mg-Fe silicate and carbonate minerals (epidote, wollastonite, calcite, grossularite, clinozoisite, tremolite) are ubiquitous at Handcamp. The abundance of carbonate minerals in the gangue was cited by Kerrich (1983) as characteristic of lode gold rather than massive sulphide deposits in Archean sequences. Although most of these minerals could have formed during hydrothermal alteration at generally accepted temperatures, wollastonite and grossularite are mainly high-temperature contact-metamorphic minerals which are commonly associated with igneous intrusions.

The only other location in the entire Robert's Arm belt at which Ca-Mg-Fe minerals have been documented is at Gullbridge (Upadhyay, 1970), where elongated amygdules in basalts are filled with vesuvianite, sphene, calcite, diopside and possibly grossular garnet. Upadhyay (1970) considered that these may have formed in response to the intrusion of Devonian granite-diorite bodies, which also formed the cordierite-anthophyllite assemblages there, by metamorphism of carbonate minerals in the amygdules. Similarly, the Ca-Fe-Mg minerals at Handcamp could have been formed as a result of metamorphism (due to the intrusion of the Twin Lakes Complex, (Figure 2) of earlier Ca-rich minerals generated by Ca-metasomatism associated with gold mineralization.

The reports of Franc de Ferriere (unpublished report, Falconbridge Nickel Mines Limited, 1978) and Burton (unpublished report, US Borax, 1981) emphasized the association between high gold content and pyritized argillites. Analyses from this current study show that quartz-sericite-pyrite rocks, originally argillites, were selectively enriched in gold. Burton (unpublished report, US Borax, 1981) also suggested that these host rocks would have provided a suitable reducing environment for gold precipitation. Boyle (1979) has discussed several mechanisms of gold precipitation and has pointed out that the reduction of gold-bearing solutions by sulphides and pyritiferous carbonaceous schists, resulting in the precipitation of gold, is a common phenomenon. It is possible then, that pyritized argillite and sulphides at Handcamp provided suitable sites for gold precipitation.

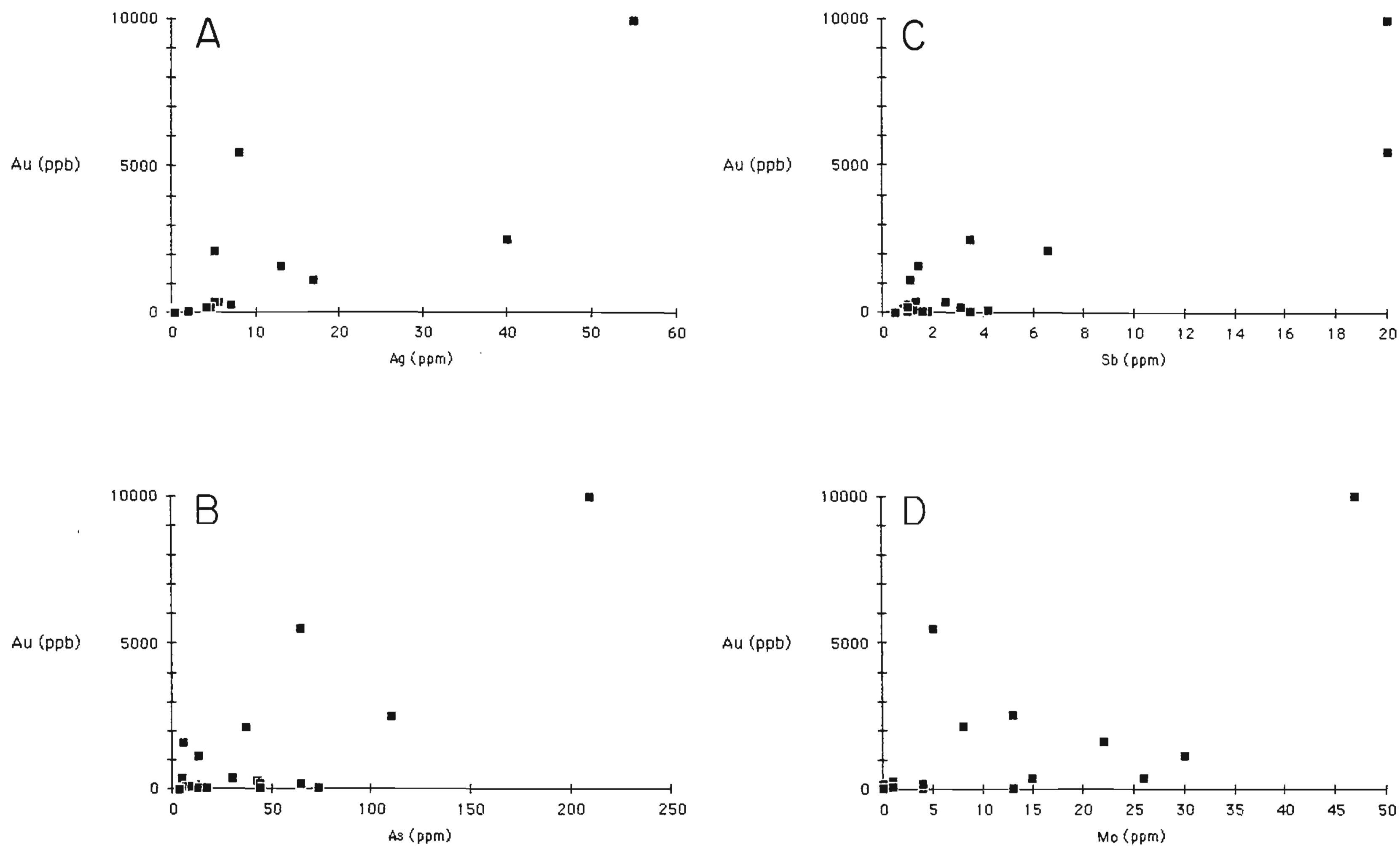


Figure 9. X-Y plots for Handcamp samples. Data used can be found in Table 2.

Table 2. Analyses and lithologies of Handcamp samples

SAMPLE	Cu ppm	Zn ppm	Pb ppm	Ag ppm	Au ppb	Co ppm	Cd ppm	Ba ppm	As ppm	Sb ppm	Mo ppm	Lithology
85S698A	48	>4000	46	<0.5	110	10	11	1290	12	0.8	1	Silicified felsic tuff, 1-2% py
85S698B	51	92	36	7.0	280	8	0	11100	43	1	1	Silicified felsic tuff, 25-30% py
85S698C	>4000	>4000	>4000	40.0	2500	19	140	242000	110	3.5	13	Red chert with cpy, gn, sp, mt, ba veins
85S698D	790	>4000	680	5.5	340	5	15	45900	4.8	1.3	26	Red chert with py, mt, ch, sc veinlets
85S698E	540	2400	1000	13.0	1600	5	9	19600	5.8	1.4	22	Red chert with sulphide veinlets
85S698F	13	150	94	<0.5	2	25	0	2180	4	0.9	0	Silicified black argillite with sc/sulphide veins
85S699A	190	870	540	5.0	340	16	3	7850	30	2.5	15	q/sc/py rock with sp veinlets
85S700A	55	1400	120	4.5	190	38	5	24700	44	3.1	0	q/sc/py rock (altered felsic tuff)
85S700B	1200	160	130	17.0	1100	5	0	3260	13	1.1	30	q/sc/py rock with minor epidote
85S701	47	350	96	<0.5	62	25	0	2350	6.8	1.3	1	Black argillite with disseminated py
85S709B	29	13	38	<0.5	62	16	0	20	10	4.2	1	Epidotized basalt with disseminated mt and sp veins
85S709C	60	51	60	<0.5	31	54	0	100	44	3.5	13	Banded ep/mt rock
85S710	91	28	40	<0.5	12	14	0	640	13	1.8	0	Silicified black argillite
85S711A	7	30	16	<0.5	12	6	0	6540	17	1.6	0	Silicified argillite with py, woll, gn veins
85S711D	1200	>4000	>4000	55.0	>10000	16	45	25500	210	20	47	q/sc/py rock with minor garnets
85S711E	330	1900	200	8.0	5500	14	1	182000	65	20	5	q/sc/py rock with mt stringers and garnets
85S711F	29	120	90	5.0	2100	4	1	11700	37	6.6	8	q/sc/py rock
86S292A	48	120	<2	>0.5	1	230	0	110	3.6	0.5	4	Mafic volcanic with ep, cal, q veins
86S293A	350	5600	740	2.0	34	100	19	380	74	1	4	Mafic volcanic with ep, cal, q veins
86S293B	53	210	12	4.0	150	180	0	320	65	1	4	Silicified mt-rich chert

Symbols/Abbreviations: q:quartz; py:pyrite; cpy:chalcopyrite; gn:galena; sp:sphalerite; mt:magnetite; ba:barite; ch:chlorite; sc:sericite ep:epidote; cal:calcite; woll:wollastonite

SUMMARY

Our field and laboratory studies suggest that, contrary to the models of some previous workers, the Handcamp prospect is probably not an exhalative volcanogenic sulphide deposit. Field mapping has shown that the mineralization occupies a major strata-parallel shear zone approximately 50 m wide, in which alteration and sulphide mineralization appear to be both synkinematic and postkinematic. The mineralized zone consists mainly of highly deformed felsic volcanic and argillaceous sedimentary rocks which have been variably altered to a quartz-sericite-pyrite-magnetite assemblage. Abundant Ca-Fe-Mg carbonates and silicates suggest that CO_2-rich fluids played an important part in the mineralizing process and invite comparisons with lode gold, rather than massive sulphide, environments. Similarly, Au enrichment is paralleled by Sb, As and locally Mo, more typical of structurally-controlled rather than volcanogenic Au deposition in central Newfoundland and elsewhere.

Our preliminary interpretation is that the Handcamp deposit may have formed through shear-related hydrothermal activity during the Middle to Late Silurian or Devonian. A similar mineralizing episode has recently been identified in the Victoria Lake Group to the southeast, where probable Silurian auriferous alteration zones are spatially, but not genetically, related to Late Cambrian massive sulphide deposits (Kean and Evans, 1988; Tuach et al., 1988). If true, this would be the first deposit of this type to be identified in the Buchans-Robert's Arm Group belt and suggests that a re-examination of some other "volcanogenic" alteration zones for evidence of auriferous shear-related mineralization may be warranted. Clearly, additional evidence is needed, and further geochemical, petrographical, mineralogical and isotopic work is in progress.

ACKNOWLEDGMENTS

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Aeromagnetic total field, gradiometer, and VLF-EM survey of an area southwest of Bathurst, New Brunswick[1]

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Geophysics Division

Ellis, B., Teskey, D.J., and Ready, E.E., Aeromagnetic total field, gradiometer, and VLF-EM survey of an area southwest of Bathurst, New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 107-113, 1989.

Abstract

A combined aeromagnetic total field, gradiometer and airborne VLF-EM survey of an area southwest of Bathurst, New Brunswick was carried out at the end of 1986 and beginning of 1987. Available information includes aeromagnetic anomaly maps, aeromagnetic vertical gradient maps and VLF-EM profiles.

Résumé

Un levé combiné aéromagnétique du champ total, gradiométrique et électromagnétique à très basse fréquence (VLF-EM) aérien couvrant une région au sud-ouest de Bathurst, au Nouveau-Brunswick, a été réalisé à la fin de 1986 et au début de 1987. L'information disponible comprend les cartes des anomalies aéromagnétiques, les cartes du gradient vertical aéromagnétique et les profils VLF-EM.

[1] Contribution to the Canada — New Brunswick Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada.

INTRODUCTION

A combined aeromagnetic total field, gradiometer and airborne VLF-EM survey of an area southwest of Bathurst, New Brunswick was flown under contract to the Geological Survey of Canada as a contribution to the Canada — New Brunswick Mineral Development Agreement(1984-1989). The location of the survey area is shown in Figure 1. The contract was awarded to Geophysical Surveys Inc. of Quebec City and the survey was flown in the period October 12, 1986 to April 12, 1987. The gradient field was measured by two Cesium vapour magnetometers, each with a resolution of 0.005 nanoteslas, separated by 2 m and suspended 30 m below a helicopter. A mean elevation clearance of 150 m was maintained with a flight line separation of 300 m. VLF-EM total field and quadrature, using signals from Cutler, Maine and Annapolis, Maryland, were recorded with a Herz Totem-2a receiver mounted on the helicopter.

Survey output is in the form of 1:20 000 aeromagnetic total field and gradiometer contour maps and 1:50 000 magnetic anomaly (Fig. 2) and gradiometer colour interval maps (Fig. 3). Figures 2 and 3 are black and white reduced photos of colour maps that can be obtained either from the New Brunswick Department of Natural Resources or the Geological Survey of Canada. The VLF-EM total field and quadrature profiles are printed on the back of the magnetic anomaly and gradiometer colour interval maps respectively in such a way that they can be viewed jointly with the magnetics data on a light table. In addition, the digital data are available for purchase from the Geophysical Data Centre, Geological Survey of Canada for further processing and interpretation.

In the survey area, the magnetic anomaly patterns can be correlated with the mapped geology as published by Davies (1979) (Fig. 4), and can be used to extend the magnetic units under sedimentary cover and to refine geological boundaries. The gradiometer trace is particularly effective for the latter purpose, as the zero gradiometer contour tends to lie directly above a magnetic contact in northern latitudes. For New Brunswick, this contour will be slightly offset due to the inclination of the Earth's field, the effect of remanent magnetization and the interference from adjacent sources. VLF-EM anomalies are associated with both contrasts in conductivity between rock units and cultural features.

The units mapped in the survey area include Devonian sediments, mafic and felsic volcanics, granites, gabbros and diabase and Ordovician sediments of the Tetagouche Group (Davies, 1979). Magnetic anomalies are associated with the Devonian volcanics, granites, gabbros and mafic volcanics.

The volcanics and granites can be identified by characteristic magnetic patterns. Analysis of these patterns indicates that the boundaries of the units can be modified considerably from those shown on the published geological map. Detailed analysis of the magnetic and VLF-EM anomalies should prove to be extremely useful for enhancing the mapping and understanding of the complex geology in this area.

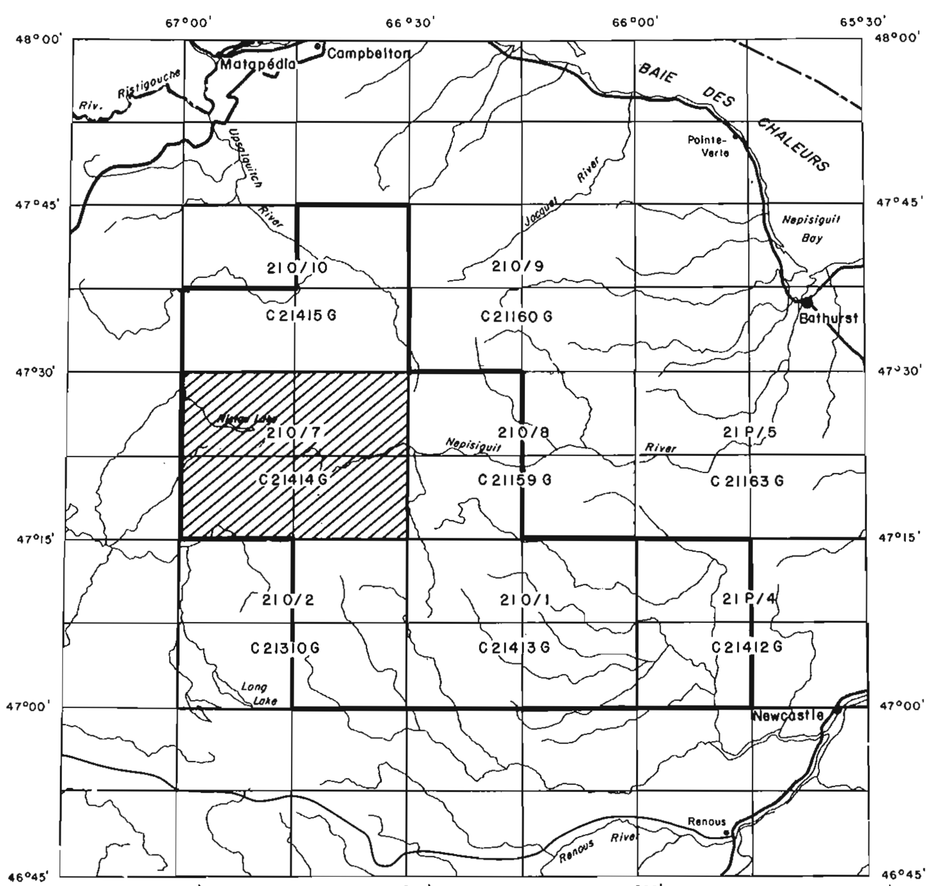

Figure 1. Location area map of the aeromagnetic, gradiometer and VLF-EM survey near Bathurst, New Brunswick, 1986/1987.

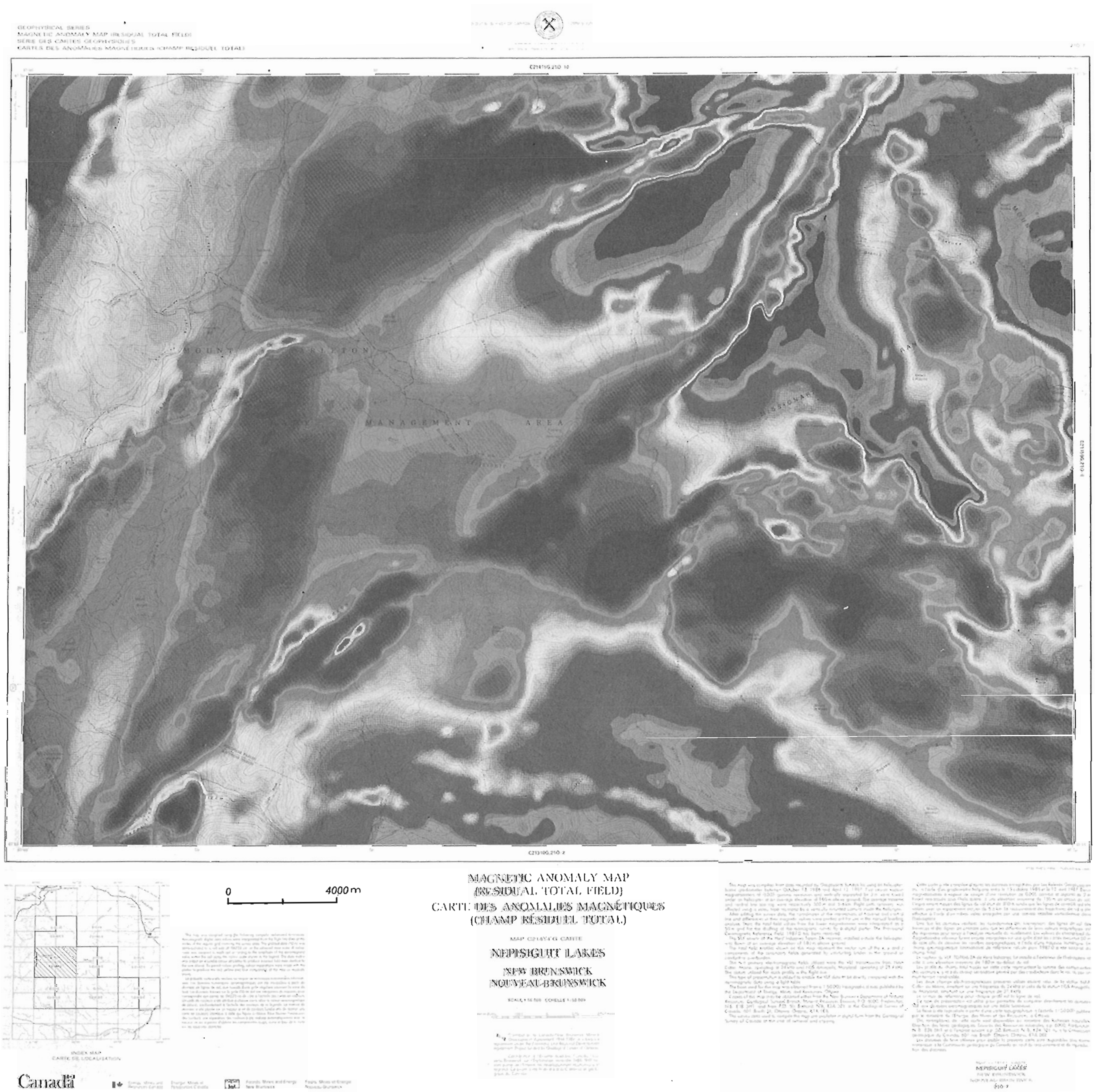

Figure 2. Black and white copy of an original colour magnetic anomaly map (residual total field) available from the Geological Survey of Canada. Map 21O/7.

Figure 3. Black and white copy of an original colour aeromagnetic vertical gradient map available from the Geological Survey of Canada. Map 21O/7.

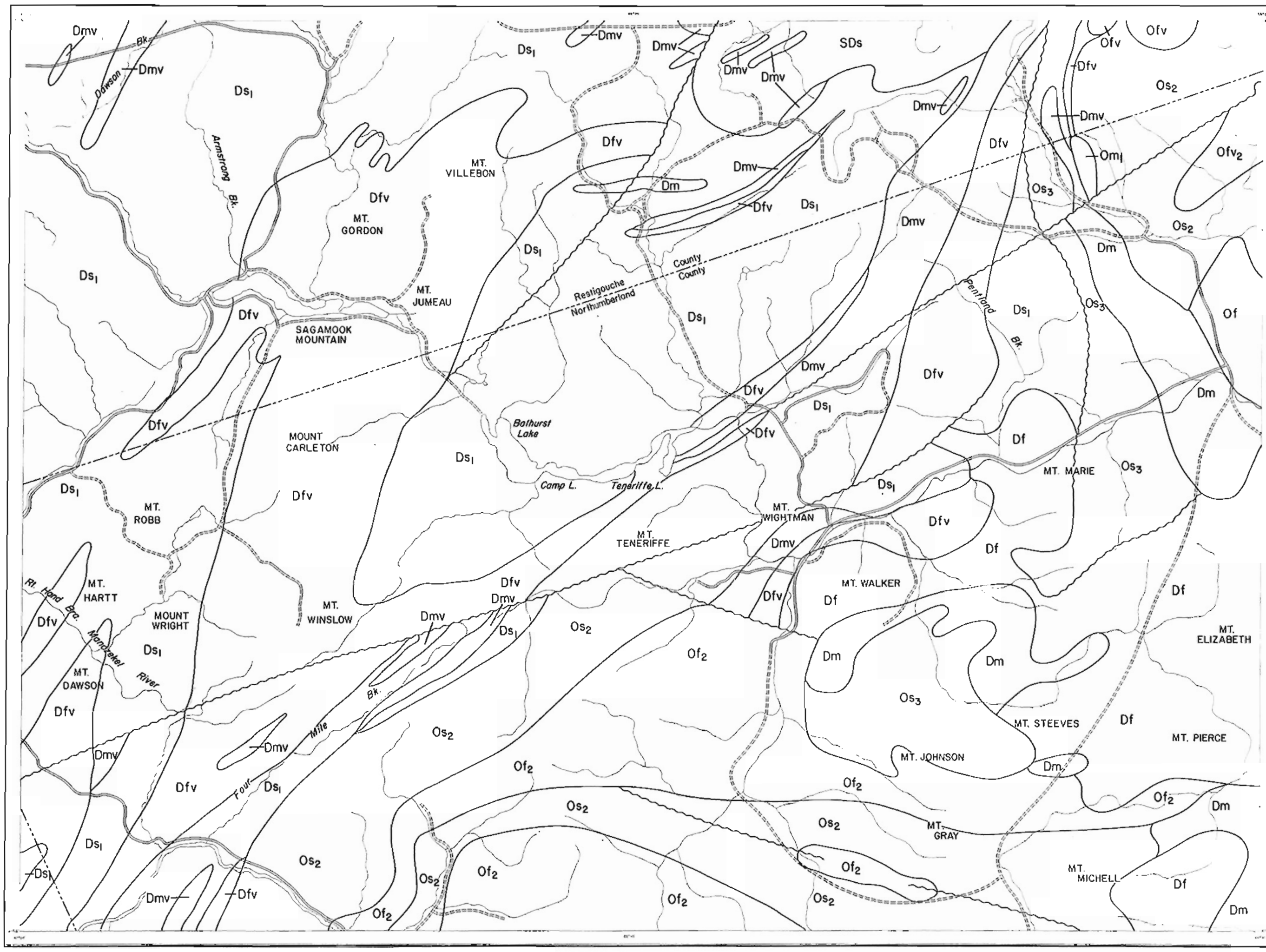

Figure 4. Section of the geological map of northern New Brunswick from Davies (1979).

LEGEND

CHALEUR AND TOBIQUE ZONES

LOWER DEVONIAN
Dalhousie Group
- Dfv — maroon and orange flow-banded and massive rhyolite, rhyolite agglomerate, tuff and breccia, dacite(?)
- Dmv — amygdaloidal basalt, basaltic tuff and breccia, palagonite tuff, andesite, minor shale, mudstone and siltstone.
- Ds1 — calcareous mudstone, siltstone, sandstone, maroon and green sandstone, siltstone, conglomerate, limestone. Includes minor felsic and mafic volcanic rock.

MIRAMICHI ZONE

DEVONIAN
- Df — granite adamellite, granodiorite, quartz monzonite, quartz feldspar porphyry and related rocks
- Dm — gabbro and diabase

ORDOVICIAN
- Of2 — gneissic and cataclastic granite
- Of1 — rhyolite metaporphyry, quartz feldspar metaporphyry.
- Om1 — metagabbro and metadiabase

ORDOVICIAN AND OLDER (?)
Tetagouche Group
- Os3 — dark grey phyllite, graphitic slate, red and green manganiferous slate and chert, feldspathic lithic and quartzose greywacke and iron formation, minor limestone and conglomerate. (Caradocian graptolites occur in this unit near the mouth of the Tetagouche River).
- Omv2— metabasalt, pillow metabasalt, basaltic metatuff, minor metatrachyte.
- Ofv2 — quartz and quartz feldspar metaporphyry, quartz sericite schist, quartz-chlorite sericite schist, crystal metatuff.
- Os2 — grey phyllite, metaquartzite, metagreywacke, minor limestone, graphitic schist, hornfels. (Arenigian brachiopods occur in uppermost strata on Tetagouche River).

CORRELATION OF MAGNETICS AND GEOLOGY — NEPISIGUIT LAKES (21O/7)

Map 21O/7, one of six aeromagnetic maps produced at 1:50 000 covering the survey area, has been selected to highlight the correlation between magnetics and geology.

The main geological feature of this map area is the tectonic boundary between the Chaleur Bay and Miramichi tectonic zones (Fig. 4). The southeastern section of the map lies on the western flank of the Miramichi Anticlinorium, a belt of highly deformed Ordovician volcanic, sedimentary and intrusive rocks. The northwestern section of the map lies on the eastern flank of the Chaleur Bay Synclinorium. These two major tectonic zones contain base metal, gold and silver occurrences (Skinner, 1982).

MIRAMICHI ANTICLINORIUM

This southeastern part of the map is underlain by Ordovician (Tetagouche Group) and Devonian rocks, most of them granite related.

The magnetic bodies indicate more fragmentation and faults than in the northwestern section of the map (Fig. 5).

The general pattern of truncation is northwest-southeast and northeast-southwest within the Devonian units. The fault identified as A, A_1 is shown on the geological map (Fig. 4); however, the magnetic patterns suggest that it continues to the southeast, is offset slightly to the northeast, then continues as indicated by B,B_1 (Fig. 5). Another similar fault, marked as D, D_1, trends in an northwest-southeast direction parallel to A,A, and B, B,. The fault identified as G, is an example of a shorter fault trending in a northeast-southwest direction, further fragmenting the magnetic bodies. VLF-EM anomalous peaks appear to trend along these faults suggesting that the faults are conductive and correlate with magnetic contacts.

The Ordovician units correlate with weaker magnetic anomalies however, VLF-EM anomalies indicate east-northeast trending conductors (E, Fig. 5). These are contacts not shown by the gradiometer.

CHALEUR BAY SYNCLINORIUM

This northwestern half of the map is underlain by Devonian (Dalhousie Group) sedimentary and volcanic rocks of the Tobique-Chaleur Subzone, which are host to all of the recently discovered gold occurrences (Burton, et al., 1987).

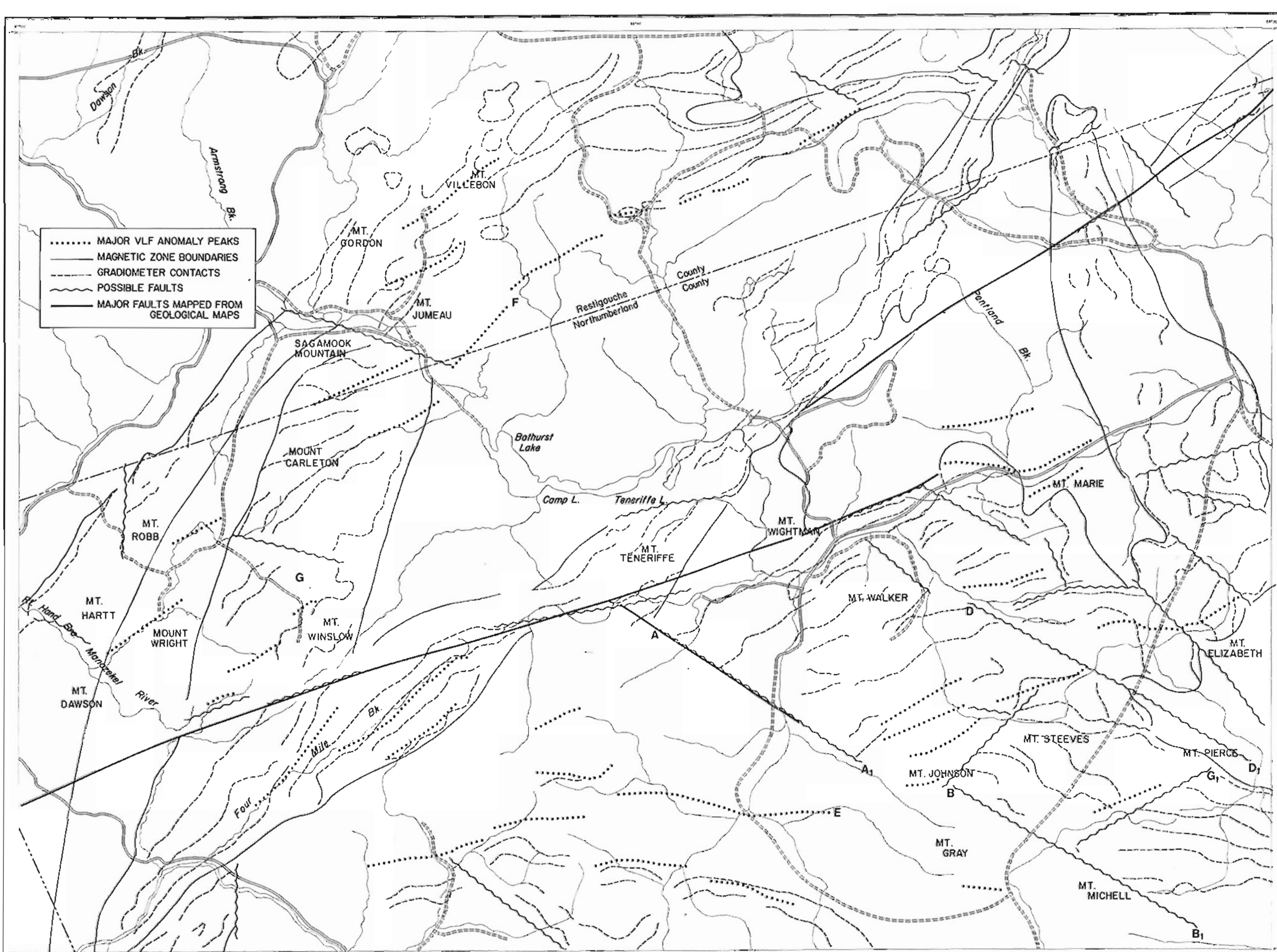

Figure 5. Geophysical interpretation of Map 21O/7.

The general magnetic trend is northeast-southwest, transected by a number of major northeast trending faults (Fig. 5). Identifier G (Fig. 5) is a good example of how magnetic contacts can delineate geological zones.

The sedimentary Ds1 unit correlates with broad magnetic lows and little or no activity on the VLF-EM profiles, except for a well-marked conductor (F, Fig. 5), trending in a northeast-southwest direction. This is another example of a contact not shown by the gradiometer but with a good VLF-EM anomaly.

The felsic and mafic volcanic rocks (Dfv, Dmv) have strong magnetic signatures; the most intense correlate with the mafic volcanics. The vertical gradient map shows that some of these units extend under sedimentary cover enabling the redefinition of the geological map boundaries.

DISCUSSION

This type of survey demonstrates the utility of gradiometer and VLF-EM for mineral exploration and geological mapping. Strong gradiometer signatures correlate very well with the mapped volcanics and show their multiple extension beneath sedimentary cover.

The combined gradiometer and VLF-EM data are a useful tool for delineating faults and contacts. Some contacts shown on the gradiometer maps show no response in the VLF-EM profiles and vice versa.

Burton et al. (1987) suggests that gold mineralization, in this environment may be volcanoplutonic or fault-controlled. If it has a volcanoplutonic origin, then the level of erosion may be favourable for deeper mineralization. If the mineralization is simply associated with selective alteration of intrusive bodies along faults, the potential is still high for a disseminated type deposit as described by Boyle (1979).

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Basin configuration, sedimentary facies, and resource potential of the Lower Carboniferous Horton Group, Cape Breton Island, Nova Scotia[1]

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Hamblin, A.P., Basin configuration, sedimentary facies, and resource potential of the Lower Carboniferous Horton Group, Cape Breton Island, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 115-120, 1989.

Abstract

Deposition of the Horton Group in the study area occurred in two fault-bounded extensional sub-basins (approximately 100 × 40 km²), interpreted as adjacent half-grabens with opposite polarities of asymmetry. Both were probably part of a more regional linear tectonic system which underwent a similar tectonostratigraphic history along its length. The lower Craignish megafacies comprises facies assemblages including red or grey alluvial fan braidplain sandstone and conglomerate and red mudflat-playa siltstone. The middle Strathlorne megafacies comprises facies assemblages including grey or green basin-centre open lacustrine mudstone, prograding shoreline red and grey fine sandstone and fault margin sandstone and conglomerate. The upper Ainslie megafacies comprises facies assemblages including red and grey fault margin conglomerate, red fluvial sandstone and basin centre fluvial sandstone and siltstone with longitudinal dispersal. Understanding the organization and geometry of facies assemblages within these half-graben sub-basins is important in predicting resource potential.

Résumé

Dans la région étudiée, la sédimentation du groupe de Horton s'est produite dans deux sous-bassins de distension limités par des failles (environ 100 x 40 km), et considérés comme des demi-grabens adjacents caractérisés par des polarités d'asymétrie opposées. Ces deux sous-bassins faisaient probablement partie d'un système tectonique linéaire de caractère régional qui aurait subi une évolution tectonostratigraphique similaire sur toute sa longueur. Le mégafaciès inférieur de Craignish comprend trois associations de faciès, notamment un grès rouge ou gris de plaine anastomosée construite sur un cône alluvial, un conglomérat, et un microgrès rouge de playa et slikke. Le mégafaciès intermédiaire de Strathorne comprend quatre associations de faciès, notamment au centre du bassin, une pélite grise ou verte mise en place dans un milieu ouvert de type lacustre, un grès à grain fin rouge et gris de ligne de rivage progradante, et un grès et conglomérat mis en place en marge d'une faille. Le mégafaciès supérieur d'Ainslie comprend trois associations de faciès, notamment un conglomérat rouge et gris mis en place en marge d'une faille, un grès fluviatile rouge, et un grès et microgrès fluvialites mis en place au centre du bassin et caractérisés par une dispersion longitudinale. Il est important de comprendre l'organisation et la géométrie des associations de faciès à l'intérieur de ces sous-bassins de type demi-graben pour mieux prévoir leur potentiel en ressources.

[1] Contribution to the Canada-Nova Scotia Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada, Lithosphere and Canadian Shield Division.
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INTRODUCTION

This progress report summarizes the results of the second field season investigating the sedimentology of the Horton Group, primarily on northern Cape Breton Island. Horton sediments comprise post-Acadian basal nonmarine clastics beneath the Windsor Group and were deposited in fault-bounded extensional basins in western and northern Cape Breton Island. A brief summary of project objectives, previous work, stratigraphic and structural setting and the first year's work were provided by Hamblin (1988). The main purpose of the study is to use facies distributions to interpret the geometry and tectonic style of Horton depositional basins.

Fieldwork was concentrated in areas of northern Cape Breton Island surrounding the Cape Breton Highlands (latitude 46°00′ to 47°15′, longitude 60°15′ to 61°15′) (Fig. 1). Horton outcrop areas occur as isolated pockets preserved around the basement complex of the Cape Breton Highlands. The lower and middle parts of the Horton Group are well represented in northern Cape Breton, whereas in western Cape Breton, the middle and upper Horton Group are best exposed. Surface data are provided by equal proportions of coastal and stream outcrops, but outcrop quality varies from poor to excellent. Most sections have some structural complications and expose only portions of the Horton Group. In addition to the previous work outlined by Hamblin (1988), the geological maps of Cameron (1948), Neale and Kelley (1960) and Neale (1963a,b, 1964a,b) were relied on throughout.

A total of 11 500 m of strata were measured in 49 outcrop sections in 1988, most in northern Cape Breton Island. In addition, lithological data from 8 onshore and 2 offshore exploration wells were assembled into measured sections. The complete Horton data base now includes 24 000 m of strata in 66 measured sections (Fig. 1). A total of 386 paleocurrent measurements were made giving a total of 930 measurements for both seasons. Palynological data from 58 samples will aid in the correlation and interpretation of this complex depositional system.

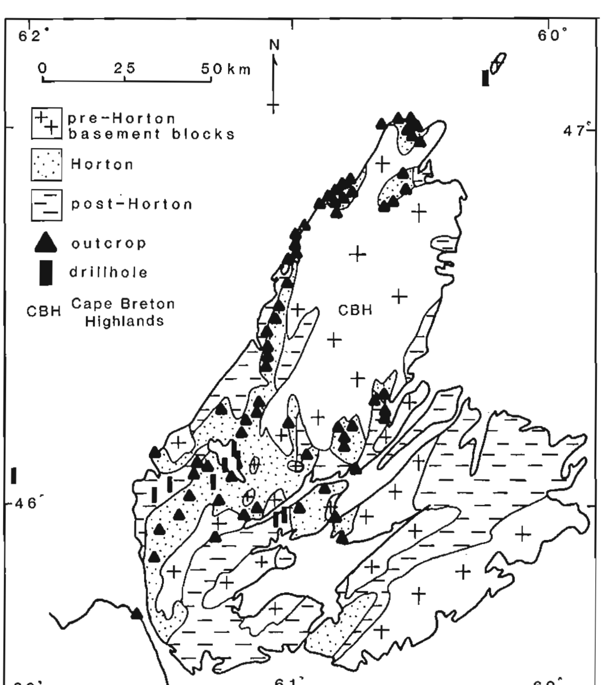

Figure 1. Simplified geological map of Cape Breton Island, locating measured sections and drillholes.

POSTDEPOSITIONAL STRUCTURE

All structural features observed in outcrop and core were postdepositional. The most likely timing for these features is the post-Canso Group, pre-Riversdale Group Alleghanian Orogeny, a regional compressional tectonic phase, which obliterated Horton-age basin margin faults. Faults are abundant in northern Cape Breton Island and break the near-continuous Horton coastal outcrop into a series of small blocks with limited defineable sedimentological connection to each other. Nowhere in northern Cape Breton Island is there a complete unfaulted section of the Horton Group.

Many topographically high basement blocks, now present at surface, separate low-lying Horton outcrop areas. These blocks have high-angle fault boundaries of post-Horton age and do not affect facies or paleocurrent patterns in the surrounding Horton sediments. These blocks were probably not positive features during deposition, contrary to some earlier interpretations, which has led to a misconception that the Horton Group was deposited in numerous small sub-basins. Repetition of sections by thrust faults is observable in outcrop and drillholes in western Cape Breton Island and has led to the erroneous attribution of large thicknesses to the Horton Group in that area. On a larger scale it appears that this (Alleghanian?) thrusting has led to the northwestward offset of much of the southeastern Horton basin margin in the western Cape Breton area (Fig. 2). This complicates sedimentological interpretations considerably and is important for interpretations of resource potential. Abundant folding also required careful section measuring and is important in resource evaluation.

THE DEPOSITIONAL SUB-BASINS

Sedimentological, stratigraphic and paleocurrent data suggest that deposition of the Horton Group was in two large fault-bounded asymmetrical sub-basins separated by the Cape Breton Highlands basement block (Fig. 2, 3). Thick sections of coarse grained, poorly sorted sediments (facies assemblages C1, S3, A1) inferred to be alluvial fan and braidplain deposits (see below) are interpreted to imply that most sub-basin margins were fault-bounded, as suggested by Belt (1968). Paleocurrent data from these facies are interpreted to represent transverse flow away from those margins. Finer grained, well sorted sediments (facies assemblages C2, S1, S2, A3) inferred to be lacustrine and fluvial deposits (see below) are interpreted to indicate deposition near unfaulted margins or near sub-basin axes.

However, the distribution of these facies and paleocurrent data is not symmetrical, suggesting deposition in half-grabens (Fig. 3). In the western sub-basin, interpreted fault margin facies are concentrated on the northern, eastern and southeastern sides while unfaulted margin facies are more prevalent on the western side. Paleocurrent data indicate sediment input from all sides. The finest grained lacustrine facies (facies assemblage S1) is thickest near the eastern fault margin side and all associated paleocurrent data indicate flow toward that margin, i.e. the basinal paleoslope dipped toward an axial locus of subsidence near the main bounding fault (Fig. 3). In the northern sub-basin, the same asymmetry is developed but in this case it is the western and northwestern margins that appear to be the main fault margins while the southeastern margin is dominated by facies interpreted to represent a margin with limited faulting.

The two sub-basins, which have opposing asymmetry of facies distribution, were apparently separated by the main Cape Breton Highlands basement block, which supplied coarse detritus to both. In addition, the dominant paleocurrent flow direction in the basinal lacustrine facies of each sub-basin indicates different, nearly opposite, paleoslopes (Fig. 3). The opposing asymmetry of facies distribution and paleocurrent data suggest that the two half-grabens may represent en- echelon rift segments with opposite polarity of asymmetry, analogous to those described from the East African Rift by Rosendahl et al. (1986) and Frostick and Reid (1987).

HORTON MEGAFACIES AND FACIES ASSEMBLAGES

The tripartite stratigraphic scheme described in the literature can be recognized throughout western and northern Cape Breton Island. The lower Craignish, middle Strathlorne and upper Ainslie formations (Murray, 1960) correspond to diachronous megafacies that always occur in the same stratigraphic order. They represent tectonically-controlled depositional settings arranged spatially in the fault-bounded sub-basins of deposition, but also record a specific temporal evolution of these sub-basins. Therefore it is of prime importance to view each outcrop in terms of its vertical position in the Horton Group and its lateral position within the sub-basin relative to original fault margins. These megafacies record an overall fining-upward trend, followed by a coarsening-upward trend. The three megafacies can be further subdivided into ten facies assemblages recognizable in both sub-basins (Table 1).

The Crainish megafacies

This megafacies up to 2000 m thick, consists of red or grey siltstone to conglomerate but is typically thick-bedded pebbly medium to coarse grained sandstone with thin finer interbeds. It conformably overlies the Fisset Brook Formation volcanics, but generally lies unconformably on basement. It thins toward basin margins apparently by pinching out from the base. It generally represents sedimentation on alluvial fans and braidplains in a broad slowly subsiding basin. Three facies assemblages are recognized, designated C1, C2 and C3.

C1) Red/orange coarse sandstone to conglomerate

This facies assemblage is most prevalent in the lower and upper Craignish near original basin margins. The main lithofacies is red to orange-grey arkosic, pebbly medium to

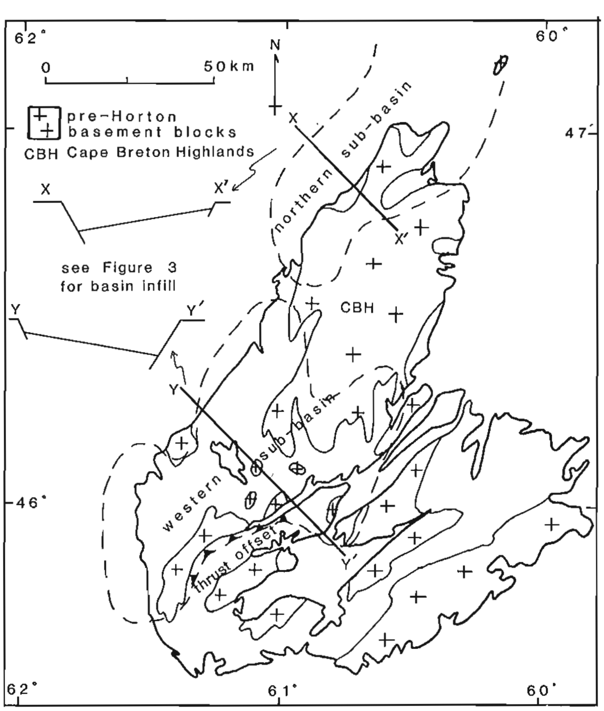

Figure 2. Horton depositional sub-basins and simplified cross-sections.

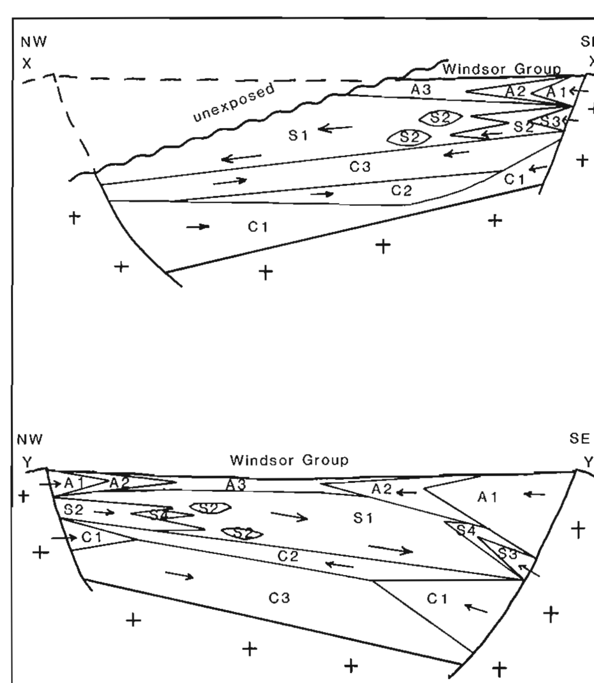

Figure 3. Simplified distribution of facies assemblages as defined in text and Table 1. Cross section locations given in Figure 2. Arrows indicate generalized paleoflow.

coarse grained sandstone and matrix-supported sandy conglomerate in sharp-based fining-upward beds with broad trough crossbedding. Red sandy siltstone to medium grained sandstone with crossbedding and ripple cross lamination is a common interbedded lithology, typically preserved only as lenses 1-2 m wide and 10-50 cm thick. Desiccation cracks, root traces and greenish-grey limestone beds are present. This facies assemblage can be interpreted as subaerial alluvial fan to braidplain deposits related to fault margins in early and late Craignish times. Paleocurrent flow was generally away from (those) defineable margins.

C2) Red siltstone to fine sandstone

This facies assemblage is thickest and most prevalent near basin centres at the top of the Craignish. It is dominated by brick red massive siltstone in units up to 50 m thick, with fining-upward interbeds 1-5 m thick of red fine to medium grained sandstone with crossbedding and ripple cross lamination. Root traces, desiccation cracks, evaporitive crystal molds and green calcareous nodules are commonly present. This facies assemblage intertongues with and is considered to be the distal equivalent of that previously described (C1). It is interpreted to represent subaerial alluvial mudflat to playa conditions with abundant exposure surfaces, developed beyond the toes of braidplains. Sparse paleocurrent data do not indicate a preferred flow direction.

C3) Grey/green coarse sandstone to conglomerate

This facies assemblage is most prevalent in basin centre positions especially in the lower Craignish of western Cape Breton and the upper Craignish of northern Cape Breton. It is dominated by grey pebbly medium to coarse grained sandstone and granulestone in sharp-based fining-upward beds with trough crossbedding or horizontal lamination. Greenish-grey siltstone to fine grained sandstone with rippling or thin lamination is a common interbedded lithology and may dominate, particularly in mid-Craignish positions. This facies assemblage can be interpreted as alluvial and braidplain sediments deposited in low-lying central parts of the depositional basin during a period of relative stability.

Table 1. Horton Group megafacies and facies assemblages. Megafacies equate to lithologically-defined formations and are in stratigraphic order. Facies assemblages do not occur in specific stratigraphic order and are placed according to their lateral position in the sub-basins. css = coarse sandstone; mss = medium sandstone; fss = fine sandstone; vfss = very fine sandstone; si = iltstone; cgl = conglomerate; bldr = boulder.

MEGAFACIES	FACIES ASSEMBLAGES	POSITION
AINSLIE	A1 red/grey pebbly css-cgl (alluvial fan/braidplain)	↑ basin margin
	A2 red fss-css & si (fluvial)	
	A3 grey/green vfss-fss & si (high sinuosity fluvial)	↓ basin centre
STRATHLORNE	S1 dark grey burrowed mudstone (open lacustrine)	↑ basin centre
	S2 grey/green vfss-fss (nearshore/shoreline)	
	S4 red si-fss (shoreline/mudflat)	
	S3 grey/green mss-bldr cgl (fan delta)	↓ basin margin
CRAIGNISH	C3 grey/green css-cgl (braidplain)	↑ basin centre
	C2 brick red si-fss (mudflat/playa)	
	C1 red/orange css-cgl (alluvial fan/braidplain)	↓ basin margin

The Strathlorne megafacies

This megafacies is up to 600 m thick, is thickest near basin centre and consists of grey thinly interbedded mudstone and very fine sandstone, with thicker fine to coarse grained sandstone beds and thin limestone beds. The great apparent thickness in a few sections is partly due to thrust repetition and 100-400 m is more typical. It is interpreted to represent a phase of predominantly lacustrine deposition punctuated by coarser grained input from fault-bounded basin margins. Based on facies distribution and paleocurrent data, deposition in the two sub-basins was likely separate. Four facies assemblages are included, designated S1, S2, S3 and S4.

S1) Dark grey mudstone

This facies assemblage, the most recognizable lithofacies of the Strathlorne, is thickest at basin centres and thins gradually toward the margins. It is dominated by dark grey thinly interbedded siltstone, claystone and very fine sandstone, typically in coarsening-upward sequences 2-50 m thick, with burrows, ripples and some desiccation cracks on the capping beds. Many thinly laminated pelletal limestones (some with stromatolitic growths, oolites, desiccation cracks and intraclastic breccias) commonly occur at the tops of these sequences. This facies assemblage can be interpreted as low energy open lacustrine deposits arranged in coarsening-upward cycles related to fault-subsidence-fill cycles. Virtually all paleocurrent data indicate a dominant paleoslope for each sub-basin, perpendicular to the basin axis. These paleoslope directions are southeastward for the western sub-basin, and northward for the northern sub-basin, suggesting two en echelon half-grabens with opposite polarity of asymmetry (Figs. 2 and 3).

S2) Grey/green fine sandstone

This facies assemblage is best developed on the hanging wall margins of the asymmetrical sub-basins, at the tops of coarsening-upward sequences, and at the base or top of the Strathlorne. The dominant lithology is grey to greenish well sorted very fine to fine grained sandstone in sharp-based units 3-10 m thick with low angle lamination, hummocky cross-stratification or trough crossbedding. Green siltstone and grey sandy oolitic limestone also occur. Paleocurrent flow was generally away from sub-basin margins. This facies assemblage represents periodic nearshore/shoreline progradation into more open lacustrine settings, probably related to subsidence-fill cycles, or as transitional units between lacustrine and fluvial settings.

S3) Grey/green medium sandstone to conglomerate

This facies assemblage is only present near inferred fault margins and is usually part of a transitional sequence near the base or top of the Strathlorne, intertonguing with finer sediments. The dominant lithology is micaceous pebbly medium to coarse grained sandstone in sharp-based fining-upward beds with trough crossbedding. Poorly sorted matrix-supported boulder conglomerate occurs in sections interpreted as proximal to fault margins, while thin interbeds of rippled siltstone to fine sandstone with roots occur in sections interpreted as distal to fault margins. Paleocurrent flows were away from interpreted fault margins and the assemblage is interpreted as high energy fault margin detritus shed directly into lacustrine environments, perhaps on fan deltas.

S4) Red siltstone to fine sandstone

This facies assemblage is a minor one, occurring as thin units associated with shoreline facies near fault margins. Brick red massive sandy siltstone with roots and desiccation cracks is interbedded with red very fine to fine grained sandstone beds with straight-crested ripples. Some red fine sandstone beds, up to 3 m thick, with scoured bases, crossbedding and ripple cross lamination, and thin greenish limestone beds with root traces also are present. This facies assemblage can be interpreted as shoreline-related sedimentation on low energy mudflats and in small fluvial channels.

The Ainslie megafacies

This megafacies, up to 600 m thick, consists of red or grey interbedded sandstone, siltstone and conglomerate which is conformably but sharply overlain by the basal Macumber Formation of the Windsor Group. Thrust repetition in some sections exaggerates the apparent thickness, but it is thickest, coarsest-grained and reddest near interpreted fault margins, and thin and grey near basin centres. It is only well exposed in the western Cape Breton sub-basin. It generally represents a period of less subsidence and consequent infill of the lacustrine basin by fluvial/alluvial sediments shed from the fault margins. Three facies assemblages are included, designated A1, A2, and A3.

A1) Red/grey coarse sandstone to conglomerate

This facies assemblage is very thick at the fault margins but extends only a short distance toward basin centres. It is dominated by red (western sub-basin) or greenish (northern sub-basin) matrix- to clast-supported conglomerate, granulestone and pebbly coarse sandstone in coarsening-upward sequences up to 75 m thick. These sharp-based, fining-upward beds have clast imbrication, broad trough crossbedding and sandstone lenses with green calcareous nodules. Paleocurrent flow was away from fault margins. This facies assemblage is interpreted as alluvial fan and proximal braidplain deposits dominated by sheet floods. More proximal debris deposits are not preserved.

A2) Red fining-upward sandstone and siltstone

This facies assemblage is prevalent in medial positions in the western sub-basin, distal to the previous assemblage. It is dominated by red micaceous fine to coarse grained sandstone in 1-5 m fining-upward beds with scoured bases, trough crossbedding, ripples and root traces. These are separated by red massive to laminated sandy siltstone which may include greenish-grey limestone beds and nodules with roots. Paleocurrent flow was away from basin margins, especially the ends of the sub-basin. These deposits are attributed to a fluvial system at the distal ends of the marginal braidplains.

A3) Grey/green fine sandstone to siltstone

This facies assemblage is dominant near basin centres and the hanging wall side of depositional sub-basins, but does not occur near fault margins. At the western sub-basin centre the Ainslie is very thin and consists only of this facies assemblage. It is characterized by interbedded greenish-grey fining-upward very fine to medium grained sandstone and grey or reddish-grey sandy siltstone. Sandstone beds are 1-5 m thick, have sharp bases with rip-ups, pebble lags, horizontal lamination, trough crossbedding and ripples. Siltstones are micaceous and laminated or burrowed; Paleocurrent data display a high variance, but may suggest flow along the length of the basin. These deposits are interpreted to represent a high sinuosity longitudinal fluvial system flowing along the basin centre, removed from fault margin sediment input.

RESOURCE POTENTIAL

The juxtaposition of dark fine grained facies and red coarser grained facies, confined in a localized structural basin and overlain by a regionally continuous carbonate/evaporite unit (Windsor Group) are all favourable characteristics for the resource potential of the Horton Group. The identification of two sub-basins and interpretation of these as asymmetrical half-grabens may help in defining areas of suitable petroleum reservoir/mineral host facies, facies pinchouts, potential source rock facies and potentially advantageous structural features.

The identification, description, eventual mapping and interpretation of the various facies assemblages will lead to more detailed conclusions regarding possible exploration philosophies. Understanding the organization and geometry of these depositional units is paramount to predicting the resource potential of this and similar areas. In an asymmetrical half-graben, the deepest quietest area where anoxic organic-rich petroleum source rock might accumulate, is in the axial zone near the main controlling fault where coarse clastic input is limited to a narrow belt adjacent to the margin. The presence of major thrust repetition of parts of the Horton Group and offset of the basin margin in western Cape Breton enhances the exploration possibilities there but requires a different exploration approach. Detailed mapping of contacts between red coarse grained and dark grey fine grained facies would delineate redox boundaries where maximum potential for metallic accumulations exist.

ACKNOWLEDGMENTS

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A preliminary report on the lithostratigraphy of lower Middle Ordovician sedimentary rocks, lower Ottawa Valley, Ontario

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Steele-Petrovich, H.M., A preliminary report on the lithostratigraphy of lower Middle Ordovician sedimentary rocks, lower Ottawa Valley, Ontario; in Current Research, Part B, Geologicial Survey of Canada, Paper 89-1B, p. 121-125, 1989.

Abstract

The Middle Ordovician sedimentary rocks of the traditional Rockcliffe, Pamelia, Lowville, Chaumont (or Leray), and Rockland formations in the eastern part of the Ottawa Valley form an uninterrupted depositional sequence that is subdivided into four lithostratigraphic units composed of recurring, interbedded lithofacies. A comparison with rocks from a similar stratigraphic interval in the western part of the valley shows considerable lithological differences between the two areas.

Résumé

Les roches sédimentaires datant de l'Ordovicien moyen des formations classiques de Rockcliffe, Pamelia, Lowville, Chaumont (ou Leray) et Rockland, situées dans la partie est de la vallée de l'Outaouais, forment une séquence sédimentaire ininterrompue qui se laisse subdiviser en quatre unités lithostratigraphiques de lithofaciès répétitifs et interstratifiés. Une comparaison avec des roches provenant d'un intervalle stratigraphique similaire dans la partie ouest de la vallée, montre qu'il existe des différences lithologiques considérables entre les deux régions.

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INTRODUCTION

This preliminary report is based on fieldwork done during the summers of 1986 and 1987. The Middle Ordovician sedimentary rocks reported on here occur in the eastern part of the Ottawa Valley, Ontario, between Ottawa and Hawkesbury (Fig. 1). The rocks are briefly described and compared with previously studied (Steele-Petrovich 1984, 1986) rocks of the same age that occur west of Ottawa.

The stratigraphic sequence considered here includes the units that have been traditionally assigned to the Rockcliffe Formation and St. Martin Member of the Chazy Series, the Pamelia, Lowville, and Chaumont (or Leray) formations of the Black River Group, and the Rockland Formation of the Trenton Group. Several recent workers (Williams and Rae, 1983; Williams and Telford, 1986, 1987) have followed Liberty's (1955, 1964, 1967) classification, originally defined for central and southwestern Ontario, and divided these rocks into the Shadow Lake, Gull River and Bobcaygeon formations of the Simcoe Group. The relationships between the different stratigraphic classifications are shown in Table 1 of this paper.

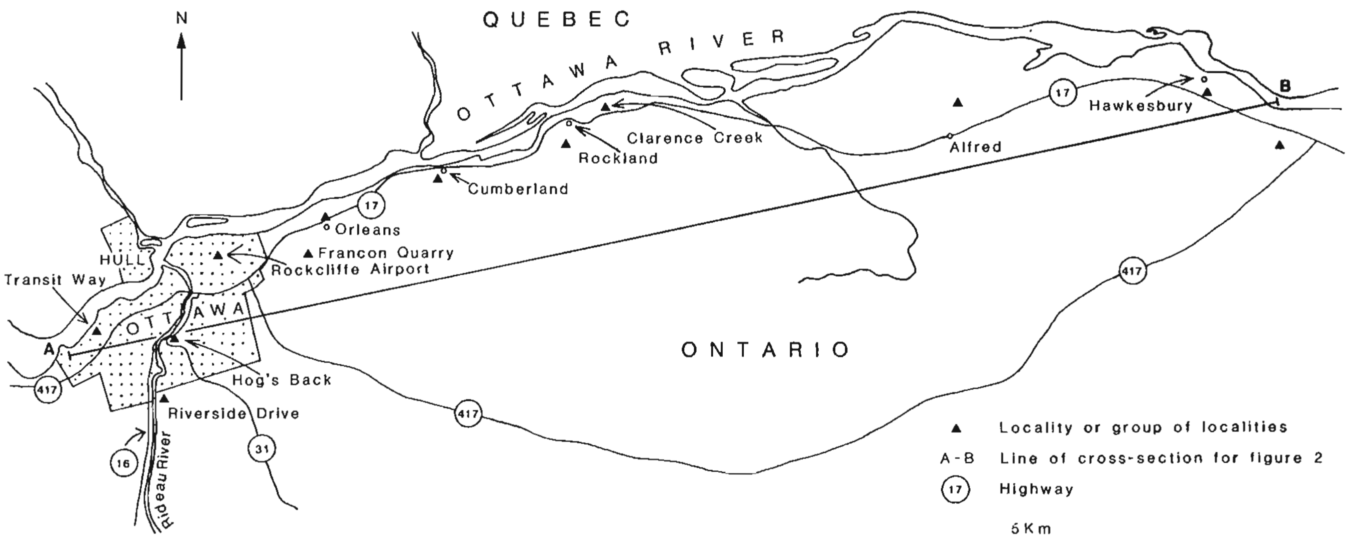

Figure 1. Locality map.

Table 1. Relationships between different lithostratigraphic classifications for the Ottawa Valley.

<table>
<tr><th colspan="2">Proposed Lithostratigraphic Units</th><th colspan="2" rowspan="2">Traditional Classification (e.g. Wilson,1946)</th><th colspan="2">After Liberty (1964,1967)</th></tr>
<tr><th>Western Part of Ottawa Valley (Steele-Petrovitch 1986)</th><th>Eastern Part of Ottawa Valley (this study)</th><th>Williams & Rae (1983)</th><th>Williams & Telford (1986)</th></tr>
<tr><td>F</td><td rowspan="3">δ</td><td>Trenton</td><td>Rockland</td><td rowspan="3">Bobcaygeon</td><td rowspan="3">Bobcaygeon</td></tr>
<tr><td>E</td><td rowspan="6">Black River</td><td rowspan="2">Chaumont</td></tr>
<tr><td rowspan="2">D</td></tr>
<tr><td rowspan="2">γ</td><td rowspan="2">Lowville</td><td rowspan="4">Gull River</td><td rowspan="3">Gull River</td></tr>
<tr><td>C</td></tr>
<tr><td rowspan="2">B</td><td rowspan="2">β</td><td rowspan="2">Pamelia</td></tr>
<tr><td>Shadow Lake</td></tr>
<tr><td rowspan="2">A</td><td rowspan="2">α</td><td rowspan="2">Chazy</td><td>St.Martin</td><td rowspan="2">Rockcliffe</td><td rowspan="2">Rockcliffe</td></tr>
<tr><td>Rockcliffe</td></tr>
</table>

METHOD OF STUDY

The studied outcrops of Middle Ordovician sedimentary rocks are relatively widely spaced on the southern side of the Ottawa River (Fig. 1) in a highly faulted terrain (Wilson, 1946; Sanford et al., 1979). Composite sections from 12 localities were studied in the field. The lithofacies were identified; the thickness and physical characteristics of each occurrence of a lithofacies were recorded, and at least one rock sample of each occurrence and as many of the contained fossils as possible were collected for laboratory studies.

LITHOSTRATIGRAPHIC SUBDIVISIONS

Certain lithofacies that occur in these rocks are highly interbedded, making correlation of individual lithofacies difficult. However, sets of interbedded lithofacies form well delineated lithostratigraphic units, provisionally called α, β, γ, δ (Table 2). These lithostratigraphic units have been correlated throughout the study area (Fig. 2) and should have formation (or member) status.

The boundary between the α and β units is the historical boundary between the Chazy Series and the Black River Group. In most of the field area, the Chazy/Black River contact is not exposed and occurs within a covered interval which, at least in part, has resulted from weathering of a brown shale horizon (cf. Wilson, 1932, 1937); in many places this covered interval occurs directly above the uppermost occurrence of the St. Martin limestone, which commonly contains abundant specimens of the rhynchonellid, *Rostricellula sp.* In the Transitway section at the western edge of the present field area (Fig. 1), the St. Martin limestone is absent, and vertical exposures are continuous across the Chazy/Black River contact; here, the boundary between the α and β units is placed provisionally where carbonates begin to dominate over clastics. The boundary between the β and γ units is placed at the top of the last occurrence of silt. The boundary between the γ and δ units is sharply defined, at the base of the very thick beds (Fig. 2). Future work on the rocks that are just higher stratigraphically than those of this study should determine whether the microbedded and microcrossbedded packstone at the top of the Rockland section should be placed in Unit δ or in a higher, as yet undefined unit. The lithostratigraphic units of this study fit into the lithostratigraphic nomenclature of the Ottawa Valley as shown in Table 1.

Table 2. Lithostratigraphic units and their lithofacies composition.

Lithostratigraphic Units	Lithofacies
δ	(?)Microbedded and microcrossbedded (?)packstone (locally at top — may belong to a higher unit) Lime mudstone to fossiliferous wackestone (massive bedding)
γ	Interbeds of: Sublithographic mudstone Ooidal, peloidal, intraclastic packstone to wackestone Skeletal packstone Lime mudstone to fossiliferous and/or peloidal wackestone
β	Mainly silty carbonate/calcareous siltstone (commonly dolomitic), interbedded with: Calcareous, silty shale/very shaly, silty carbonate (commonly dolomitic) Lime mudstone to fossiliferous and/or peloidal wackestone Calcareous shale and very shaly limestone Skeletal packstone (locally) Ooidal, peloidal, intraclastic packstone to wackestone (locally silty)
α	Mainly terrigenous sandstones and shales. Some sandy carbonates, primarily toward the top (commonly dolomitic and/or shaly and containing the brachiopod, *Rostricellula* sp.)

STRATIGRAPHIC DIFFERENCES BETWEEN EASTERN AND WESTERN PARTS OF THE OTTAWA VALLEY

There are considerable differences between the Ordovician rocks of the eastern and western parts of the valley, and the changes occur in the vicinity of Ottawa. The differences are summarized here and will be dealt with in greater detail in a later publication.

1. In the studied sequence (traditional Rockcliffe to Rockland formations), certain lithofacies occur in both parts of the valley but others are unique to either the east or the west; for example, lithofacies 8, 9, 10, which occur west of Ottawa (Steele- Petrovich, 1986) are absent from the eastern part of the valley. The same stratigraphic interval in the east is occupied by the very thick beds of unit δ.
2. The carbonates in the east have a much higher shale content than in the west.
3. The deposition of silt-sized quartz grains ceased relatively abruptly in the west with the first appearance of Black River carbonates. In contrast, silty carbonates are common in the lower part of the Black River sequence in the east, where they are interbedded with non-silty carbonates.
4. Different lithofacies are associated with the sublithographic limestone in the two parts of the valley. In the west, the sublithographic limestone (Lithofacies 6) is usually interbedded with a lime mudstone/wackestone (Lithofacies 7) (Steele- Petrovich, 1986); in the east it appears to be closely associated with an oolitic lithofacies.
5. Rocks of the Trenton Group that are stratigraphically higher than the traditional Rockland Formation are common east of Ottawa, but are very rare, if present at all, in the western part of the valley.

THE CHAZY QUESTION

In a previous study (Steele-Petrovich, 1984, 1986) the lowermost Ordovician sandstones and shales in the western part of the Ottawa Valley were assigned to the lower part of the Pamelia Formation, within the Black River Group. This designation was based on Wilson's (1932, 1936) descriptions of the Lower Pamelia and on recent field evidence (Steele-Petrovich, 1984, 1986) showing that the terrigenous and carbonate rocks that occur there form a continuous depositional sequence; it is also consistent with a recent interpretation of a stratigraphic sequence of rocks of similar age in New York State, where the basal terrigenous rocks are included in the Pamelia Formation (Walker, 1973). However, the lowermost Ordovician sandstones and shales west of Ottawa fit equally well Wilson's (1946) description of the Rockcliffe Formation, and Kay (1942) previously assigned the terrigenous rocks of the western part of the Ottawa Valley to the Chazy Series.

East of Ottawa, the Ordovician sandstones and shales that occur immediately below the Black River carbonates are clearly of Chazyan age and belong to the Rockcliffe Formation: a thin limestone unit (St. Martin Member) containing *Rostricellula* sp., a diagnostic Chazyan brachiopod, is commonly found near or at the top of the sandstone-shale sequence in the eastern part of the valley (but becomes thinner toward the west and essentially disappears in the vicinity of Ottawa). As the Chazyan sandstones and shales are the stratigraphic equivalents of the lowermost sandstones and shales that occur west of Ottawa, both sequences are considered here to be Chazy Series.

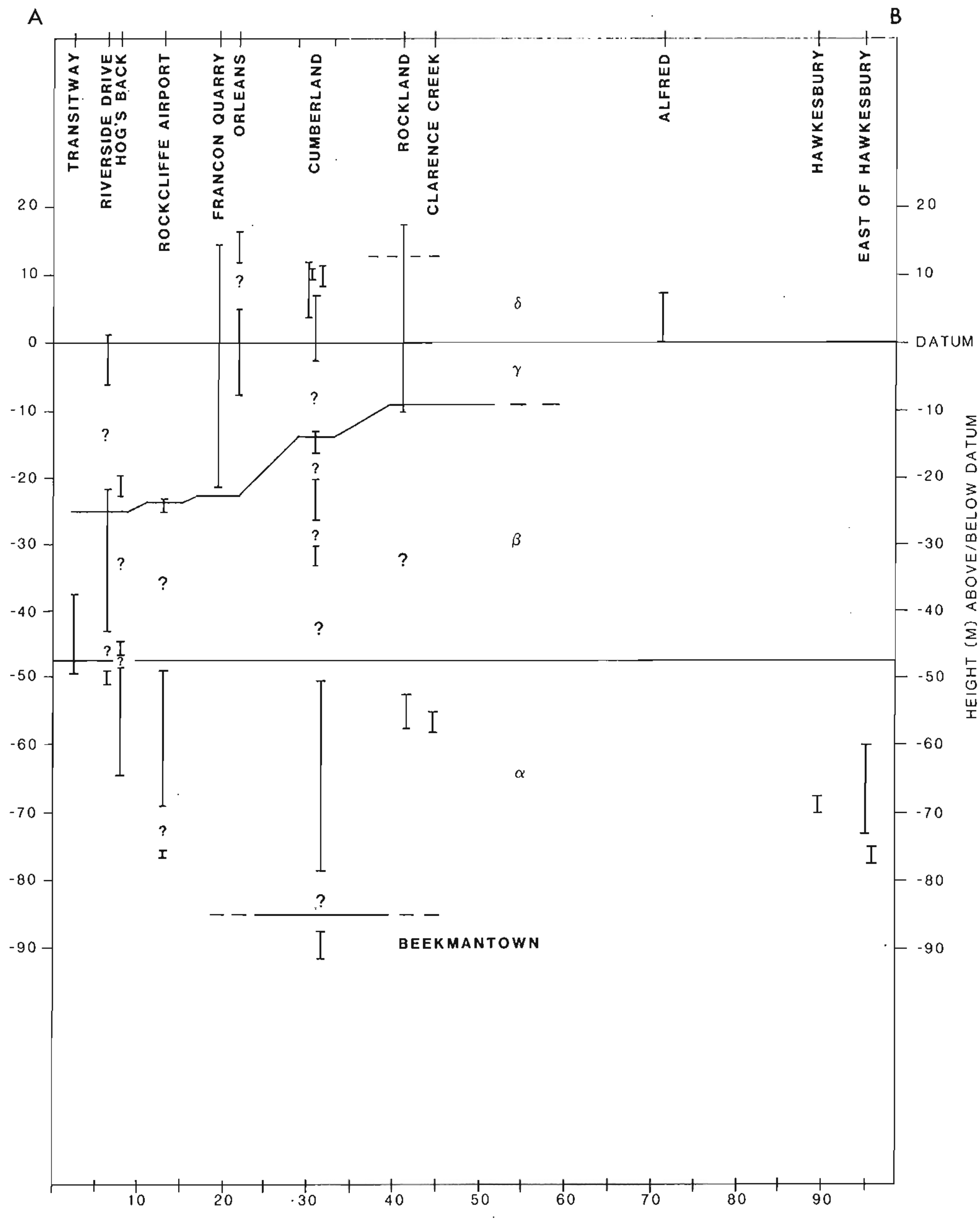

Figure 2. Measured sections as projected onto AB of Figure 1. Datum: base of very thick beds forming δ unit. Secondary datum: α/β boundary. $\alpha\beta\gamma\delta$ proposed lithostratigraphic subdivisions; AB, line of sections shown in Figure 1; ?, covered interval of unknown thickness.

Traditionally, the Black River rocks of the Ottawa Valley have been thought to be separated from the Chazy by a disconformity (eg., Wilson 1932, 1937,1946; Cooper 1956; Barnes et al., 1981), although Wilson (1937) noted difficulty in distinguishing the disconformity because of lithological similarities between the rocks on both sides of it. However, results of recent fieldwork east of Ottawa suggest, as did results of previous fieldwork farther west (Steele-Petrovich, 1984, 1986), that the Chazy sandstones and shales and the overlying Black River carbonates form an uninterrupted depositional sequence. There is no locality east of Ottawa where the outcrop is continuous vertically from the lower sandstones and shales up through the higher carbonates (a covered interval invariably occurs at the top of the Chazy), and, although a continuous sequence across the boundary is seen in the Transitway section near the western side of Ottawa, the depositional regime there is more closely related to that farther west. However, in the east, at least one relatively common and very distinctive type of silty carbonate that occurs in the Chazy rocks below the covered interval also occurs above it, interbedded with Pamelia carbonates, implying continuous deposition across the Chazy/Black River boundary.

The above evidence raises further questions about the relationships between the Pamelia, Rockcliffe, Shadow Lake, Black River, and Chazy units, both east and west of Ottawa. It also raises questions about Cooper's (1956) stages, as Cooper equated the generally accepted hiatus between the Chazy and Black River rocks in the Ottawa Valley with two of his brachiopod stages. Many of these questions should be answered as this work continues.

ACKNOWLEDGMENTS

I thank M. J. Copeland for help in numerous ways, S. Petrovich for assistance in the field, I. Munro for helpful discussions at some outcrops, R. Petrovich, G. S. Nowlan for editorial comments, and C. Bolton for drafting the figures. This work was performed under GSC contract nos. 54SZ.23233-6-0630 and 34SZ.23233-7-0179.

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Glacial dispersal of Precambrian Shield and local Appalachian rocks in the lower St. Lawrence region in western Gaspésie, Quebec, and in adjacent New Brunswick[1]

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Rappol, M. and Russell, H., Glacial dispersal of Precambrian Shield and local Appalachian rocks in the lower St. Lawrence region in western Gaspésie, Quebec, and in adjacent New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 127-136, 1989.

Abstract

The distribution and frequency of Precambrian erratics in surficial deposits south of St. Lawrence River valley have been controlled by an early Late Wisconsinan invasion of Laurentide ice across the area in an east-southeasterly direction, and by subsequent flow from Appalachian ice divides toward St. Lawrence Valley and other areas of drawdown, redistributing much of the Precambrian debris. High frequencies of Precambrian erratics are found along the Lac Témiscouata low and in areas of Late Wisconsinan Appalachian ice divides. Lower values occur along St. Lawrence Valley (above marine limit), and especially west of Lac Matapédia, where strong late glacial outflow removed most of the Precambrian material.

Observed dispersal of a number of Appalachian rock types provides an additional control on variations in ice flow directions and supports the sequence of events outlined above.

Résumé

La répartition et la fréquence des roches erratiques précambriennes dans les dépôts en surface accumulés au sud de la vallée du Saint-Laurent ont été déterminées par une invasion des glaces laurentidiennes au début du Wisconsinien supérieur à travers la région dans une direction est sud-est, puis par un écoulement des glaces à partir des lignes de partage des glaces des Appalaches vers la vallée du Saint-Laurent et d'autres zones d'ablation des glaces; ces deux épisodes ont redistribué une grande partie des débris précambriens. Il existe un grand nombre de roches indicatrices d'âge précambrien, en particulier le long de la dépression du lac Témiscouata, dans les zones des lignes de partage glaciaire des Appalaches datant du Wisconsinien supérieur. On trouve des concentrations de blocs plus faibles le long de la vallée du Saint-Laurent (au-dessus de la limite marine), et surtout à l'ouest du lac Matapédia, où un fort écoulement tardiglaciaire a éliminé une grande partie des matériaux précambriens.

Les observations faites sur la dispersion de plusieurs types de roches provenant des Appalaches viennent s'ajouter aux contrôles dont on dispose déjà sur les variations que montrent les directions d'écoulement des glaces, et confirment la séquence d'événements brièvement passée en revue ci-dessus.

[1] Contribution to the Canada Economic Development Plan for Gaspé and Lower St.Lawrence, Mineral Program 1983-1988. Project carried by the Geological Survey of Canada.

INTRODUCTION

The presence of Precambrian Shield erratics above marine limit in the Appalachian region south of St. Lawrence River represents unequivocal evidence for a Laurentide ice cover at some time. However, little is known about the actual distribution and frequency of occurrence of these erratics. This situation allows considerable variation in reconstructions of the glacial history of the region, either with minimal portrayal of Laurentide influence (e.g., Prest, 1984; Rampton et al., 1984) or with supposed major overriding of almost the entire Altantic region by Laurentide ice (e.g. Hughes et al., 1985). Moreover, the timing as well as main flow direction of the Laurentide ice invasion are disputed issues among glacial geologists working in the area (cf. e.g., Rampton et al., 1984; David and Lebuis, 1985; Hughes et al., 1985; Prichonnet and Desmarais, 1985; Lowell and Kite, 1986; Lortie and Martineau, 1987.

During the summer of 1988 more than 500 boulder counts were done in parts of the Lower St. Lawrence region, western Gaspésie, and northwestern New Brunswick to establish a first approximation of the distribution and frequency of Precambrian Shield erratics in surficial deposits of the study area (Fig. 1). This was combined with a study of the glacial dispersal of several distinctive local igneous rock types and an inventory of bedrock striation sites. This report discusses some initial results of the information obtained. These results put some important constraints on future reconstructions of the glacial history in the region.

METHOD

In general 1000 cobbles and boulders (in only 3 samples, less than 500) in the 10-30 cm fraction were inspected at each site. Sample sites include gravel pits and roadcuts, boulder piles in fields, surface boulders in forestry areas, and some natural exposures and streambeds. The sampled material occurs in a variety of glacial deposits, such as eskers and kames, glaciomarine deltas and subaqueous fans, and till sheets. In the case of till, counted material generally represents only the bouldery surface layer; few counts were actually done within massive and dense subglacial till.

At each site, the number of shield erratics and selected Appalachian lithological indicators of glacial transport were determined. Anorthosites, being distinctive shield indicators, were counted separately during part of the project.

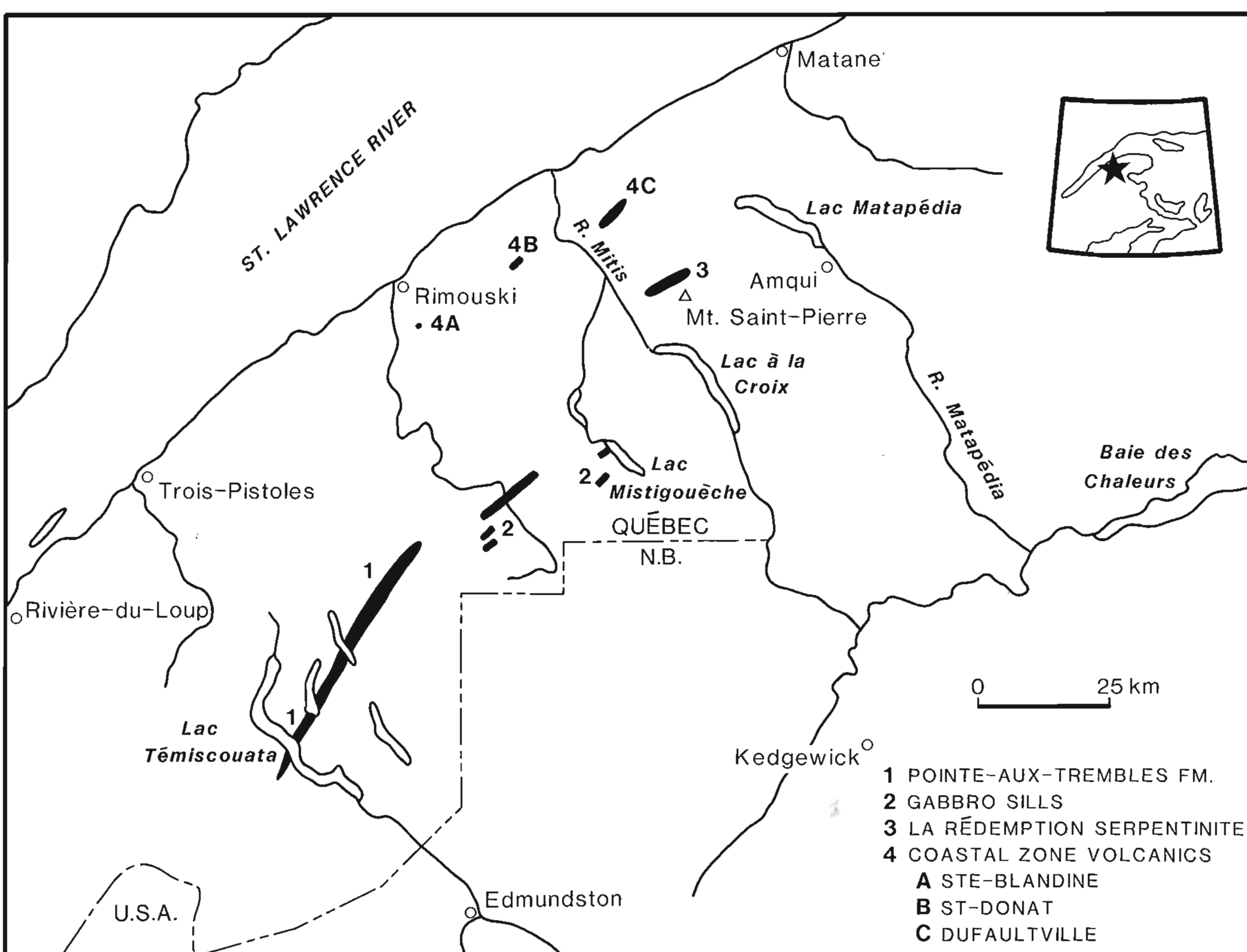

Figure 1. Location map.

Although it was attempted to spread sample sites more or less evenly over the area, poor access and exposure, scarceness of surficial deposits, and also limited available time caused some areas to be insufficiently sampled.

In general, no counts were done below marine limit where the abundance of Precambrian Shield erratics is well established and attributed to glaciomarine deposition and ice rafting during and after deglaciation (see Dionne, 1971, 1987).

BEDROCK SETTING

The estuary of St. Lawrence River separates the Grenville Province of the Canadian Shield from the Appalachian mountain range south of the river. Appalachian rocks of the study area are predominantly of sedimentary origin and Paleozoic age. A distinct northeast-southwest structural trend prevails, with structural complexity being highest along St. Lawrence River. With one exception, that is, amphibolite associated with the La Rédemption serpentinite, metamorphism never exceeds phyllitic grade.

The coastal zone consists of a series of thrust-imbricated slices of Cambro-Ordovician sediments, which form the Québec anticlinorium (Poole and Rodgers, 1972). In the northern part of the study area, a number of distinctive volcanic rocks occur in this zone (Fig. 1): the Dufaultville basalt (Aubert de la Rue, 1941), the St-Donat basalt (Liard, 1972), and the Ste-Blandine andesite (this report).

The remainder of the area is located within the Gaspé-Connecticut Valley synclinorium. The Silurian and Devonian sediments of this area rest unconformably upon the Cambro-Ordovician basement. Among other places, this contact is exposed along the northern flank of Mont Saint-Pierre.

North of Lac Témiscouata, the Silurian Pointe-aux-Trembles Formation comprises volcanic conglomerates and a few lava flows (Lespérance and Grenier, 1969; David et al., 1985). In the vicinity of Lac du Pain de Sucre, the Pointe-aux-Trembles Formation is intruded by a number of small dioritic bodies (Lespérance and Grenier, 1969). Farther to the northeast, the stratigraphic equivalent of the Pointe-aux-Trembles Formation, the Lac Raymond Formation, is intruded by a series of gabbroic sills (Lajoie, 1971). Related gabbro sills were mapped by Béland (1960) west of Lac Mistigouéche and we observed several previously unmapped occurrences.

The La Rédemption serpentinite complex is exposed within the package of Cambro-Ordoviaian rocks of the northern flank of Mont Saint-Pierre (Fig. 1) and comprises mainly serpentinite and amphibolite (Aubert de la Rue, 1941; Béland, 1960).

A number of small igneous intrusives occur within the New Brunswick part of the study area, that is, north of Edmundston and west of Campbellton (Potter et al., 1979). These consist of distinctive quartz-feldspar porphyry and diorite.

PREVIOUS WORK

The frequency of Precambrian Shield erratics in surficial deposits of the coastal area of Rivière-du-Loup and Trois-Pistoles has been studied by Dionne (1971). He found that glacial deposits at or above marine limit contain relatively few Precambrian clasts (0.3 % in till, 1.3 % in deposits of the St-Antonin Moraine, and 1.5 % in glaciofluvial deposits) when compared to deposits below marine limit, for example marine beach deposits (over 10 %).

Results of many boulder counts for an area in northwestern Maine are given by Lowell and Kite (1986). Surficial deposits in their study area contain up to 4.4 % shield erratics, but on average only 0.4 %.

David and Lebuis (1985) indicated areas of greater and lesser abundances of Precambrian erratics in western Gaspésie, and infer the presence of a Precambrian indicator train extending in a southeastern direction along lower Matane River valley. Prichonnet and Desmarais (1985) presented some results of boulder counts in the Lac Matapédia area.

Dispersion of local rock types was studied by Lebuis (1973), Lebuis andDavid (1977), Martineau (1979), David and Lebuis (1985), Prichonnet and Desmarais (1985), Halter (1986), and Chauvin and David (1987). In general, two main dispersal trains are observed: one toward the southeast and one in a northern or northwestern direction.

Available striae information was recently reviewed by Lowell and Kite (1986), Rappol (1986), and Lortie and Martineau (1987).

The main features of the glacial history of the region emerging from these studies are an early invasion of Laurentide ice moving toward the east or southeast, followed by a change in direction of ice movement due to the development of an Appalachian ice divide in response to rising marine level and drawdown. With the progression of a calving bay up St. Lawrence River valley (Thomas, 1977), ice movements north of the Appalachian ice divide gradually changed from northeastward towards northward and northwestward. A late and strong westward movement of ice is observed in the Lac Matapédia area, north of Mont Saint-Pierre.

RESULTS

Precambrian erratics

Precambrian erratics from the Canadian Shield comprise a wide variety of igneous and metamorphic types. Among these, anorthosite, mangerite (both commonly sheared), gneiss, and granite-gneiss of the Grenville Province are most characteristic, but granitic rocks are also common. With the exception of local igneous rocks referred to in a preceding section, all igneous and metamorphic rocks were considered to be derived from the Canadian Shield. No indications were found, either in the field or in the geological literature, that igneous material from the Miramichi Highlands, central New Brunswick, or from central Gaspésie was transported into the study area by Appalachian ice.

Precambrian erratics are found throughout the study area (Fig. 2), and include anorthosites in all parts of the area; however, the frequency of occurrence shows large regional variation. Highest frequencies are found in the southern part of the study area around Lac Témiscouata, with a maximum of 5.0 %. Precambrian erratics generally make up 1.5-3.0 % of the total boulder assemblage in this area. Toward the coast, percentages become slightly lower: $\pm 1.0\,\%$ for deposits of the St-Antonin Moraine complex. Also farther north, highest frequencies of Precambrian erratics are generally not found along the coast, but somewhere inland. A pronounced zone of relatively high percentages of Precambrian erratics extends southward from Mont Saint-Pierre, connecting up with relatively high percentages along the Québec-New Brunswick border.

Precambrian erratics are, on the other hand, extremely scarce in the area north of Mount Saint-Pierre. This coincides with an area of strong westward flow through the Lac Matapédia low just prior to deglaciation. Ice of this late, westward flow event originated from the Notre-Dame Mountains east of Matapédia Valley.

Another area of low Precambrian erratics percentages is located around Kedgewick, New Brunswick, perhaps suggesting that Laurentide ice did not penetrate much farther into central New Brunswick. To the south, however, Precambrian erratics are still abundant in the Grand Falls area.

Assemblages of Precambrian erratics vary areally. A detailed study of these variations was not undertaken, but some observations are worth mentioning. Anorthosite constitutes more than 10 % of all Precambrian erratics counted in NTS map area 22B/4, and anorthosites are abundant among Precambrian erratics in most map areas adjacent to 22B/4. However, south and west of Lac Témiscouata, anorthosites are relatively rare (less than 2 % in map sheets 21N/10 and 21N/11). Instead, rocks of the mangerite-charnockite series, derived mainly from the Laurentian Mountains north of Québec City (Laurin and Sharma, 1975), represent a major part of the Precambrian erratics in this area. In the northeastern part of the study area, mangerite is extremely rare, and instead pinkish granitic or pegmatitic rocks are fairly common. Abundant mangerite in the

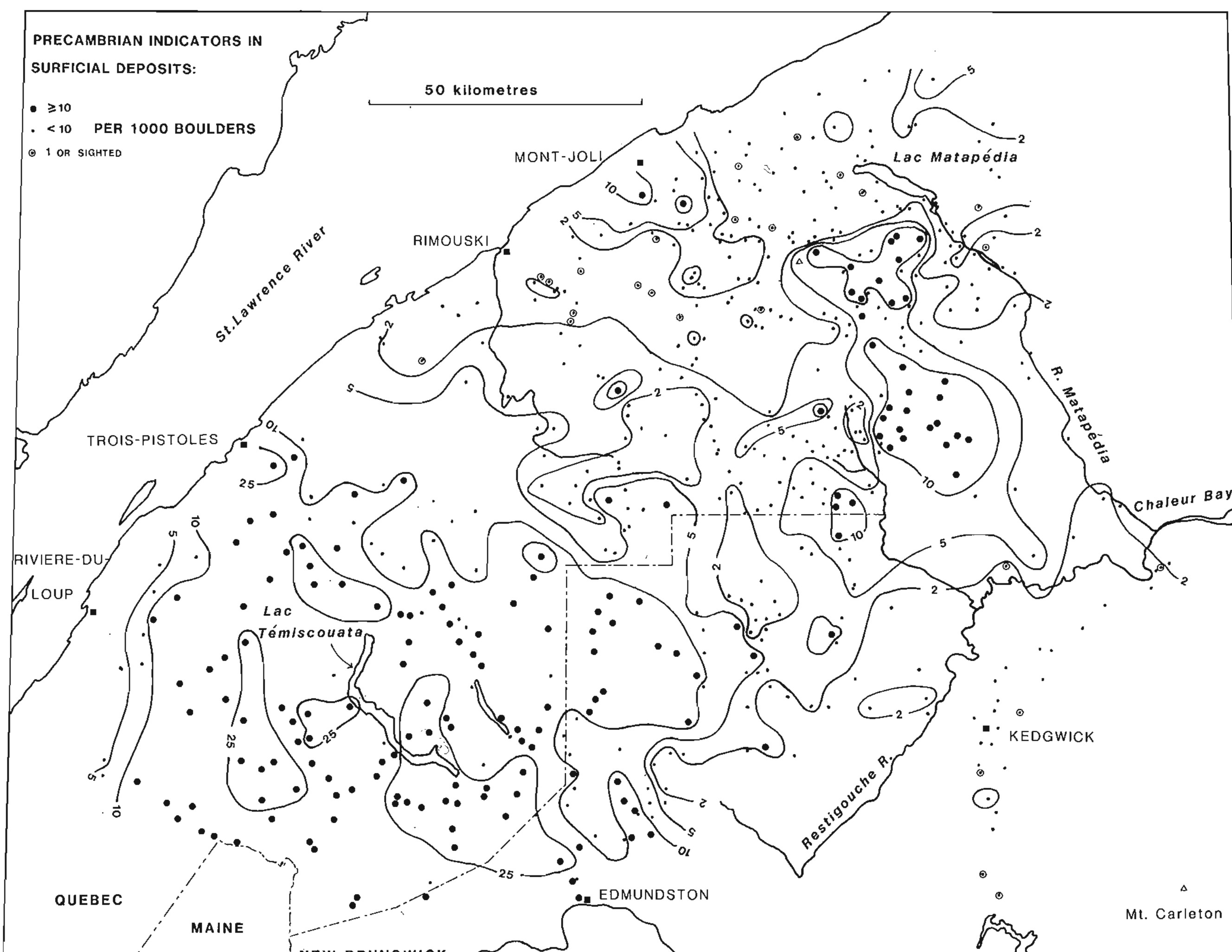

Figure 2. Frequency of Precambrian erratics, per thousand boulders counted, in surficial deposits of the study area.

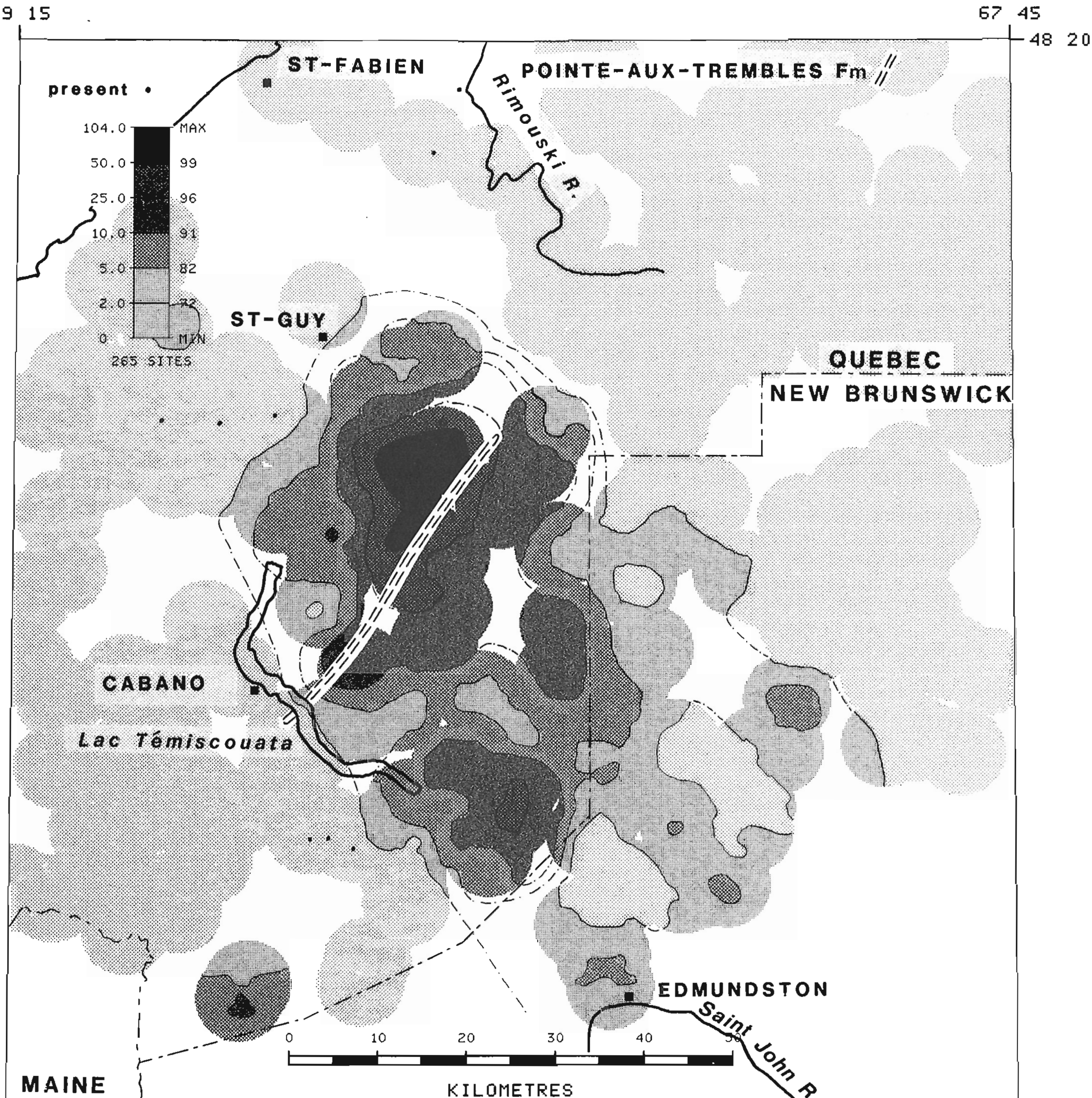

Figure 3. Dispersal of Pointe-aux-Trembles andesite; frequencies in number per thousand boulders.

southern part of the study area, and anorthosite generally more common north of this (the latter having a major source in the Lac St-Jean area), strongly suggests a main dispersal towards east or east-southeast for Precambrian erratics, rather than toward southeast.

Andesite of Pointe-aux-Trembles Formation

The Pointe-aux-Trembles Formation forms a southwest-northeast extending unit of volcaniclastic material between Lac Témiscouata and Lac du Pain de Sucre (Fig. 3). It consists of conglomerates, tuffs, and rare lavas (Lespérance and Greiner, 1969; David et al., 1985; David and Gariépy, 1986). Andesitic components of the formation, which have a distinct trachytic structure, were used to determine the dispersal pattern. Rocks derived from an occurrence of similar andesite in the Témiscouata Formation (Lespérance and Greiner, 1969), south of Lac du Pain de Sucre, were included in counts of Pointe-aux-Trembles andesite.

Because of the elongate shape of the source body, and because the andesitic components are not distributed equally along the entire length of the outcrop, detailed interpretation of the dispersal pattern is seriously hampered.

Combined with available striae information, the data of Figure 4 indicate an early dispersal toward east or southeast and a younger one toward north or northwest. For a quartz-diorite intrusion within the Pointe-aux-Trembles Formation just north of Lac du Pain de Sucre, Martineau (1979) found dispersal toward three directions — southeast, north, and northwest. Because Martineau (1979), and later Lortie and Martineau (1987), did not recognize the evidence for an important northward ice movement preceding the ice movement toward the northwest, he suggested that the northward dispersal is only apparent and possibly related to another outcrop of the same rock type. However, there is abundant evidence now for northward flow in the area. It even appears that this northward movement was more important as a debris transporting and till forming event than was the later northwestward flow. For example, a fairly extensive till sheet northwest of Saint-Guy (22C/2) overlies bedrock showing only northward striae and has a north-south fabric. Northwestward-trending striae were only formed on bedrock knobs protruding through the till sheet.

Dispersal patterns do not show a gradual increase of the frequency of indicators toward their source. Among other factors, this is probably due to disturbances produced by younger ice movements that redistributed part of the dispersed material.

Gabbro between Lac des Echos and Lac Mistigouèche

Gabbroic sills trending northwest-southeast between Lac des Echos and Lac Mistigouèche have been reported by Lajoie (1971) and Béland (1960). These sills have not been mapped in full detail, probably as a result of poor access and exposure, and their occurrence is therefore indicated only schematically in Figure 4. Although David and Gariépy (1986) suggested a fairly uniform texture and composition for these intrusives, glacially transported boulders derived from them show a wide variation in texture and occasionally contain xenoliths of the country rock.

Glacial dispersal from the gabbroic sills (Fig. 4) indicates two main modes of transport: one in a southeastern direction, another toward the north. Minor modes are recorded toward the northeast and northwest.

Within the southeastern dispersal train, two distinctive components may be distinguished. One of these is directed in a more easternly direction reaching down toward Baie de Chaleur. In the western part of the train, a strong dispersal towards the south-southeast is recorded.

Minor dispersal toward the northwest and northeast is supported by striae evidence, but the main dispersal in a northern direction appears to be almost due north, reaching down to St. Lawrence Valley.

La Rédemption serpentinite complex

The outcrop of serpentinite and associated rocks near La Rédemption occurs on the lower north flank of Mont St-Pierre, west-southwest of Lac Matapédia. Serpentinite and amphibolite are the major rock types of the unit, with minor occurrences of associated gabbro and diorite (Aubert de la Rue, 1941; Béland, 1960). Serpentinite constitutes most of the western part, whereas amphibolite forms the eastern part of the outcrop area as indicated in Figure 5.

Results of boulder counts in the area around the serpentinite complex suggest glacial dispersal in almost all directions, except toward southwest (Fig. 5). This is supported by observed bedrock striations in the region, indicating many different directions of ice movement. Older flow events were toward the southeast, east, and northeast; within the northern part of the area an old northwestward flow phase predates the northeastward flow event. Younger ice movements were toward the north, northwest, and finally west. These latter movements seem related to the progression of a calving bay up St. Lawrence Valley and subsequent deglaciation.

A pronounced northward dispersal train, as observed for the gabbro, is not present for the serpentinite or the amphibolite. This appears to be due primarily to the late westward ice movement north of Mont St-Pierre, that is also considered the reason for the near-absence of Precambrian indicator rocks in this area.

Most remarkable, however, is the near-absence of southeastward dispersal from the serpentinite body, whereas the remnant of the southeastward dispersal train from the amphibolite represents the strongest developed dispersal direction of the amphibolite. Also, eastward and northeastward dispersal from the serpentinite appears less important than for the amphibolite.

The main dispersal of the serpentinite is toward the northwest, extending to Mitis Valley, with some very high values ($>15\,\%$) in ice contact deposits just north of the outcrop area. Immediately north and east of these deposits, ice contact deposits related to a deglaciation phase of the westward flowing ice from the Lac Matapédia area (St. Cleophas lobe, cf. David and Lebuis, 1985) contain virtually no debris derived from the serpentinite complex (see also Prichonnet and Desmarais, 1985). This sharp compositional boundary represents the latest contact during deglaciation between competing ice flows from the Lac Matapédia area (westward) and Mitis Valley area (north to northeastward).

It should be noted here, that Prichonnet and Desmarais (1985) claimed that the main flow event in the region was toward the south or south-southeast, and that northward dispersal from the serpentinite complex could be due to fluvial transport preceding the ice cover. However, we were unable to relocate some of their sites with southward striae; at other sites we found clear indications (rat-tails) for northward ice movement. Moreover, dispersal from the serpentinite complex as presently established (Fig. 5) is incompatible with fluvial transport. The most northerly locations found, where bedrock striations indicate ice movement toward southeast and east, are on the south flank of Mont Saint-Pierre.

Coastal zone volcanics

Three occurrences of volcanic rocks have been identified in the coastal zone of the northern part of the study area: that is, Dufaultville basalt (Aubert de la Rue, 1941), St. Donat basalt (Liard, 1972), and Ste-Blandine andesite (see Fig. 1). Of these, the Dufaultville basalt has the largest outcrop area

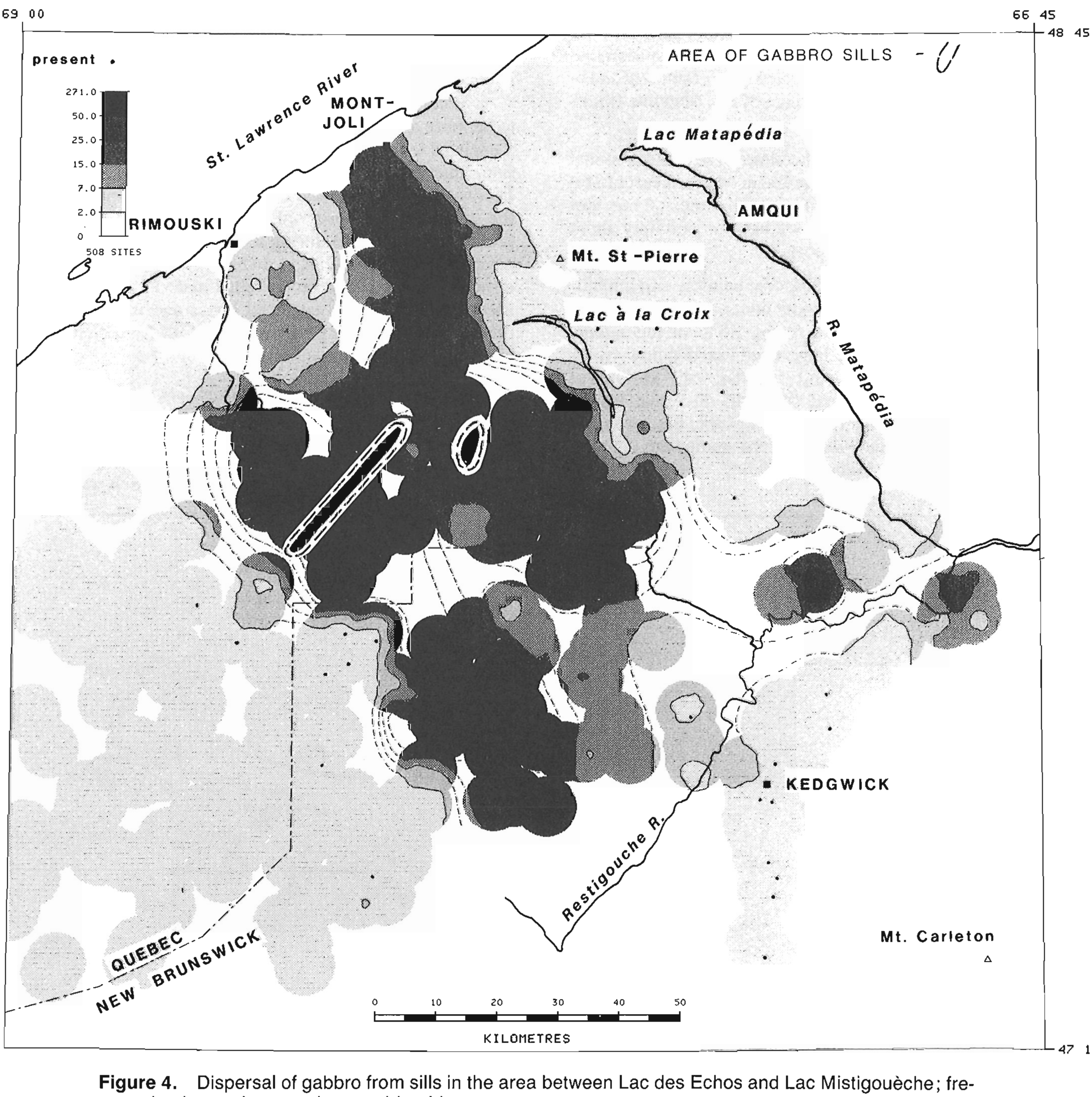

Figure 4. Dispersal of gabbro from sills in the area between Lac des Echos and Lac Mistigouèche; frequencies in number per thousand boulders.

(about 5 km along strike and up to 1.5 km wide), forming the major part of Mont Michel Nadaud. The other volcanic occurrences are found in close proximity to the base of the Neigette Fault and are of small areal extent.

The Dufaultville and St-Donat basalts are similar in appearance. Depending on the degree of alteration, both are purple or greenish, massive to vesicular, with pillow structures in outcrop. The Ste-Blandine andesite is a strongly sheared pink to purple aphanitic rock with more than 40 % calcite, filling amygdules. Its presence is indicated by a small boulder deposit exposed in the base of a gravel pit immediately north of St-Blandine (539450/5357100); the rock type was not found in situ. This boulder deposit consists almost exclusively of angular volcanic boulders and smaller rock fragments, cemented by calcite, and is overlain by ice-contact deposits. The occurrence of the Ste-Blandine andesite is probably related to a strike-slip fault that intersect the Neigette Fault at the village of Ste-Blandine (Mukherji, 1971).

Ste-Blandine andesite was encountered in boulder counts at two sites, about 4 km northeast and north-northeast of Ste-Blandine. However, because the areal extent of the Ste-Blandine andesite is unknown, we have no certainty about the direction of glacial transport.

The observed dispersal of St-Donat basalt is also limited. It was only recorded in one boulder count, located just north of the outcrop area, where it forms 1.5 % of the sample (556700/5371200). Also a huge truck-sized pillow basalt erratic (~ 120 m^3) just south of St-Donat (556050/5370300) is derived from the St-Donat basalt. Earlier, Locat (1978) referred to this block, but, probably unaware of the volcanic outcrop in the immediate vicinity, suggested a source in Matapédia Valley.

Observed dispersal from the Dufaultville basalt is in a northern and northwestern direction. No remnants of an older dispersal train toward southeast, related to Laurentide ice invasion, were encountered. Farther to the east, a near-coastal volcanic occurrence at St-Adelme, southeast of Matane, shows only significant northward dispersal (Lebuis, 1973).

DISCUSSION AND CONCLUSION

Results of current glacial research in the study area support the concept of a Late Wisconsinan invasion of Laurentide ice, followed by a drastic change in ice flow configuration, presumably as a result of marine drawdown (see Chauvin et al., 1985). Some characteristics of lithological indicator of glacial transport and their implications for the glacial history of the area are discussed below.

It is commonly observed that Precambrian erratics tend to be more abundant in the upper bouldery zones of till and associated ice contact deposits formed during final deglaciation, than in silty and compact subglacial till (see also Dionne, 1971; Lebuis and David, 1977). This indicates that the majority of observed Precambrian erratics were introduced to the area during its latest ice cover, and transported directly across St. Lawrence Valley. Precambrian erratics thus constitute far-travelled material transported in part at relatively high englacial levels and were deposited predominantly during the late phase of till formation (melt-out?) and deglaciation.

Flow velocities in active glacier ice increase rapidly from the base upwards. Precambrian material in high-level transport position therefore represents the most mobile element of the glacial debris assemblage present in the ice. Given that the ice flow configuration in the Appalachian

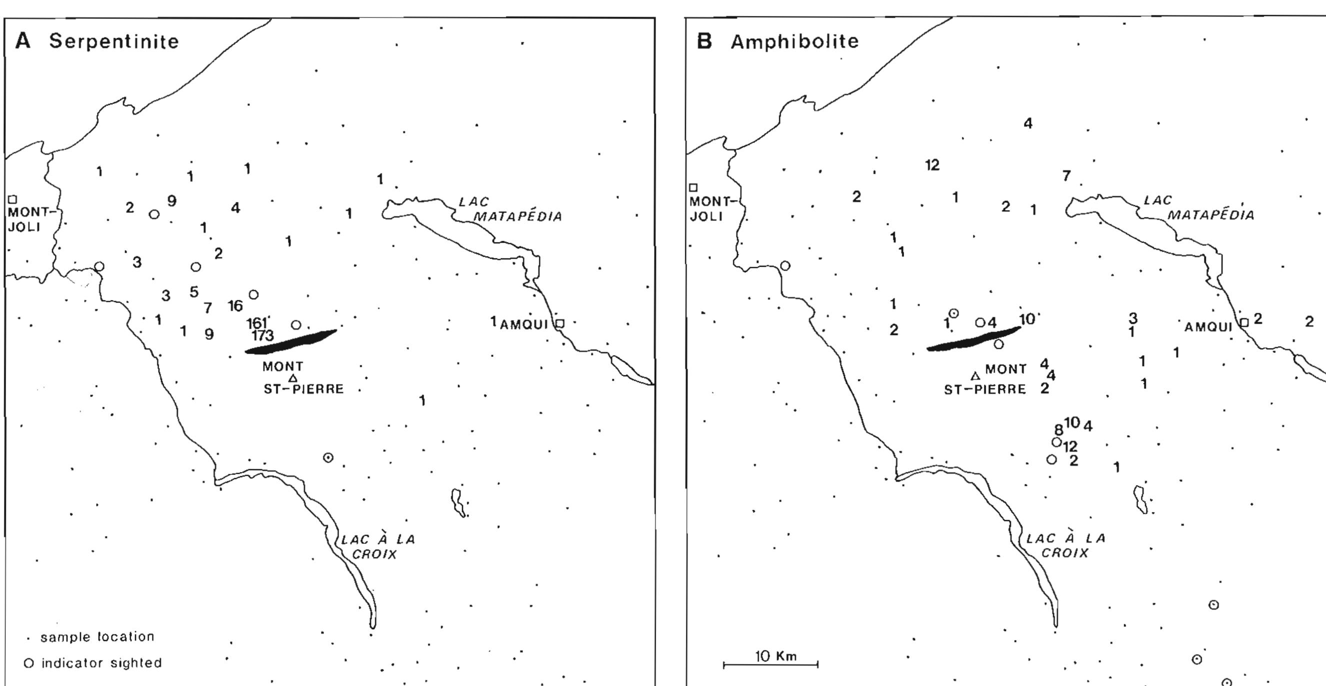

Figure 5. Dispersal of serpentinite and related gabbro (A) and amphibolite (B) from La Rédemption serpentinite complex; frequencies in number per thousand boulders.

region changed radically in response to late glacial marine drawdown, much of this high-level transport material was again removed from the area and discharged toward and down St. Lawrence Valley. The present distribution of Precambrian erratics is therefore as much a reflection of late glacial flow from Appalachian ice divides as of the earlier Laurentide invasion. After all, most Precambrian erratics were actually deposited by ice flowing from residual Appalachian ice divides.

The distribution of mangerite and anorthosite, combined with available striae evidence, suggests that Laurentide ice invaded the western part of the study area with an eastward or east-southeastward flow direction. This is in concert with findings of Lowell and Kite (1986) for northern Maine. On its way east, Laurentide ice collided with local Appalachian ice, of which that over the Miramichi Highlands represented a major barrier. As a result, Laurentide ice movement was split into two major branches: one moving eastward towards Baie des Chaleurs, the other southward down Saint John Valley. This bifurcation is also reflected in the dispersal of Appalachian gabbro as portrayed in Figure 4.

It may be assumed that during the initial invasion of Laurentide ice, there existed a gradual decrease in the concentration of Precambrian debris along ice flow lines in the down-ice direction. It is evident from Figure 2 that such a compositional gradient has not been preserved during subsequent ice movements. In fact, highest values for Precambrian erratics in surficial deposits above marine limit are not found in the more proximal coastal zone, but usually somewhere farther inland. This must be due to a further dilution of Precambrian material by incorporation of local material along flow lines away from the late glacial Appalachian ice divides.

Transport and erosion by late glacial flow from Appalachian ice divides did not affect all parts of the study area equally. North of Mont Saint-Pierre, westward and northward ice movements almost completely removed all evidence of an earlier Laurentide ice cover. Most counts in this area contained less than 0.1 % Precambrian erratics. The Precambrian erratics train along lower Matane River valley as recognized by David and Lebuis (1985) is therefore a residual phenomenon, rather than a primary dispersal train. South of Mont Saint-Pierre, however, a highf concentration of Precambrian erratics is found. We suggest that this reflects a zone of minimal outflow, being protected from drawdown by the high of Mont Saint-Pierre.

Studies on glacial dispersal of local Appalachian rock types in Gaspésie and the Lower St. Lawrence have demonstrated the presence of two major directions of glacial transport: one in a southeastern direction, the other in a northern direction (Lebuis, 1973; Martineau, 1979; David and Lebuis, 1985; Halter, 1986; Chauvin and David, 1987). Results of the present study support these findings. Strength and orientation of the dispersal trains may vary areally, particularly in connection with their position with respect to the late glacial Appalachian ice divides.

For a better understanding of the glacial history of the Appalachian region, similar research should be conducted in other parts of the region. The absence of such data from adjacent areas now hampers detailed interpretation of the presently available information. It would, moreover, be of great importance to apply quantitative counts within the Precambrian erratics assemblages in order to determine more precisely the source of Precambrian material in different areas.

ACKNOWLEDGMENTS

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Surficial geology of Saint-Joseph-de-Beauce map area, Chaudière River valley, Quebec

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Blais, A. and Shilts, W.W., Surficial geology of Saint-Joseph-de-Beauce map area, Chaudière River valley, Quebec; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 137-142, 1989.

Abstract

Surficial geology mapping in the Saint-Joseph-de-Beauce, Quebec area directed particular attention toward an inventory of bedrock striations and paleocurrent directions in ice-contact stratified drift. Preliminary results indicate that widespread, late glacial northward striations, observed to postdate southeastward Laurentide glacial striations in adjacent areas of Maine and Quebec, are also prominent in the highlands and interfluves of the St-Joseph-de-Beauce region. However, in the middle and lower reaches of Chaudière and Etchemin valleys, ice contact stratified deposits have structures indicating southward paleocurrents. This suggests that northward glacial flow and melting of a remnant Appalachian ice mass was followed by limited southward expansion and readvance of Laurentide ice up major trans-Appalachian valleys, an event of regional climatic significance.

Résumé

La cartographie des formations en surface de la région de Saint-Joseph-de-Beauce au Québec, a suscité beaucoup d'intérêt pour l'établissement d'un inventaire des stries du socle et des directions des paléocourants observées dans les matériaux stratifiés de contact glaciaire. Les résultats préliminaires indiquent que des stries tardiglaciaires dirigées vers le nord et très répandues, plus récentes que des stries faites par les glaces laurentidiennes et dirigées vers le sud-est dans les régions adjacentes du Maine et du Québec, sont tout aussi bien développées dans les hautes terres et les interfluves de la région de Saint-Joseph-de-Beauce. Cependant, dans les portions moyenne et inférieure des vallées des rivières Chaudière et Etchemin, les dépôts stratifiés de contact glaciaire présentent des structures qui indiquent l'existence de paléocourants dirigés vers le sud. Ces observations semblent indiquer que l'écoulement glaciaire vers le nord et la fonte des masses glaciaires résiduelles des Appalaches ont été suivis d'une expansion limitée vers le sud et d'une nouvelle avancée des glaces laurentidiennes en amont des grandes vallées trans-appalachiennes, événement qui a joué un rôle important du point de vue du climat régional.

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INTRODUCTION

Mapping of the surficial geology of the St-Joseph-de-Beauce map area (1:50 000, NTS 21 L/7) was completed in 1988. Expanded from preliminary reconnaissance mapping carried out by N.R. Gadd and Pierre LaSalle in the 1960s, this project completes mapping of the upper and central Chaudière Valley (Gadd, 1964, 1978; Shilts, 1981). The study area occupies a critical geographical spot in southeastern Quebec for both economic and academic reasons (Fig. 1). It contains areas of high gold potential in glacial and preglacial sediments and an ophiolite belt that has been explored for platinum group elements. Geochemical or overburden drilling exploration for both these commodities depends on an adequate, up-to-date interpretation and map of surficial sediments (Shilts and Smith, 1988).

Indicators of ice flow directions and paleocurrents in ice contact deposits were mapped systematically using airphoto interpretation of the surficial geology, building on observations made by Gauthier (1975) and Lortie (1976), and investigation of several well exposed stratigraphic sections and cores from deep boreholes in and near Rivière des Plantes in the southern part of the map area (Shilts and Smith 1986a,b, 1988; Smith and Shilts 1987). The results of this research will be expanded upon in an MSc thesis to be completed by the first author.

PRELIMINARY RESULTS

Surficial deposits in the St-Joseph-de-Beauce area consist mainly of thin, stony, olive-grey, weathering tan till draped over rolling bedrock hills. In Chaudière Valley, till is thicker, but bedrock outcrops are still common. In both Chaudière and Etchemin valleys, massive deposits of ice contact sand and gravel stand as prominent hills or terraces. These deposits are thought to be subaqueous outwash facies, mainly on the strength of their occurrence well below 396 m[1], which is the approximate elevation of the main outlet eastward to Saint John River valley of a series of lakes that were dammed by glaciers each time they entered the valleys of the northward flowing Chaudière and its tributaries (Shilts, 1981). In addition to the massive ice contact deposits at Vallée-Jonction (Fig. 2), there are several other geomorphically less prominent deposits of poorly sorted gravel, diamictons, and well sorted sand and gravel in bands across Chaudière Valley as far south as St-Martin (Fig. 3). In every one of these deposits in Chaudière Valley, current measurements indicate discharge southward, opposite the direction of present drainage. In the middle reaches of Etchemin Valley, current directions are similarly southward, indicating, as in Chaudière Valley, that they were deposited from an ice mass lying to and advancing from the north.

In upper Etchemin Valley and on the interfluves east and west of Chaudière Valley, ice contact deposits commonly have structures indicating northward discharge (Fig. 3). Likewise, the abundant striae on outcrops flanking the valleys, and at some places in the valleys themselves, are oriented predominantly toward the north; in places the northward oriented striae are superimposed on southeastward oriented striae.

[1] A lower col at 305 m could carry water westward to the Saint-François drainage system thence by a complicated route through Lake Champlain to Hudson River. This col is still considerably higher than the surface of the ice-contact deposits in Chaudière Valley (213 m at Vallée-Jonction).

DISCUSSION

This juxtaposition of striae in the Estrie-Beauce region has long been known (Lamarche, 1971; Gadd et al. 1972; Lortie and Martineau, 1987). It is attributed to the formation of a late glacial ice divide in the Appalachians east of Rivière Saint-François. The ice divide developed as the result of the creation of a saddle in the ice sheet over St. Lawrence Valley in response to drawdown of ice into marine water encroaching up the lower St. Lawrence estuary (Gadd et al., 1972; Thomas, 1977). This saddle cut off the southward to southeastward flow of Laurentide ice across the St. Lawrence and Appalachians. Till deposited by this southeastward flow contains Precambrian and unmetamorphosed fossiliferous Paleozoic erratics derived from within and north of the St. Lawrence Lowlands. Development of the saddle caused ice flow to reverse toward the St. Lawrence, thus accounting for northward trending striae being superimposed on southeastward trending striae.

Notwithstanding this erosional evidence on outcrops that flank it, the southward current directions of ice-contact sediments in Chaudière Valley and the elevations of the lowest cols in the Chaudière drainage basin suggest that the last glacial event was an ice lobe that advanced up the valley in a deep lake. The lobe would represent a readvance from a body of Laurentide ice that lay in the St. Lawrence Lowlands. Similar lobes may have projected up other major trans-Appalachian valleys, such as the Saint-François and Etchemin. However, the fact that northward paleocurrent directions are present in the ice contact deposits of upper Etchemin Valley suggests that once that part of the valley was deglaciated by southward retreat of northward-flowing ice, it was not reoccupied by glaciers.

The results of this and previous studies indicate that the paradox of stratified sediments deposited by southward flowing glaciers lying in major valleys surrounded by evidence of late glacial northward flow is explained best by the readvance theory. At several places in Chaudière and tributary valleys, deposits of clayey till lying on intensely deformed lacustrine sediments were observed to rest directly on typical Lennoxville Till as defined by McDonald and Shilts (1971). Taken together these sites may provide depositional evidence for a readvance.

CONCLUSION

In the Saint-Joseph-de-Beauce map area the understanding of a complex history of ice movements has been augmented by elucidation of the details of Lennoxville glaciation, based largely on bedrock striations and paleocurrent data from ice contact stratified deposits. Recent work by Shilts and Smith (1986a,b, 1988) has confirmed the pre-Lennoxville Appalachian stratigraphy first discussed by McDonald and Shilts (1971). From the present and earlier studies, a picture of at least three glaciations has emerged. Deposits of each

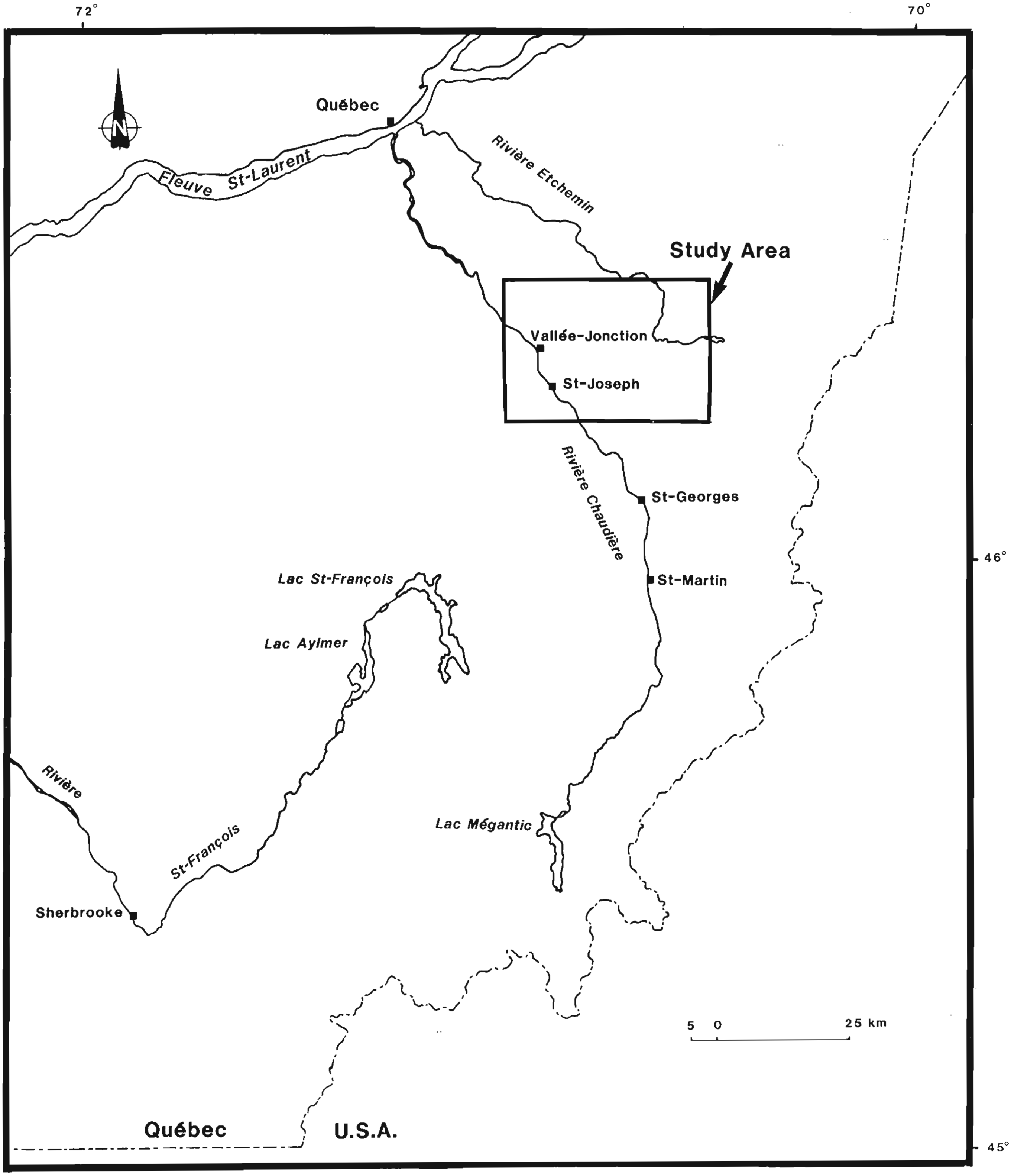

Figure 1. Location map of Saint-Joseph-de-Beauce map area.

glaciation with accompanying interstadial or interglacial beds are well represented in the Rivière des Plantes sections at the south border of the Saint-Joseph-de-Beauce map area (Shilts and Smith, 1987; Poliquin, 1987; Paul, 1987). From oldest to youngest, they are:

1) Johnville glaciation: Glaciers crossed the study area southeastward from a Laurentide source.

2) Chaudière glaciation: Glaciers crossed the study area first west-southwestward from an Appalachian ice cap to the east, shifting to flow southeastward from a Laurentide ice cap.

3) Lennoxville glaciations: Glaciers crossed the study area first southeastward from a Laurentide ice cap, flowing across the Appalachians to the New England coast (Shilts, 1976, 1981). During the general deglaciation of New England and eastern Quebec, ice south of St. Lawrence River flowed northward to northeastward from a local ice divide developed in response to drawdown along the axis of St. Lawrence Valley into a marine calving bay that formed in the lower St. Lawrence estuary. After retreat of this northward flowing ice to an ice divide located near St-Georges, about 40 km south of the map area, Laurentide ice again flowed up Chaudière and Etchemin valleys in the form of tongues of thin ice advancing into deep lakes. The southward flowing lobes of ice apparently covered only the middle Chaudière and Etchemin river valleys and the lower Appalachian foothills flanking these rivers to the northwest and northeast, respectively. It is probable that other major Appalachian valleys, such as the Saint-François, were similarly affected, because the youngest ice contact deposits also show southward current directions (Banerjee and McDonald, 1975). It is not possible to say whether the readvance started from the Appalachian front south of St. Lawrence River or whether ice advanced across St. Lawrence Valley through proglacial lakes or the Champlain Sea. Because the readvance appears to postdate melting of the remnant northward flowing Appalachian ice, it probably represents at least a regional paleoclimatic event — a deterioration of climate causing expansion of the eastern Laurentide Ice Sheet.

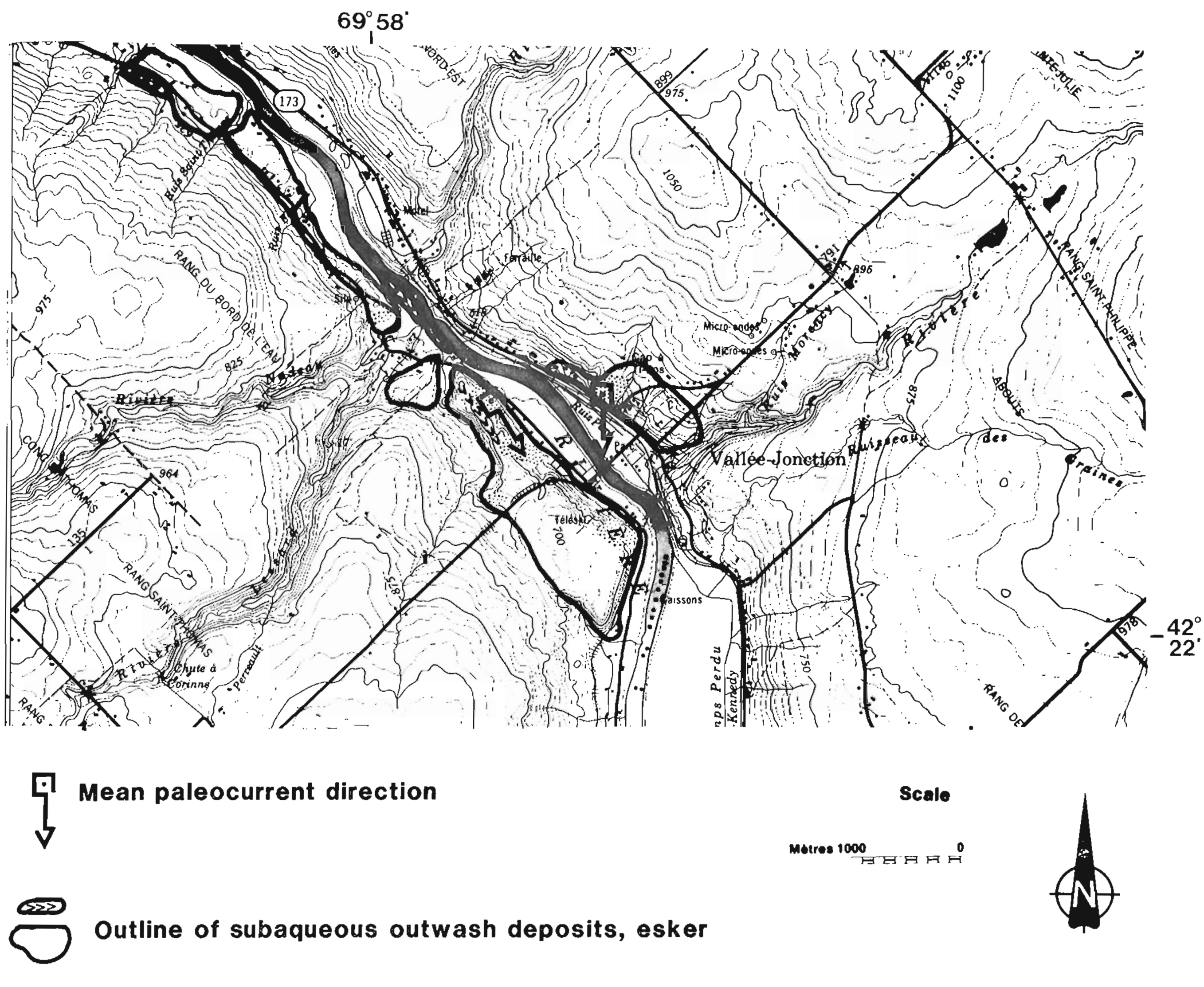

Figure 2. Ice-contact stratified drift at Vallée-Jonction, Quebec. Esker complex fed subaqueous outwash fan. Paleocurrents measured from ripple cross-laminae in medium to fine sand.

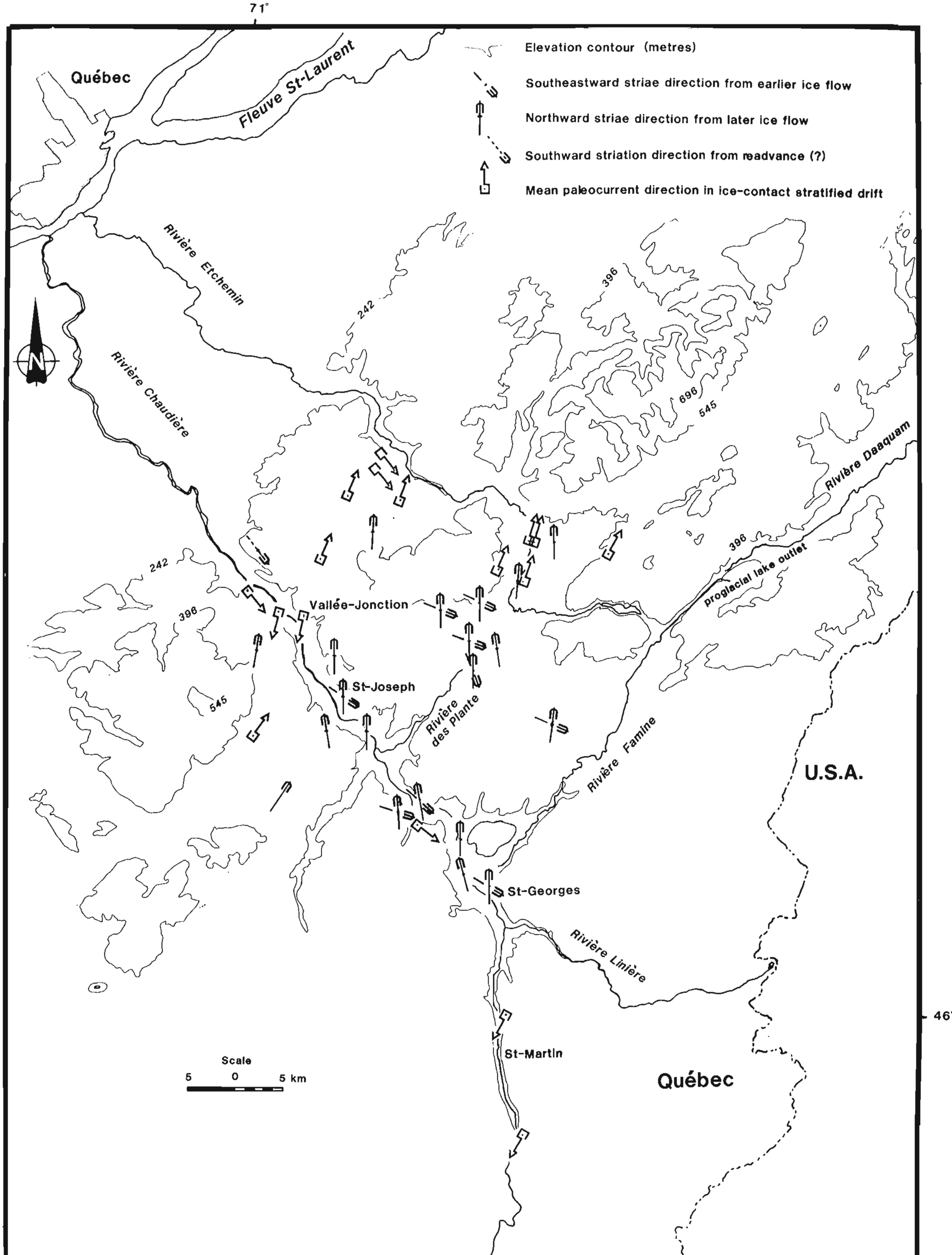

Figure 3. Selected striations and mean directions of paleocurrents in ice contact stratified drift in Saint-Joseph-de-Beauce and adjacent map areas. On many outcrops, northward striae clearly cut earlier regional southeastward striae, but no unequivocal evidence of later southeastward striations, formed during the readvance, has been found.

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Subbottom profiling of Quebec Appalachian lakes and its potential application to assessing seismic hazard

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Abstract

Subbottom acoustic profiling was carried out on four Quebec Appalachian lakes located within 100 km of major seismic zones in St. Lawrence Valley and Miramichi Highlands of New Brunswick. Two of these lakes, Lac Témiscouata and Grand Lac Squatec, are thought to have been damaged by earthquakes because:

1. *Bottom disturbance is widespread in both lakes and appears to be multicyclic;*
2. *The hummocky appearance of both resedimented and eroded sediments is similar to features found in lakes observed to be damaged during historical earthquakes;*
3. *The geographical relationship between these lakes and the Charlevoix seismic zone provides a plausible mechanism for dislodging sediment from slopes beneath the surface of the lakes.*

Résumé

Au Québec, on a réalisé des profils acoustiques du fond de quatre lacs de la région appalachienne, situés dans un rayon de 100 km à partir des grandes zones sismiques de la vallée du Saint-Laurent et des hautes terres de Miramichi au Nouveau-Brunswick. On estime que deux de ces lacs, le lac Témiscouata et le Grand Lac Squatec, ont été endommagés par des séismes, puisque:

1. *les perturbations du fond couvrent un grande surface des deux lacs et semblent être multicycliques;*
2. *l'apparence bosselée des dépôts resédimentés et érodés rappelle des structures découvertes dans des lacs ayant subi des dommages observés durant les séismes de la période historique; et,*
3. *la relation géographique entre ces lacs et la zone sismique de Charlevoix peut agir à titre de mécanisme plausible responsable du décollement des sédiments de pentes situées au-dessus de la surface des lacs.*

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INTRODUCTION

Subbottom acoustic profiling was carried out on three lakes that occupy major linear topographic depressions transverse to the northeast-trending structure of the eastern Quebec Appalachian mountains. These lakes, along with a smaller, more equidimensional lake that was also surveyed, are located within 100 km of two major seismic zones in St. Lawrence Valley (Fig. 1). The principal objectives of the surveys were to determine the style and history of sedimentation in these basins.

Profiling was accomplished with a Raytheon, RTT-1000A-1 "Portable Survey System" operating with dual 7kH$_z$ and 200 kH$_z$ transducers, deployed from an inflatable 5 m boat. Clear, high-resolution profiles were obtained by slow traversing (<5 km/h) coupled with chart speeds of 2.5 to 5 cm/minute. Vertical exaggeration on the resulting profiles averages about 20 ± 5. Navigation was by "dead reckoning", from headland to headland. A total of 220 line km of profile were surveyed — 103 km on Lac Témiscouata, 30 km on Grand Lac Squatec, and 87 km on Lac Matapédia.

RESULTS

Lac Matapédia

Lac Matapédia (Fig. 1), located about 40 km southeast of the Lower St. Lawrence seismic zone (Adams and Basham, in press), is typical of the many lakes that occupy linear depressions with axes that trend at right angles to the structure and topographic grain of the northeastern Appalachian Mountains. The lake is steep-sided with a gently undulating bottom underlain by several metres of acoustically nonlaminated sediment (Fig. 2), interpreted elsewhere as postglacial organic sediment or gyttja (Klassen and Shilts, 1982). At various places along the lake, the bottom is interrupted by symmetrical ridges, thought to be segments of the discontinuous esker that can be seen to enter the lake at its south end. On none of the 87 km of profiles surveyed on this lake was there any evidence of sediment disturbance. In this respect the lake is similar to almost all of the more than 150 lakes that we have profiled on the Canadian Shield and in the Appalachians over the past eight years.

Lac Etchemin

Lac Etchemin is a relatively shallow lake with sides notably less steep than those of Lac Matapédia. It lies about 100 km south of the Charlevoix seismic zone (Adams and Basham, in press; Fig. 1). Its bottom is mostly underlain by acoustically nonlaminated sediment similar to the upper sediment in Lac Matapédia. Like Lac Matapédia, the sediment shows no signs of disturbance at the surface, but in one part of the lake, a subbottom feature resembling a mud flow extends down the south side of the basin and a short distance up the north side (Fig. 3). The sediment mass is completely buried by overlying, acoustically nonlaminated sediment and has no expression on the present lake bottom. If this is a mass wasting deposit, it probably was formed at the onset of postglacial sediment accumulation and cannot be related to any specific cause. This example is presented mainly to demonstrate how ancient slump features may be detected within the sediment column using the acoustic profiling system.

Lac Témiscouata

Lac Témiscouata is a long (~30 km), low elevation lake lying along the Trans-Canada highway, midway between Rivière-du-Loup, Quebec and Edmunston, New Brunswick, about 70 km east of the Charlevoix seismic zone (Fig. 1). It is steep sided and flat bottomed and was deglaciated by southward retreating ice that flowed northward from an ice flow centre in New Brunswick. The central part of the lake bed is almost totally covered by hummocky deposits that appear to be made up of sediment slumped from the sides (Fig. 4, 5, 6). Like similar deposits in lakes observed to have been shaken by earthquakes, such as Lac Tee, Quebec (Shilts, 1984) and Comox Lake, British Columbia (Clague et al., 1989), these slumped sediments are either acoustically opaque or impede penetration of the acoustic signal into underlying sediment. In the northern third of the lake, acoustically parallel laminated sediments, interpreted to have been deposited in a proglacial lake expanding in front of the southward retreating ice front, are generally undeformed. In places the parallel laminated sediments are cut by channels, and a considerable volume of sediment can be seen to have been removed (Fig. 7). This is clearly demonstrated by profile 87-TC-32 (Fig. 7) which was made during a north to south traverse along the axis of the northern part of the lake. The channels were cut by an agent, presumably a sediment flow, moving at right angles to the axis of the lake.

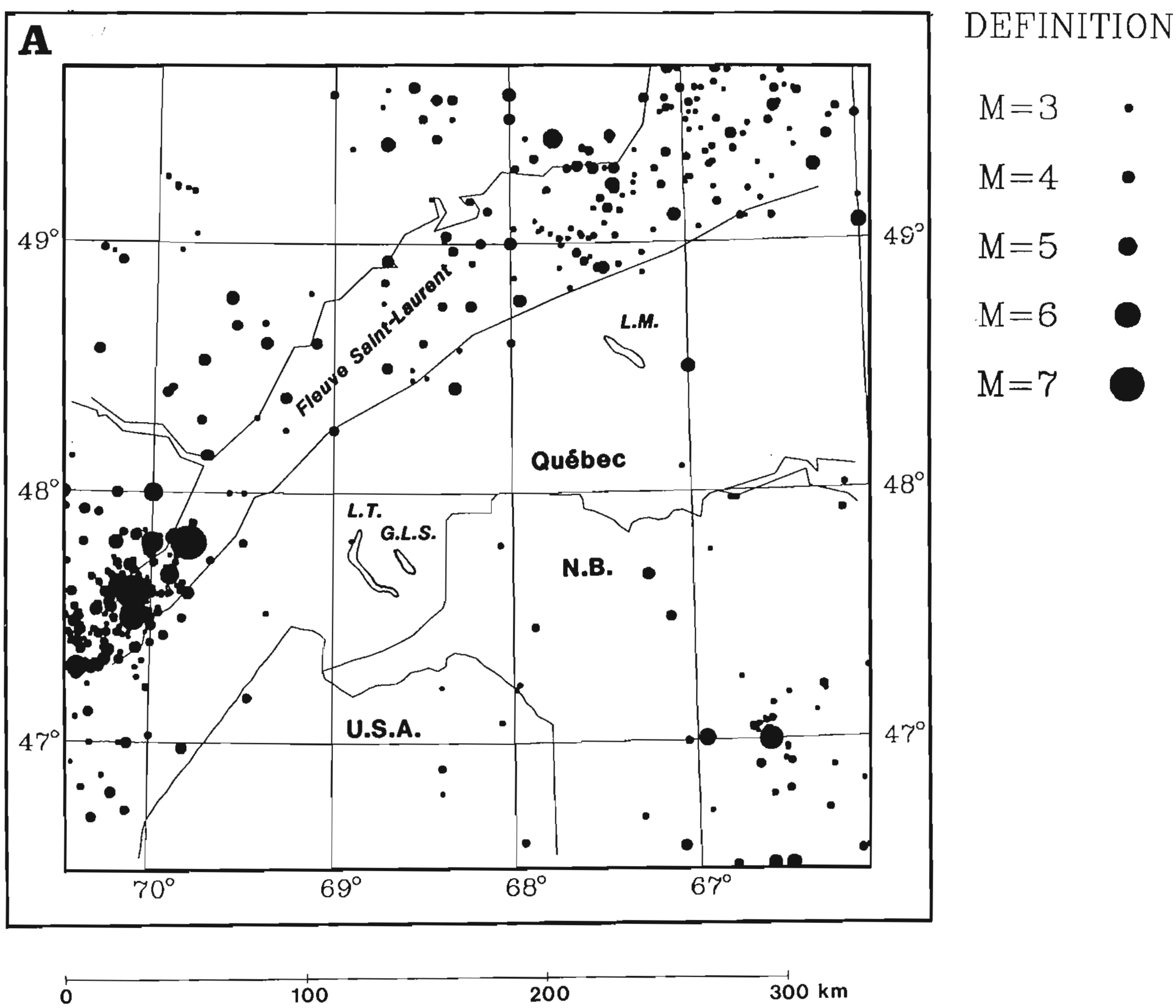
Grand Lac Squatec

Grand Lac Squatec is located about 15 km east of the south end of Lac Témiscouata (Fig. 1). It has bottom morphology and sediment characteristics similar to both Lac Matapédia and Lac Témiscouata, and its late glacial history is similar to that of Lac Témiscouata. Many profiles surveyed on Grand Lac Squatec showed signs of sediment erosion and chaotic deposition (Figures 8, 9). This suggests that whatever caused the disturbance of the bottom of Lac Témiscouata, may have caused similar disturbance in Grand Lac Squatec.

DISCUSSION

One of the principal problems in using sediment disturbance in lakes as an indicator of past seismic activity is distinguishing earthquake-induced slumping from slumping associated with late glacial sedimentation processes, normal shoreline processes, or hydrological processes. In an earlier report, the first author incorrectly attributed massive slumping in Lac Mégantic, Quebec, a basin bathymetrically and structurally similar to those discussed here, to seismic activity (Shilts, 1984); subsequent to publication of that report, an eyewitness described how an oversteepened sand spit had failed, with photographic evidence of the undisturbed spit, in September of 1946. This failure produced the distinctive slump features seen on the bottom of Lac Mégantic.

Nevertheless, the contrast between the undisturbed Holocene sediment in Lac Etchemin and Lac Matapédia on one hand and the disturbed sediment in Lac Témiscouata and Grand Lac Squatec on the other, suggests that the latter pair have been subjected to processes not recorded in the former.



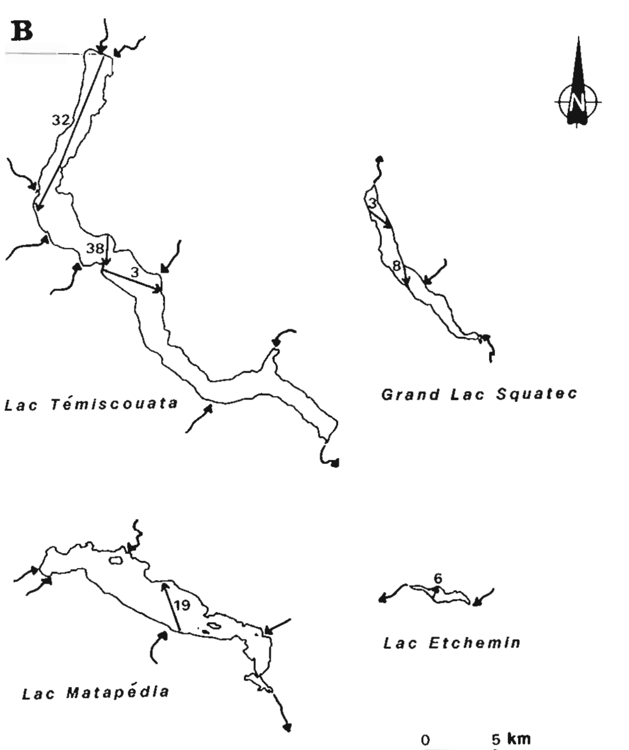


Figure 1. 1A. Location of lakes and earthquakes, three seismic zones are evident: Charlevoix at left center, Lower St. Lawrence at upper right, and Miramichi at lower right. L.T. = Lac Témiscouata; G.L.S. = Grand Lac Squatec; L.M. = Lac Matapédia; Lac Etchemin is just below lower left corner. 1B. Location of subbottom profiles discussed. Lac Témiscouata and Grand Lac Squatec are in true spatial relationship.

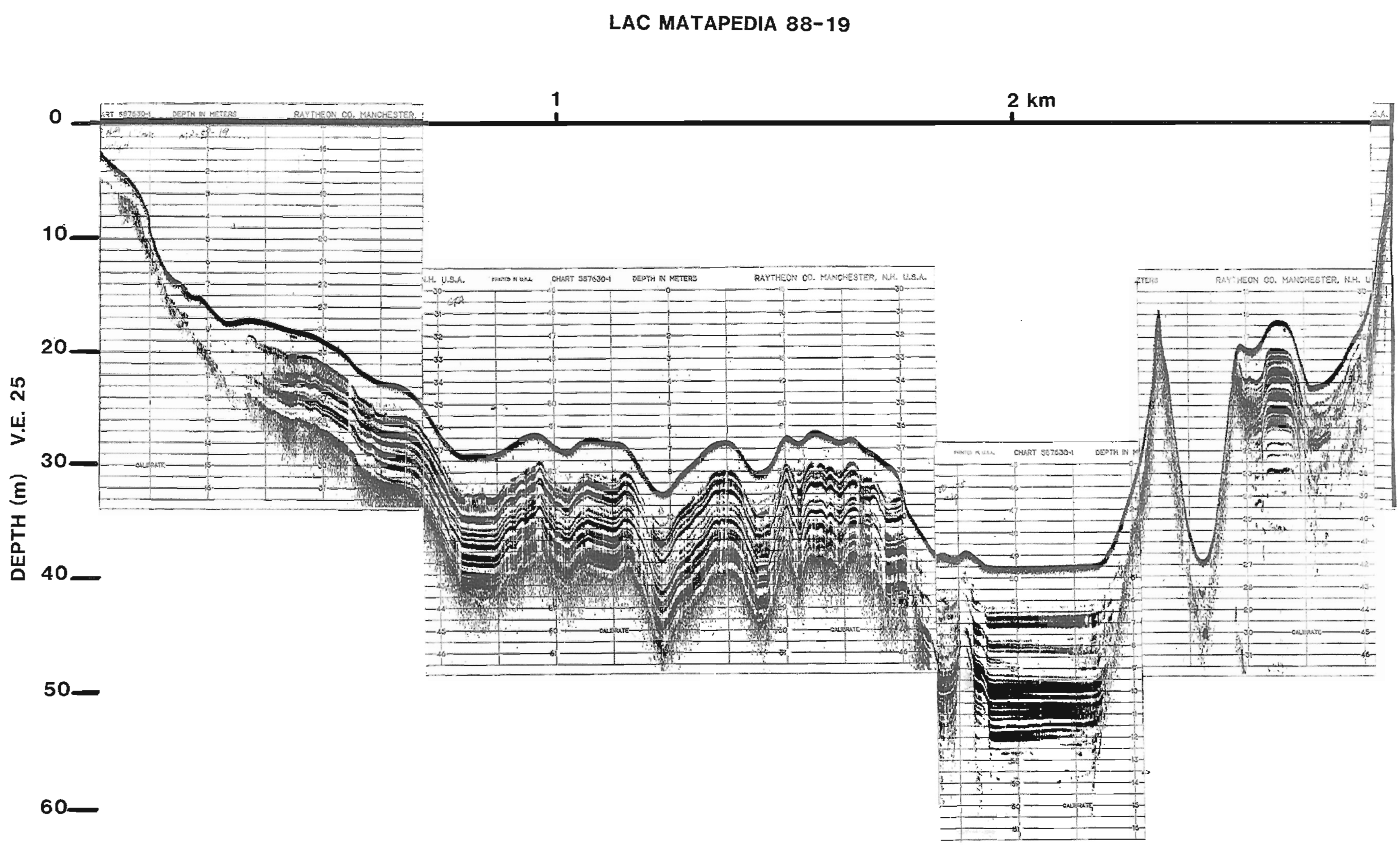


Figure 2. Undisturbed, acoustically parallel laminated sediment overlain by acoustically ''clear'' modern sediment in profile typical of Lac Matapédia. Sharp ''spike'' at about 2.3 km is an esker which can be traced on land. Note the undisturbed gyttja on the flanks of the esker. Slope of the esker flanks is 15°-17°.

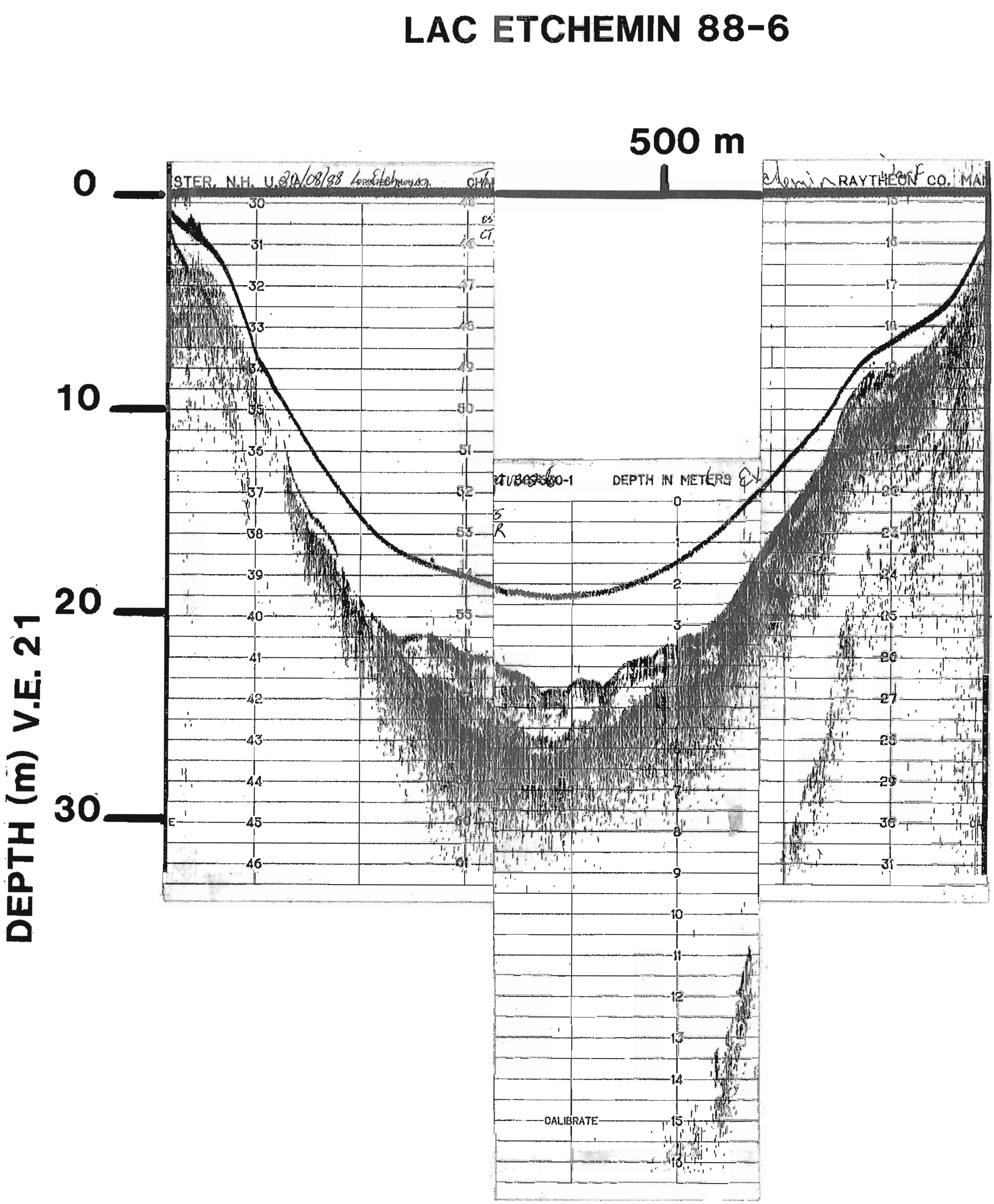


Figure 3. Late or immediately postglacial slump deposit buried beneath acoustically clear gyttja in Lac Etchemin. Direction of slumping was from right to left. Lat. 46°23'10'' Long. 70°30'10'' start of profile; Lat. 46°23'35'' Long. 70°29'30'' end of profile.

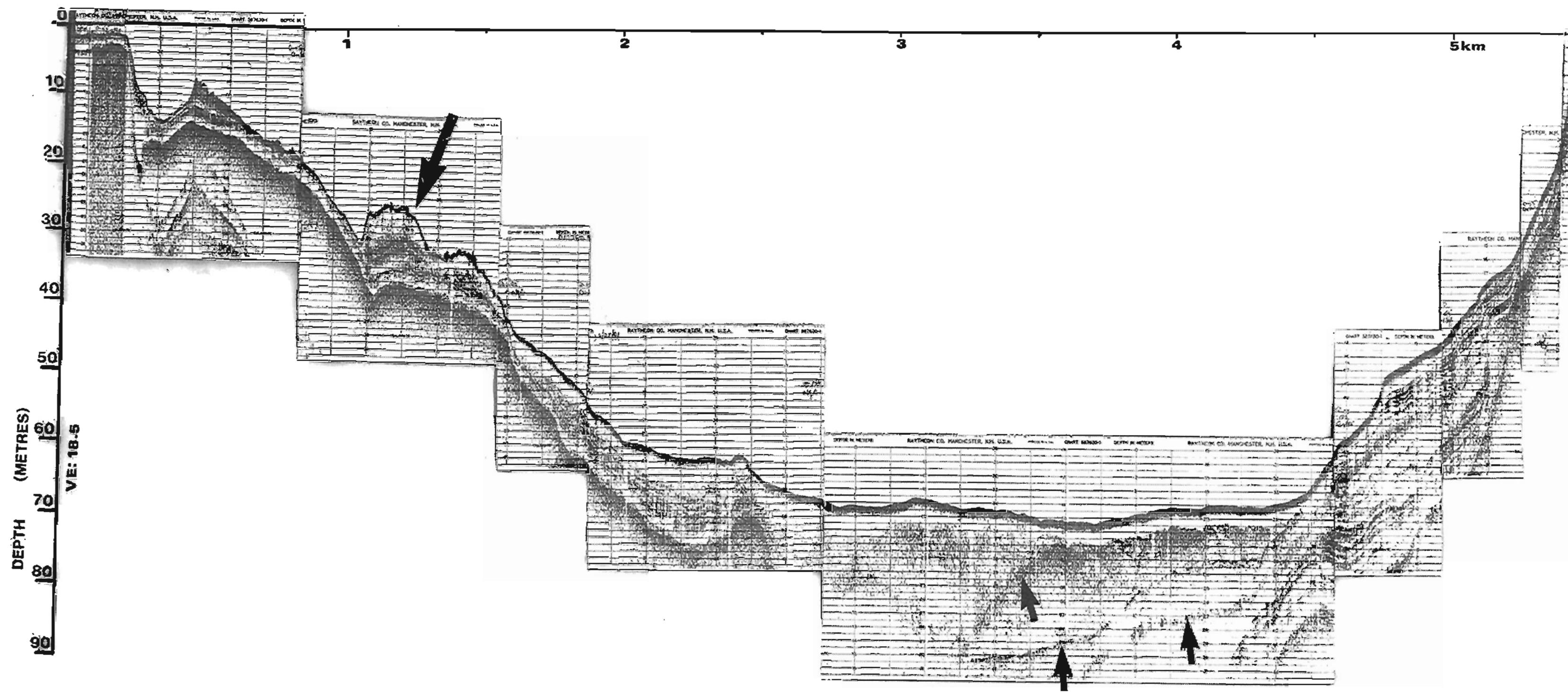


Figure 4. Profile showing typical hummocky bottom in central Lac Témiscouata. Small arrows point to buried hummocky erosional or depositional surfaces that may mark tops of old, seismically induced, slumped deposits. Large arrow points to disturbed (eroded or slumped), acoustically laminated sediment.

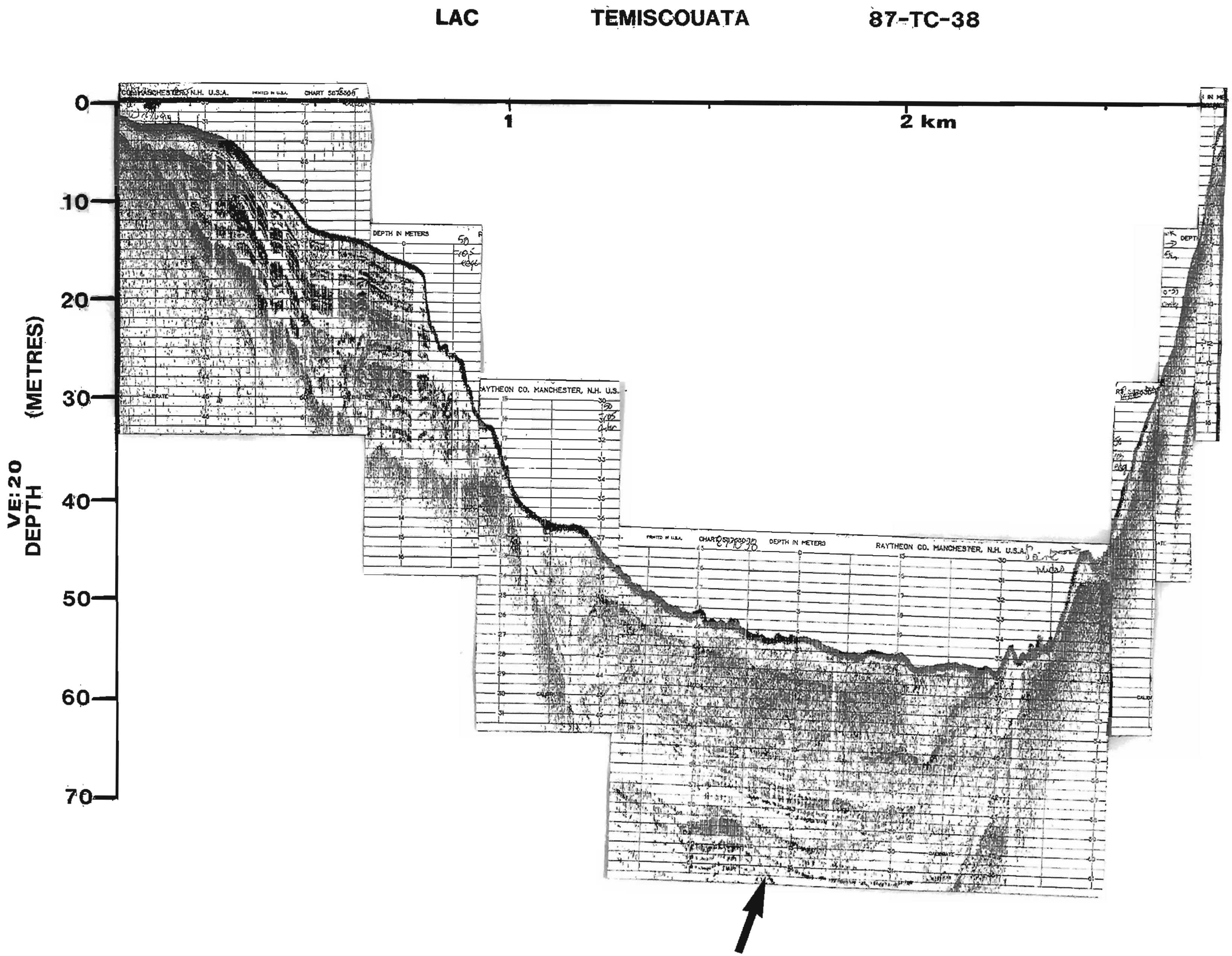


Figure 5. Hummocky slump deposits overlying acoustically parallel laminated sediments. At least two periods of slumping are represented. Origin of angular unconformity between parallel laminated sequences (arrow) not known.

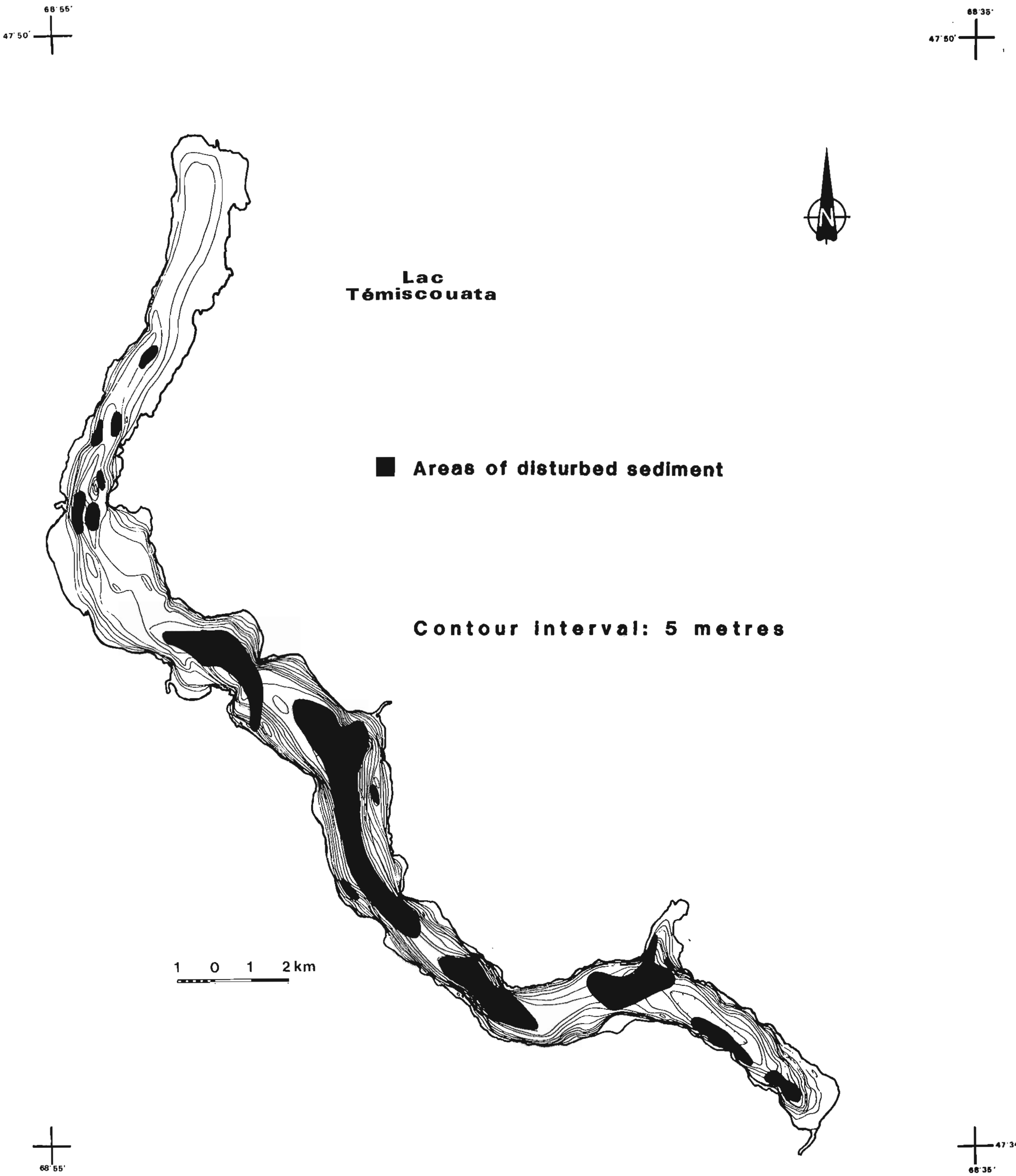


Figure 6. Map of areas where slumping has disrupted bottom sediments in Lac Témiscouata.

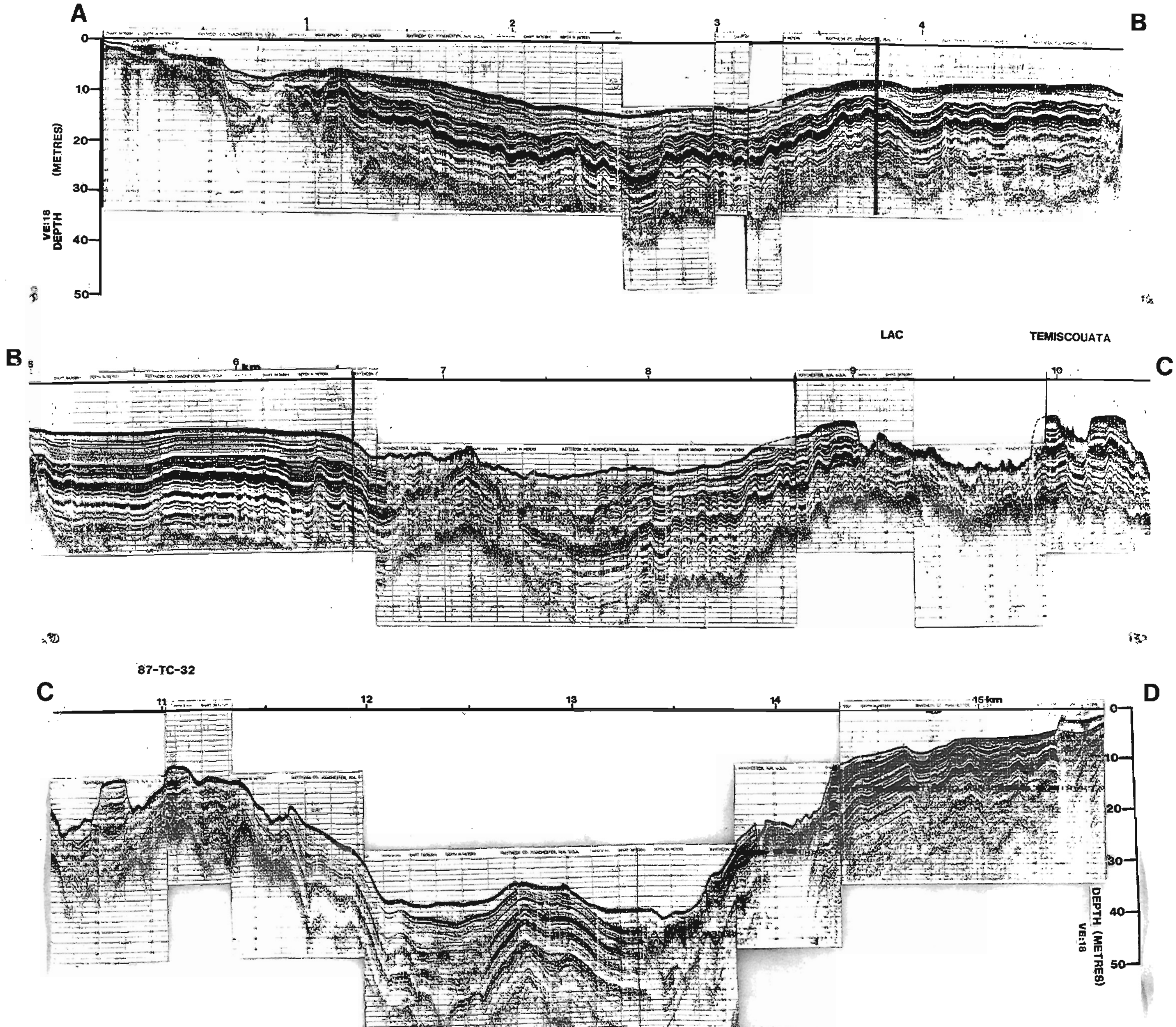


Figure 7. Longitudinal profile through northern third of Lac Témiscouata, from north to south. Hummocky slump deposits lie on undistrubed, parallel laminated sediment between 7-8 km. From 9-11 km and at 14 km, parallel laminated sediment is eroded and disturbed, presumably as a result of seismically induced slumping from basin sides.

Furthermore, except for a few examples of postglacial sediment disturbance related either to factors such as shoreline sedimentation, described above for Lac Mégantic, or to unknown late glacial causes (Larocque, 1987), features similar to those observed on the bottoms of Lac Témiscouata and Grand Lac Squatec have been observed only in lakes that have suffered visible disturbance during historical earthquakes (e.g., Shilts, 1984; Clague, et al., 1989).

Large earthquakes are presently far more common near Lac Témiscouata and Grand Lac Squatec (Fig. 1) because they are located near the Charlevoix seismic zone, which has been the site of at least five historical earthquakes of magnitude 6, or greater (1663, 1791, 1860, 1970, and 1925; Smith, 1962; Adams and Basham, in press). The 1663 and 1925 earthquakes were of magnitude 7; the 1925

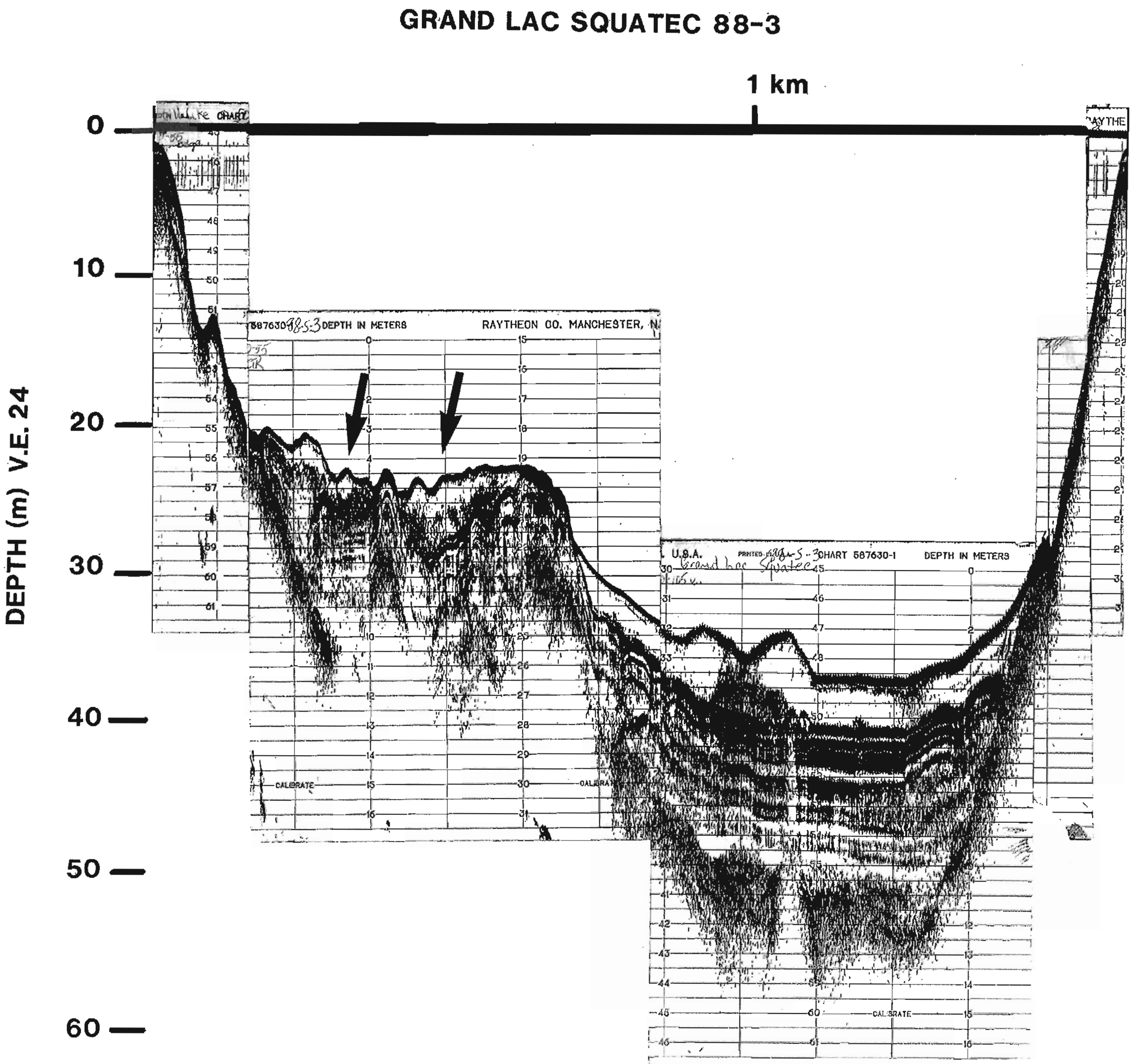


Figure 8. Slumping in Grand Lac Squatec (arrows). Two large hummocks on acoustically clear sediment in deep basin 1 km to right of arrows are similar to side echos from masses of slumped sediment generated in 1946 earthquake in Comox Lake, Vancouver Island, B.C. (see Clague et al., 1989).

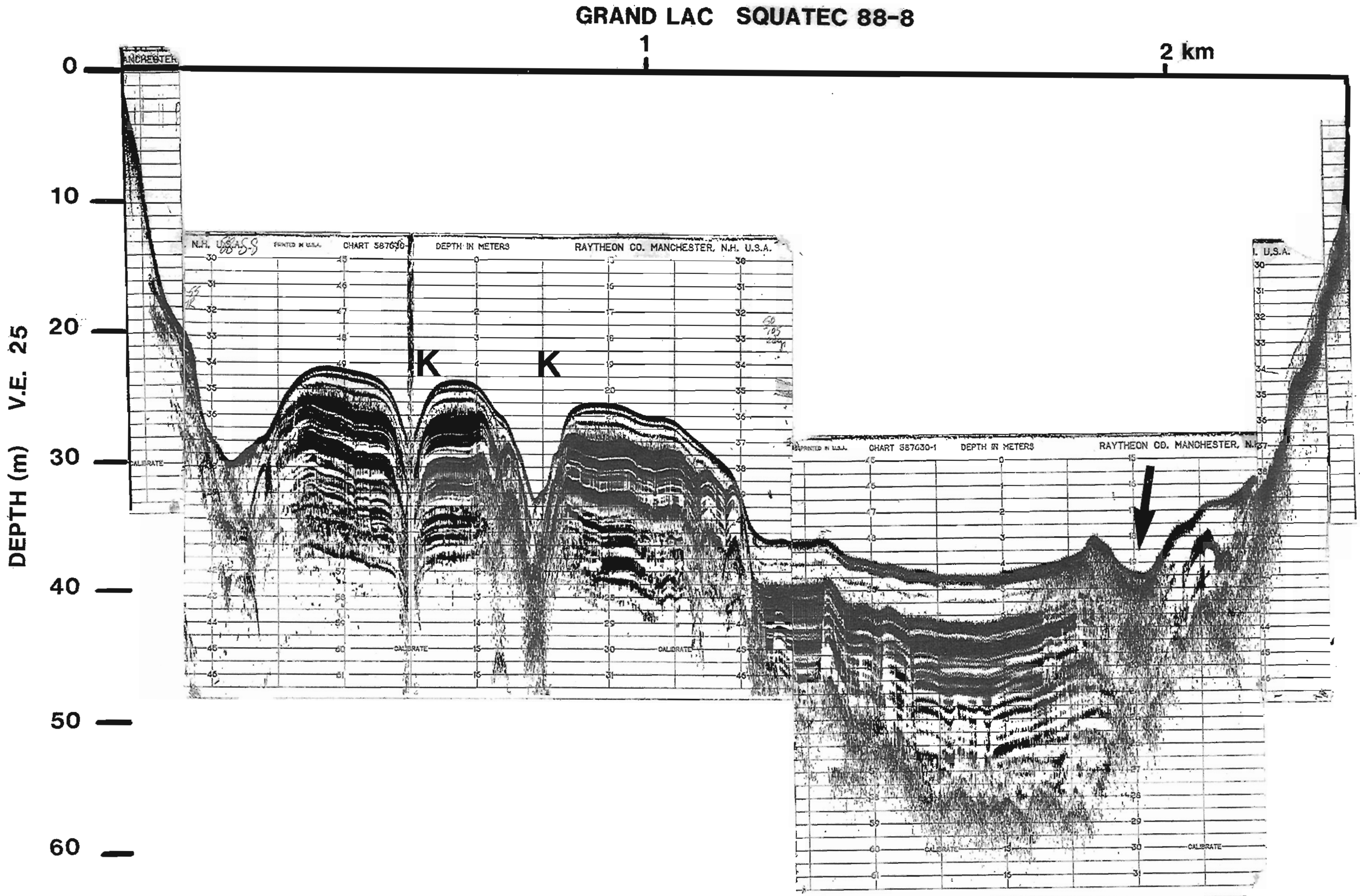


Figure 9. Hummocky slump deposits and sediment deformation in Grand Lac Squatec (arrow). Sharp depressions in parallel laminated sediment are sections through kettle holes (k) caused by collapse over buried, melting glacier ice.

event produced Modified Mercalli Intensities (MMI) of VI to VII at Lac Témiscouata, but only about V at Lac Matapédia (Smith, 1962). Lac Témiscouata and Grand Lac Squatec are also relatively close to the Miramichi Highlands of central New Brunswick, the site of four strong earthquakes in 1982 that produced MMI IV at Lac Témiscouata.

Sims (1973) suggested that earthquakes generating MMI $\geq$ VI were required to produce even slight sediment deformation in Van Norman reservoir in California. Thus, it is unlikely that the strongest recorded earthquakes in the region (1663, 1925 at Charlevoix) would have caused sediment disturbance in Lac Matapédia. By contrast, the intensity of the 1925 shaking at Lac Témiscouata and Grand Lac Squatec (MMI > VI) might well have been sufficient to cause liquefaction and slumping.

The slumping and sediment disruption on the bottom of Lac Témiscouata is presently interpreted to have been caused by one or more large postglacial earthquakes. This conclusion is based largely on the extensive area of the lake floor involved in slumping (Fig. 6), the exceptionally thick acoustically unlaminated sequences of sediment in the deepest basins, and the presence of buried, hummocky surfaces in acoustically nonlaminated sediments (Fig. 5). It should be emphasized again that such sediment disruption is extremely rare in the lakes that we have profiled, being confined almost exclusively to lakes that were chosen for profiling because of their known disturbance by historical earthquakes.

Was the slumping in Lac Témiscouata caused by the 1925 or earlier Charlevoix earthquakes? At this point we know of no documentation of changes in the character of Lac Témiscouata after the 1925 earthquake. Suspended sediment was reported in Lac Tee, Quebec after the magnitude 6.2 earthquake of 1935 (Hodgson, 1936; Shilts, 1984). It may be possible to detect resedimented clay layers in gyttja cores using methods developed by Doig (1987) to date the slump events in Lac Témiscouata. Multiple slump events were recently documented by Doig (1987) in Lac Tadoussac, adjacent to the Charlevoix seismic zone. A remaining possibility is that the disturbances might have been caused by an earthquake close to Lac Témiscouata, even though the current level of seismicity is very low (Fig. 1).

Although Lac Matapédia is close to the seismically active Lower St. Lawrence seismic zone (Adams and Basham, in press), this zone has not been known to have produced any earthquakes larger than magnitude 5 over the past 100 years. Although larger earthquakes probably do occur in the Lower St. Lawrence seismic zone, the present evidence from Lac Matapédia — especially the lack of slumping of gyttja from the esker sides — suggests that the lake basin may not have been shaken by a magnitude 7 or larger earthquake in the last 10 000 years. Such a conclusion, if verified, would place a strong constraint on the size of the largest earthquakes to be expected from the Lower St. Lawrence seismic zone, and hence improve future hazard estimates.

Finally, a set of seismic hazard calculations, equivalent to those developed for the National Building Code of Canada, were made for each of the lakes discussed in this paper (Lac Témiscouata and Grand Lac Squatec were considered together). For an annual probability of 10 % in 50 years, the peak horizontal accelerations expected are 12 %g for Lac Matapédia, 16 %g for Lac Etchemin, and 21 %g for Lac Témiscouata. These estimates are in line with the relative amounts of disturbance observed in the lakes.

Because of the potential of this highly portable acoustic subbottom profiling system for providing high resolution records of sediment conditions in small to moderate sized lakes, we intend to continue to survey lakes of the Canadian Shield and Appalachians with an eye toward improving our understanding of the various processes causing slumping. The expansion of this activity into the Canadian Cordillera (Clague et al., 1989) has and will increase our ability to discern among various causes of slumping, yielding, hopefully, a technique that we can apply confidently to improve our understanding of the geological record of prehistoric earthquakes and hence the estimation of seismic hazard.

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