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Commission géologique du Canada

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CURRENT RESEARCH PART B
EASTERN AND ATLANTIC CANADA

RECHERCHES EN COURS PARTIE B
EST ET RÉGION ATLANTIQUE DU CANADA

1988

Canada

GEOLOGICAL SURVEY OF CANADA
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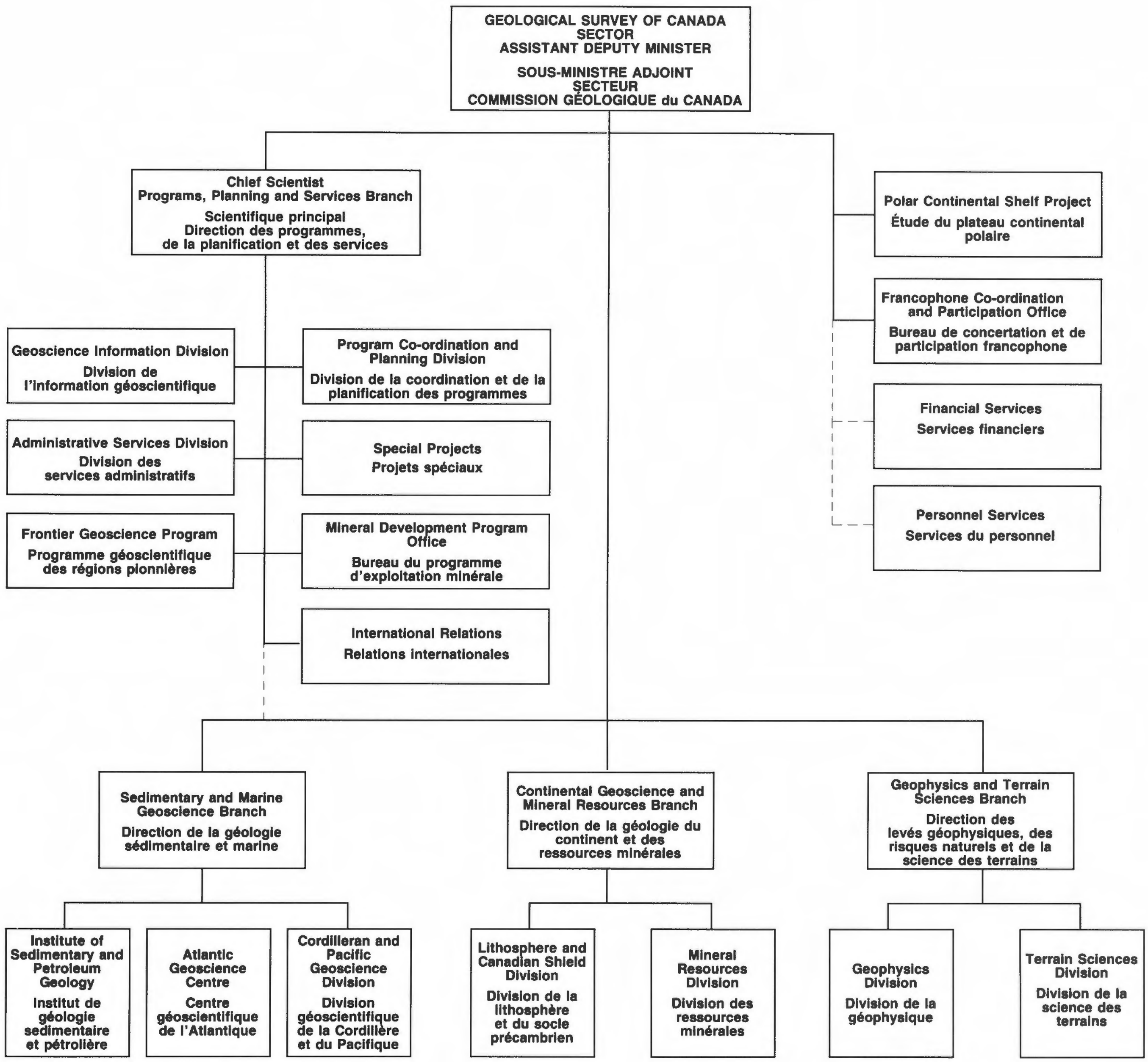
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INTRODUCTION

In 1987 the Geological Survey of Canada became a Sector within the Department of Energy, Mines and Resources, and was re-organized into the four Branches shown on the accompanying organizational chart. The primary role of the Survey, which was founded in 1842, continues to be to provide an overview of all facets of Canadian geology as a basis for national policy, for planning by government and industry, and for public information.

In order to provide interim results of its program a publication titled "Summary of Research" was initiated in 1963. The title was changed to "Current Research" in 1978 and the report was released three times a year (Part A, B and C). After 1982 Part C was no longer issued and Part B was discontinued in 1987 to encourage greater use of journal publication for short contributions.

Current Research, however, is the one series of GSC publications that gives the public a yearly overview of the range of the Geological Survey of Canada Sector activities. From time to time Current Research has been criticized for its size, as it was necessary for the user to buy a large volume to obtain a few pertinent papers. To introduce greater flexibility, this issue of Current Research is therefore available in six parts that can be purchased separately: four regional volumes, one volume of national and general programs, and a volume that contains abstracts of all the reports. The Parts are:

- Part A: Abstracts/Résumés
- Part B: Eastern and Atlantic Canada
- Part C: Canadian Shield
- Part D: Interior Plains and Arctic Canada
- Part E: Cordillera and Pacific Margin
- Part F: National and General Programs

Identification of the Parts by letters is for convenience only and may be subject to change. Each of Parts B to F includes a paginated Table of Contents for the volume: Table of Contents for the other Parts of this series will be found at the back of each volume.

En 1987 la Commission géologique est devenue un secteur à l'intérieur du ministère de l'Énergie, des Mines et des Ressources et a été réorganisée en quatre directions indiquées sur l'organigramme d'accompagnement. Organisme fondé en 1842, la Commission a comme rôle principal de procurer un cadre d'ensemble de toutes les facettes de la géologie du Canada comme base d'une politique nationale pour la planification du gouvernement et de l'industrie et pour informer le public en général.

Afin de fournir les résultats préliminaires de son programme de recherche une publication ayant titre « Summary of Research » est apparue en 1963. Une nouvelle publication, ayant les mêmes buts, est apparue en 1978 sous le titre « Recherches en cours »; cet ouvrage était diffusé trois fois par année (parties A, B et C). Après 1982 la partie C a été abandonnée et ce fut de même pour la partie B en 1987. L'arrêt de ces publications avait pour but d'adopter une nouvelle forme d'édition pour satisfaire davantage l'usager.

La publication « Recherches en cours » appartient à part entière à la série des publications de la CGC et apporte à chaque année une vue d'ensemble des activités de la Commission géologique maintenant au niveau de secteur. De temps à autre la publication « Recherches en cours » a été critiquée pour son fort volume, plusieurs ont constaté qu'il était nécessaire d'acheter un gros volume uniquement pour n'avoir accès qu'à un petit nombre d'articles. Maintenant, cette publication est disponible en six parties en vente séparément, ce qui procure une plus grande flexibilité pour l'usager. La publication est répartie comme suit: quatre volumes régionaux, un volume couvrant les programmes nationaux et généraux et un dernier contenant les résumés de tous les articles. On y trouve les parties suivantes:

- Partie A: Abstracts/Résumés
- Partie B: Est et région atlantique du Canada
- Partie C: Bouclier canadien
- Partie D: Plaines intérieures et région arctique du Canada
- Partie E: Cordillère et marge du Pacifique
- Partie F: Programmes nationaux et généraux

L'identification des parties par une lettre a été adoptée uniquement par commodité; on pourra éventuellement utiliser une autre façon. Les parties B à F possèdent une table des matières paginée; il est à noter qu'à chacune des parties de cette série on y trouvera à l'endos une table des matières indiquant le contenu des autres parties.

Répartition du Cr, Pt, Pd, et Ir dans les dépôts de surface de l'Estrie-Beauce, Québec[1]

Y.T. Maurice
Division des ressources minérales

Maurice, Y.T., Répartition du Cr, Pt, Pd, et Ir dans les dépôts de surface de l'Estrie-Beauce, Québec; dans Recherches en cours, partie B, Commission géologique du Canada, Étude 88-1B, p. 1-8, 1988.

Résumé

Des analyses chimiques de concentrés de minéraux lourds alluvionnaires, récoltés sur environ 9000 km² de la région de l'Estrie-Beauce, révèlent des traînées de dispersion de chrome qui proviennent de l'érosion glaciaire de roches du complexe ophiolitique de la bande de Serpentine, une des unités géologiques majeures qui traverse la région échantillonnée. Les traînées les mieux définies sont longues de plusieurs dizaines de kilomètres et sont orientées vers le sud-est, direction d'écoulement du dernier glacier d'importance (Lennoxville) à parcourir le sud du Québec.

Il existe une étroite relation entre le Pd et les traînées de chrome en provenance des régions de Thetford Mines et de Saint-Joseph. Le rapport Pd/Cr est sensiblement plus élevé dans le secteur de Saint-Joseph, ce qui pourrait indiquer un potentiel plus élevé pour des minéralisations en éléments du groupe du platine (EGP) dans cette région. Plusieurs autres anomalies d'EGP pointent vers différentes zones de roches ultrabasiques où il pourrait y avoir des concentrations intéressantes en ces métaux. Une traînée d'échantillons très anomaux en Pt, Pd et Ir semble débuter dans la partie nord du granite de Saint-Sébastien. Quoique nous n'ayons pas encore confirmé la présence de concentrations anomales d'EGP dans les roches de cette région, cette anomalie pourrait signaler la présence d'un type de minéralisation d'EGP dont on ne soupçonne pas l'existence en Estrie-Beauce.

Abstract

Chemical analyses of alluvial heavy mineral concentrates, collected in a 9000 km² area of the Eastern Townships, show chromium dispersal trains derived from glacial erosion of ophiolitic rocks contained in the Serpentine belt, a major geological unit that crosses the area. The better-defined trains are several tens of kilometres long and are oriented towards the southeast, parallel to the flow direction of the last major glacial ice sheet (Lennoxville) to overrun southern Quebec.

Pd shows a close association to the chromium dispersal trains originating in the Thetford Mines and Saint-Joseph regions. The Pd/Cr ratio is significantly greater in the Saint-Joseph area, which could indicate a higher potential for platinum group elements (PGE) mineralization in that region. Several other PGE anomalies point towards various areas of ultrabasic rocks where significant concentrations of these metals may occur. One dispersal train, which contains samples that are highly anomalous in Pt, Pd and Ir, seem to originate from the northern part of the Saint-Sébastien granite. Although no anomalous PGE concentrations have yet been found in the rocks of that region, this anomaly could indicate the presence of an unsuspected type of PGE mineralization in the Eastern Townships.

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[1] Contribution aux mesures fédérales relatives à l'amiante (1984-1987)

INTRODUCTION

Depuis 1984, la Commission géologique du Canada effectue des levés géochimiques de minéraux lourds alluvionnaires dans les régions de l'Estrie et de la Beauce du sud du Québec. L'extraction des minéraux lourds s'effectue au moyen d'appareils mécanisés à partir de volumes considérables de gravier (de l'ordre de 200 litres), afin de s'assurer que les échantillons soient représentatifs, particulièrement en ce qui concerne les métaux rares. À date, environ 16 000 km^2 ont été couverts par ces levés à un taux de prélèvement moyen d'un échantillon par 12.5 km^2 (figure 1).

Les échantillons sont analysés par un laboratoire commercial pour une trentaine d'éléments. Initialement, la liste n'incluait pas les éléments du groupe du platine (EGP) mais en raison de l'intérêt croissant de la part de plusieurs sociétés minières qui recherchent ces métaux dans les roches ultrabasiques de l'Estrie-Beauce, nous avons jugé à propos d'y ajouter le Pt, le Pd et l'Ir.

Notons que les roches ultrabasiques de l'Estrie-Beauce ne sont généralement pas considérées comme favorables aux minéralisations d'EGP (e.g. Crocket, 1981). Cependant, Gauthier et Trottier (1987), après avoir effectué plusieurs analyses d'EGP sur des chromitites de la région de Thetford Mines, concluent que ces roches pouraient revêtir un certain potentiel encore non identifié.

Dans le présent rapport, il ne sera question que des résultats de la campagne d'échantillonnage de 1985 (figure 1). Les échantillons cueillis en 1984 n'ont pas été analysés pour les EGP et les résultats de ceux récoltés en 1987 ne sont pas disponibles au moment de la préparation de ce rapport. Les méthodes de prélèvement des échantillons dans les cours d'eau et de séparation des minéraux lourds ont été décrites en détail par Maurice et Mercier (1985b et 1986). Les méthodes et résultats d'analyses (sauf pour le Pt et le Pd), ainsi qu'une interprétation de ces résultats ont également fait l'objet de divers rapports (Maurice et Mercier, 1985a; Maurice, 1985, 1986a et b). L'objet de ces travaux est de cartographier la répartition des métaux analysés, dans le but de faire ressortir des patrons de dispersion qu'on associe à des processus géologiques et géochimiques et, ultimement, à des sources de métaux dans les roches. Dans certains cas, ces patrons ou anomalies constituent des cibles d'exploration intéressantes.

GÉOLOGIE

La région de l'Estrie-Beauce se situe dans la province structurale des Appalaches et se subdivise en bandes sédimentaires et volcano-sédimentaires dont les âges varient du Protérozoïque supérieur (Hadrynien) au Dévonien inférieur. Ces bandes sont orientées en direction nord-est et comportent des unités miogéosynclinales et eugéosynclinales, pour la plupart légèrement métamorphisées. Certaines de ces formations ont été pénétrées par des intrusifs granitiques vers le milieu du Dévonien (figure 1).

L'unité qui nous concerne principalement dans ce rapport est la bande de Serpentine qui renferme un complexe ophiolitique alpin. Cette bande, que l'on peut retracer de façon intermitente à partir de Terre-Neuve jusqu'au lac Brompton près de la frontière internationale dans le sud du Québec, est interprétée comme une suture océanique résultant de la fermeture de l'océan Iapétus durant l'Ordovicien inférieur (Laurent, 1980). La séquence ophiolitique comporte un membre inférieur composé surtout de harzburgite, superposé de cumulats ultrabasiques et mafiques à la base desquels on retrouve des dunites et des dunites-chromifères. Ces dernières donnent lieu par endroit à des chromitites massives à semi-massives, à caractère podiforme à plus ou moins stratifié. Vers le haut, la séquence se poursuit avec d'abord des pyroxénites, suivies de gabbros, de laves basiques et d'argillites. Selon Crocket (1981), ce sont les gabbros à cumulats, en particulier ceux qui contiennent des sulfures magmatiques juste au dessus des cumulats ultramafiques, qui offrent les meilleures chances d'être enrichis en Pt et Pd.

Dans la région de Thetford Mines, la séquence ophiolitique est à peu près complète mais ailleurs le long de la bande de Serpentine, on n'en retrouve que des parties démembrées. Notons également qu'on retrouve des roches ophiolitiques à l'extérieur de la bande de Serpentine comme, par exemple, le dyke de Pennington (unité 2, figure 1) au nord-est de Thetford Mines, qui aurait été mis en place le long de failles de chevauchement dans les schistes Cambriens de la bande des monts Sutton/Notre-Dame (Laurent, 1975).

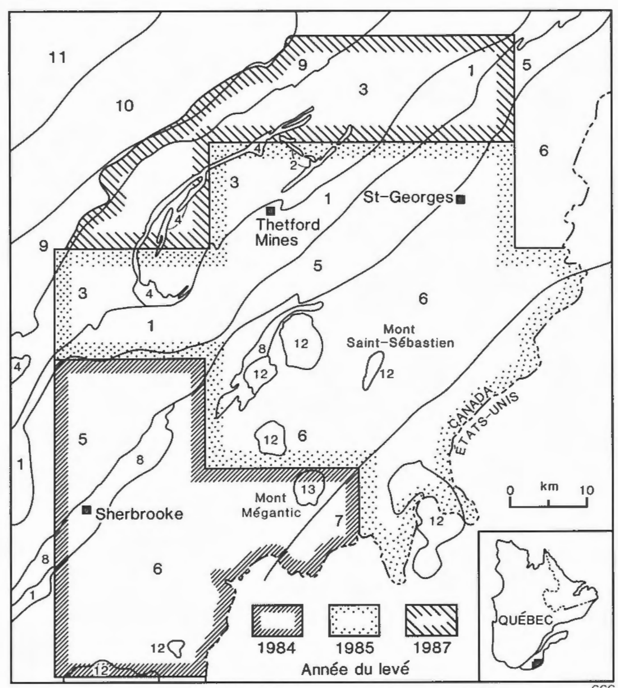


Figure 1. Géologie régionale et localisation des travaux de 1984, 1985 et 1987. Géologie selon Harron (1976): 1: bande de Serpentine; 2: dyke de Pennington; 3: bande des monts Sutton/Notre-Dame; 4: volcaniques du Tibbit Hill; 5: synclinorium de Saint-Victor; 6: synclinorium de Gaspé-Connecticut Valley; 7: anticlinorium de Boundary Mountain; 8: bande des monts Stoke; 9: bande flyschoïde; 10: klippes; 11: basses-terres du Saint-Laurent; 12: granites dévoniens; 13: intrusion alcaline montérégienne.

GÉOLOGIE DU QUATERNAIRE

La région de l'Estrie-Beauce a subi au moins trois avancées glaciaires au cours du Wisconsinien et chacune est représentée par un till séparé du suivant par des dépôts lacustres ou fluviatiles. Le plus ancien de ces tills et le plus récent, respectivement connus sous les noms de Johnville et Lennoxville, ont été déposés par des glaciers qui ont parcouru la région en direction sud-est (McDonald et Shilts, 1971). Le till intermédiaire, ou Chaudière, aurait été déposé par un glacier s'écoulant vers l'ouest ou le sud-ouest à partir d'un centre situé dans les Appalaches. Il existe cependant une controverse au sujet de l'ampleur de ce mouvement glaciaire: Shilts (1978) et Shilts et Smith (1987) prétendent que le mouvement vers le sud-ouest a été interrompu par un glacier laurentidien en provenance du nord-ouest. Selon ce point de vue, la partie supérieure du till Chaudière proviendrait de sources situées au nord-ouest tout comme les tills Johnville et Lennoxville. De son côté, Parent (1987) est d'avis que la glace a parcouru la région d'est en ouest durant toute la durée de l'épisode Chaudière.

Vers la fin du Wisconsinien, avec la formation d'une baie de vêlage dans la vallée du Saint-Laurent et l'isolement d'une masse glaciaire sur les terrains élevés de l'Estrie-Beauce, il s'est produit une inversion de l'écoulement glaciaire vers le nord qui a affecté la moitié nord de la région couverte par nos travaux (Lortie et Martineau, 1987). Ce mouvement est responsable d'un système de stries et nervures (crag-and-tail) que l'on retrouve en abondance dans les régions de Thetford Mines et d'Asbestos (Lamarche, 1971 et 1974).

DISPERSIONS GLACIAIRES

Les travaux de McDonald et Shilts (1971) et Shilts (1973) sur la géologie et la géochimie des dépôts quaternaires de l'Estrie-Beauce ont illustré l'importance de la glaciation comme agent de dispersion des matériaux meubles dans la région. La présente campagne d'échantillonnage démontre que la géochimie des minéraux lourds alluvionnaires reflète la composition de ces dépôts de surface et permet de confectionner des cartes d'éléments qui illustrent remarquablement bien les dispersions glaciaires.

Les cours d'eau choisis pour les prélèvements des minéraux lourds sont généralement de premier ou de second ordre, de sorte que la composition de leurs alluvions s'apparente à celle des dépôts glaciaires sur lesquels ils coulent. Certes, les forces hydrauliques et l'action chimique associées au ruissellement ont pu modifier quelque peu les proportions entre les différentes phases minéralogiques, mais nos données montrent que ces processus n'ont pas masqué les tendances géochimiques régionales clairement attribuables à la glaciation.

La figure 2 montre la répartition du chrome dans la région échantillonnée en 1985. On note une large zone de concentrations élevées (> 10 % Cr) au sud et sud-est du complexe ophiolitique de Thetford Mines, qui s'étend jusqu'à la frontière internationale. On note également une interruption des valeurs élevées au nord-est de cette zone et une reprise de celles-ci au nord de la rivière Chaudière, qui correspond à des terrains situés en aval glaciaire de l'amas de roches ultrabasiques situé au sud-est de Saint-Joseph. La zone de valeurs basses, entre les deux zones chromifères, forme un couloir d'environ 25 km de large, orienté vers le sud-est. Cette zone croise la bande de Serpentine le long d'un segment qui est apparamment dépourvu de roches ultrabasiques, quoiqu'elle intersecte plus au nord le dyke de Pennington.

Une traînée de valeurs très élevées en chrome (> 20 % Cr jusqu'à 50 % Cr) débute dans la partie la plus massive du complexe ophiolitique de Thetford Mines, plus précisément dans le canton de Coleraine, endroit bien connu pour ses gisements de chromite. Cette traînée s'étend sur plus de 50 km au sud-est de sa source et possède une largeur moyenne de 15 km (figure 2). On note une zone plus petite de concentrations supérieures à 20 % Cr à l'intérieur de la zone de valeurs élevées (> 10 % Cr) au nord de la rivière Chaudière et quelques autres moins importantes réparties à l'intérieur de la zone principale.

Il ne fait aucun doute, d'après la répartition régionale du chrome, que les minéraux lourds récoltés à partir des dépôts alluvionnaires reflètent surtout la répartition de la chromite dans les dépôts glaciaires du Lennoxville. Il est probable, cependant, qu'il existe une certaine composante de dispersion vers l'ouest ou le sud-ouest correspondant à l'épisode Chaudière, mais celle-ci est difficile à isoler sur la figure 2. On doit s'attendre normalement à ce qu'une partie des dépôts de l'épisode Chaudière ait été repris par le glacier Lennoxville pour produire des dispersions de débris vers le sud-est, mais déplacées à l'ouest par rapport à leurs sources. Aussi, certains des cours d'eau échantillonnés sont probablement suffisamment encaissés dans les dépôts glaciaires pour que leurs sédiments soient composés en partie de matériaux déposés par le glacier Chaudière. L'examen des cartes de distribution de la trentaine d'éléments analysés laisse entrevoir dans quelques cas, une composante que l'on peut interpréter comme appartenant à la glaciation Chaudière, mais les patrons les plus prononcés sont toujours attribuables à l'épisode Lennoxville. Ceci ne signifie pas que l'on doit ignorer les effets de la glaciation Chaudière lorsqu'on interprète nos données. Comme on le verra à la section suivante, certaines anomalies, rendues faibles par la dilution mais non moins importantes, peuvent être expliquées en fonction d'une dispersion Chaudière ou, le plus souvent, d'une combinaison Chaudière-Lennoxville.

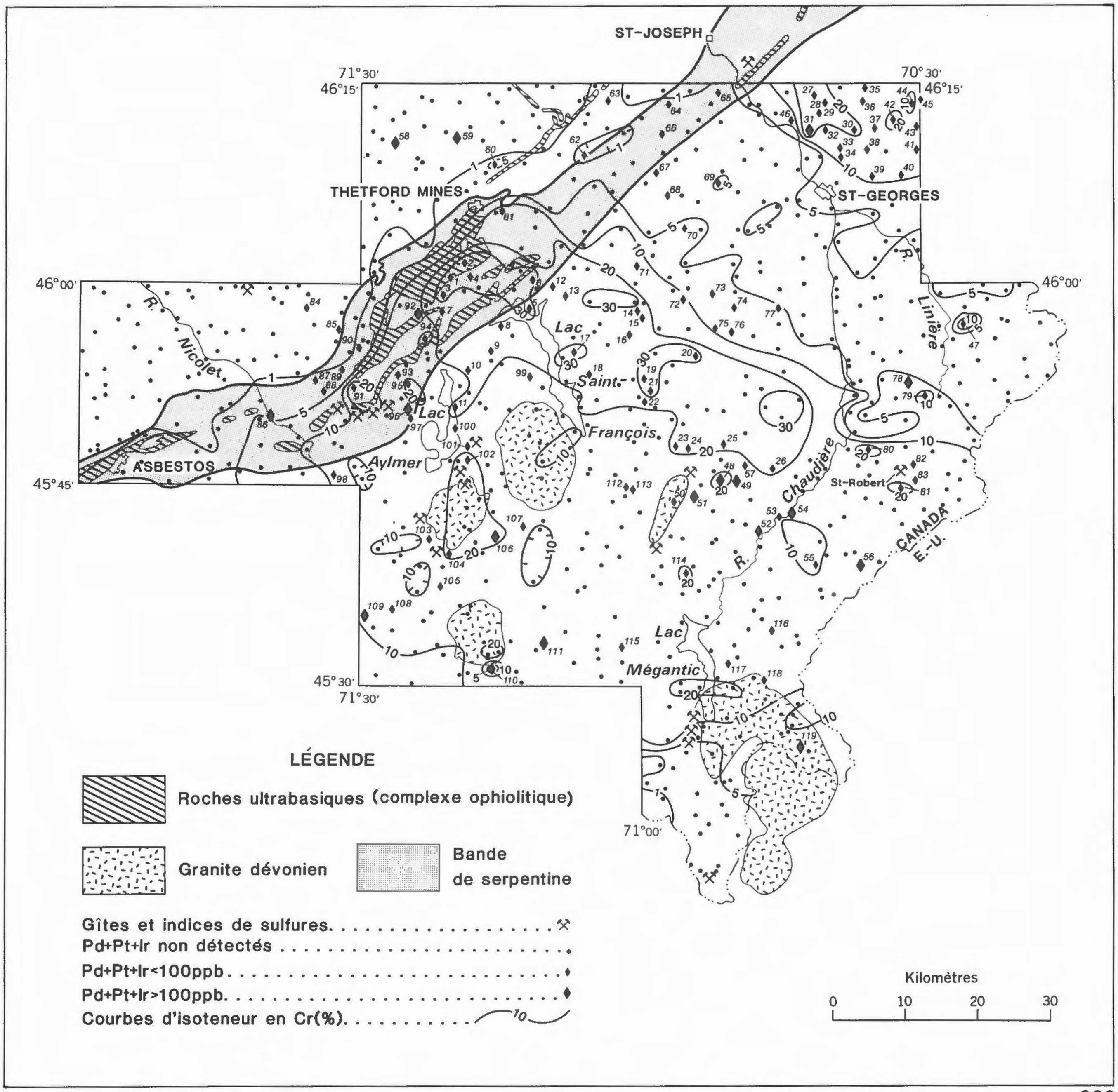
Le mouvement glaciaire tardif vers le nord ne semble pas avoir entraîné de matériaux du complexe ophiolitique à en juger par les basses teneurs en chrome au nord de la bande de Serpentine dans les régions de Thetford Mines et d'Asbestos. Cette constatation confirme les observations de Shilts et Smith (1987).

RÉPARTITION DU Pt, DU Pd ET DE L'Ir

La figure 2 localise tous les échantillons qui contiennent des quantités détectables de Pt, Pd ou d'Ir. L'appendice A donne les concentrations de ces éléments dans chaque échantillon numéroté à la figure 2. Ces données ont été obtenues par l'analyse des mêmes échantillons de minéraux lourds qui furent utilisés pour le dosage du Cr. L'Ir a été analysé simultanément au Cr par activation neutronique de 10 g d'échantillon pulvérisé sans préconcentration par pyroanalyse (irradiation directe). Pour le Pt et le Pd, on a effectué une concentration par pyroanalyse de 5 g d'échantillon suivie d'une mesure au plasma à courant continu, en se servant d'une fraction de l'échantillon différente de celle utilisée pour l'analyse de l'Ir et du Cr. Étant donné que les EGP peuvent se retrouver sous forme de petites particules métalliques libres ou pépites, et que ces pépites ne sont pas distribuées uniformément dans l'échantillon même après broyage, l'analyse de deux portions différentes de l'échantillon peut expliquer pourquoi nous avons enregistré dans certains échantillons des concentrations anormales en Ir et non en Pd et Pt, et vice-versa. Normalement, on s'attendrait de retrouver les trois éléments ensemble.

Les limites de détection analytique sont de 100 ppb pour l'Ir, de 50 ppb pour le Pt et de 10ppb pour le Pd. La figure 2 et l'appendice A montrent tous les échantillons qui contiennent des teneurs supérieures à ces limites de détection.

Le Pd, avec sa limite de détection analytique considérablement plus basse que celles des deux autres éléments, nous permet d'observer la répartition des EGP là où ils sont présents en concentrations très faibles. C'est le cas par exemple des deux traînées de chrome, celle de Thetford Mines et celle située au nord de la rivière Chaudière. Un bon nombre des échantillons contenant plus de 20 % Cr dans



CGC

Figure 2. Répartition du chrome et de certains éléments du groupe du platine (EGP) dans les dépôts alluvionnaires (minéraux lourds) de la région échantillonnée en 1985. Les numéros d'échantillons correspondent à ceux inscrits à l'appendice A.

la traînée de Thetford Mines et plus de 10 % Cr dans celle au nord de la rivière Chaudière, contiennent des quantités détectables de Pd.

Il est fort probable que dans nos échantillons, le Pd se trouve dans la chromite même. Oshin et Crocket (1982) ont remarqué que les EGP s'associent de préférence à la chromite plutôt qu'aux silicates ferromagnésiens dans les roches du complexe ophiolitique de Thetford Mines. Ils indiquent également que dans les chromites provenant des amas de chromitites à olivine (comme celles des gisements de chromite du canton de Coleraine) les EGP se trouvent à l'intérieur des grains de chromite, tandis que dans les harzburgites, ils se situent plutôt en bordure des grains.

Il est intéressant de constater que le rapport Pd/Cr est d'environ trois fois plus élevé dans la traînée au nord de la rivière Chaudière que dans celle de Thetford Mines si on ne tient compte que des échantillons avec plus de 20 % Cr et en ignorant ceux de la traînée de Thetford qui contiennent moins de Pd que la limite de détection analytique (tableau 1).

Près de l'extrémité sud-est de la traînée de chrome de Thetford Mines et légèrement au sud de celle-ci, proviennent six échantillons dont les teneurs totales des trois EGP analysés varient de 374 ppb à 869 ppb. Quatre des échantillons sont enrichis en Ir tandis que les deux autres le sont en Pt et Pd (voir appendice Ac). Cette séparation apparente des EGP peut résulter du fait que les analyses d'Ir et de Pt-Pd aient été effectuées sur différentes portions des échantillons

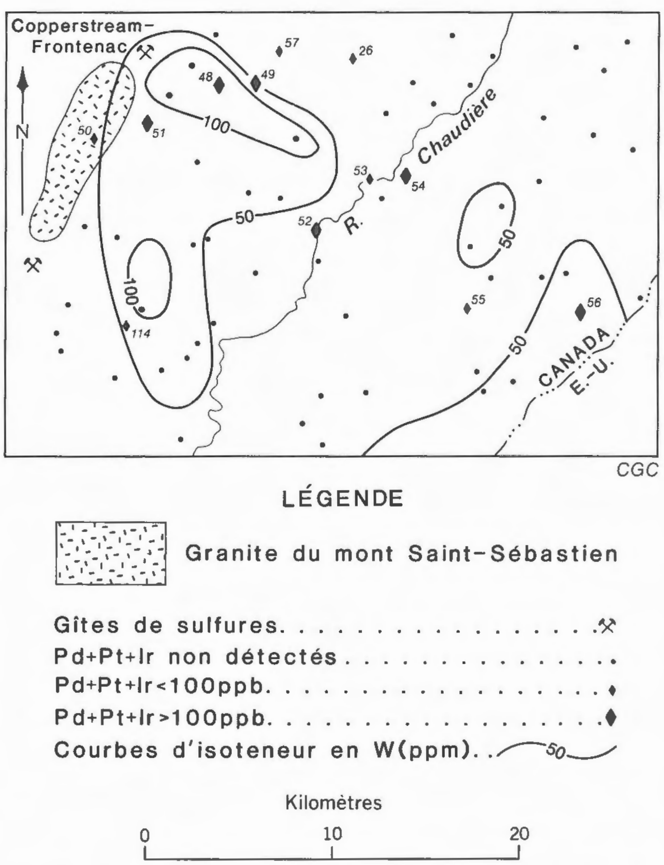


Figure 3. Traînées de tungstène et d'EGP émanant des environs du mont Saint-Sébastien.

Tableau 1. Rapports Pd/Cr des traînées de chromite de Thetford Mines et de la rivière Chaudière.

Traînée	n	*Cr % ($\bar{X}$)	**Pd ppb ($\bar{X}$)	Pd/Cr (x10^7)
Thetford Mines	26	27.8	13.7	0.5
R. Chaudière	5	23.9	35.8	1.5

*Cr>20% n = nombre d'échantillons
**Pd>10 ppb

(voir discussion à ce sujet plus haut). Les échantillons anomaux sont alignés parallèlement à la traînée de chrome ce qui porte à croire que leur répartition résulte d'une dispersion glaciaire Lennoxville (figure 2).

La traînée d'EGP semble débuter à l'extrémité nord du granite du mont Saint-Sébastien, à l'endroit où affleure le gîte de mobybdène-cuivre de la Copperstream-Frontenac. Ce gisement est formé de veines de quartz minéralisées en pyrrhotine, molybdénite et chalcopyrite recoupant des cornéennes et divers types de roches granitiques et quartzofeldspathiques (Kelly, 1975). On sait que le glacier Lennoxville a érodé et transporté vers le sud-est une partie de la zone minéralisée comme en témoigne une traînée de tungstène qui semble prendre source à l'endroit du gisement (figure 3). Bien qu'on n'ait jamais rapporté la présence de tungstène associé à ce gisement, l'analyse d'un échantillon de cornéenne récolté par l'auteur a révélé une teneur de 150 ppm W. L'analyse de cet échantillon et de quelques autres minéralisés en sulfures n'a cependant pas révélé la présence d'EGP.

Un certain nombre d'échantillons enrichis en Pd (10 à 30 ppb) proviennent de la région située au sud-ouest de la rivière Chaudière, entre les deux traînées de chrome (figure 2). Quelques-uns (#70 à 77) se situent en bordure de la traînée de chrome de Thetford Mines mais les autres (#62 à 69) sont nettement à l'extérieur de celle-ci. Quoique le lien ne soit pas évident, ces échantillons se situent tous en aval glaciaire du dyke de Pennington. Nous n'avons cependant pas de données concrètes qui indiquent que les roches du dyke ont subi une érosion glaciaire importante. Néanmoins, malgré les basses teneurs en chrome (1 à 5 %) de la région au sud-ouest de la rivière Chaudière, ces teneurs demeurent plus élevées que celles du secteur au nord du dyke de Pennington. Ceci pourrait indiquer qu'il y a effectivement eu érosion glaciaire des roches du dyke et si le Pd était relié à cette composante des débris glaciaires, le rapport Pd/Cr du dyke de Pennington serait passablement plus élevé que celui des roches ultrabasiques des régions de Thetford Mines ou de Saint-Joseph, ce qui lui donnerait une certaine importance du point de vue exploration pour les EGP. De plus, l'analyse d'échantillons de minéraux lourds récoltés près du dyke, à East-Broughton, a révélé la présence d'importantes quantités de sulfures de nickel (pentlandite) dans cette région (Maurice, 1986b). Ces sulfures, comme toutes concentrations de sulfures associées à des roches ultrabasiques, pourraient être très intéressants comme cible d'exploration pour les EGP.

Il se pourrait, d'un autre côté, que le Pd et le Cr que l'on retrouve dans la zone située entre les deux principales traînées de chrome, aient été déposés initialement par le glacier Chaudière s'écoulant vers l'ouest et repris par le glacier Lennoxville (voir « Dispersions Glaciaires »). Dans ce cas, le Cr et le Pd pourraient être associés à une fraction très diluée des dépôts glaciaires provenant des roches ultrabasiques au sud de Saint-Joseph. De même, les deux valeurs élevées, une en Ir (#58 = 170 ppb) et l'autre en Pt (#59 = 164 ppb), signalées au nord-ouest du dyke de Pennington, pourraient représenter des vestiges d'un mouvement glaciaire vers l'ouest. Ce secteur est pauvre en chrome, comme le montre la figure 2, mais on y a rapporté plusieurs blocs erratiques de roches ultrabasiques dont on ignore la provenance exacte (voir Bouchard et al., 1987, figure 48). Ces blocs pourraient avoir été transportés vers l'ouest à partir du dyke de Pennington ou des roches ultrabasiques de la région de Saint-Joseph, ou encore vers le nord à partir des roches ophiolitiques de la région de Thetford Mines au cours de la phase tardive d'écoulement glaciaire. Les très faibles concentrations de chrome dans cette région vont, cependant, à l'encontre de la dernière de ces hypothèses.

On remarque quelques valeurs élevées en Pd et une en Pt dans les échantillons provenant des environs du gisement de Saint-Robert (#80 à 83). Ces échantillons se situent en ligne avec la traînée de chrome de Thetford Mines mais leur teneur en EGP est passablement plus élevée que celle des échantillons de la traînée de chrome ce qui suggère une source distincte. Notons que le gisement de Saint-Robert et celui de la Copperstream-Frontenac possèdent plusieurs caractéristiques communes (Harron, 1976).

Le long du flanc ouest de la zone enrichie en chrome ($Cr > 10\,\%$) on remarque plusieurs échantillons contenant des valeurs au dessus des limites de détection analytique en Pd et quelques uns en Pt (#86, 103, 110) et en Ir (#92, 96, 106, 109, 111, 119). La répartition de ces échantillons pourrait indiquer un enrichissement en EGP dans le segment de la bande de Serpentine situé à l'ouest de la zone d'ophiolites massives de Thetford Mines.

DISCUSSION

Les concentrations et la répartition du Pt, du Pd et de l'Ir suggèrent que les EGP sont présents dans les concentrés de minéraux lourds alluvionnaires de l'Estrie-Beauce sous deux formes :

1 : dans la chromite, en concentrations relativement basses mais réparties assez uniformément et,

2 : sous forme de particules de métal natif donnant des concentrations plus élevées mais beaucoup plus erratiques.

On croit que les particules d'EGP natifs dans les dépôts glaciaires pourraient provenir soit de la destruction de minéraux porteurs, en particulier des sulfures, ou avoir été libérés des espaces intergranulaires des roches ultrabasiques lors de leur désagrégation, ou encore, de l'érosion d'anciens placers platinifères.

Nos données semblent indiquer que la chromite des roches ultrabasiques de la région de Saint-Joseph contient plus d'EGP que celle de la région de Thetford Mines. Notons, cependant, qu'à cause du mélange résultant du transport glaciaire, ces résultats ne représentent que des moyennes et qu'à l'intérieur de chacune de ces régions, il pourrait y avoir des zones où les EGP sont plus concentrés. Par exemple, Gauthier et Trottier (1987) rapportent au moins trois sites dans la région de Thetford Mines où les chromitites contiennent des concentrations nettement anomales en EGP. Cependant, leurs résultats d'analyse de chromitites d'une vingtaine de sites dans cette même région, corroborent nos données en étant en moyenne inférieurs aux limites de détection de nos analyses.

La teneur régionale en EGP des chromites pourrait tout de même être un indicateur de potentiel, mais ceci reste à démontrer. À cet égard, les pépites de platine natif et possiblement d'autres alliages d'EGP qui ont été signalées dans certains cours d'eau au nord de la rivière Chaudière (Logan et al., 1863; Mackay, 1921) pourraient être reliées à ces teneurs régionales plus élevées.

Les hautes valeurs en Pt, Pd et Ir au sud-est du mont Saint-Sébastien demeurent énigmatiques. La présence d'Ir est particulièrement curieuse car cet élément, plus que le Pd ou le Pt, tend à se concentrer dans les premières phases à crystalliser durant l'évolution des magmas basiques. Cependant, on connaît plusieurs cas d'enrichissement d'EGP dans des cuivres-porphyriques associés à certaines intrusions felsiques à caractère alcalin (Finch et al., 1983; Werle et al., 1984). De plus, les recherches des dernières années, encouragées par les forces du marché et le perfectionnement des méthodes analytiques, ont permis de constater la présence d'EGP dans des environnements géologiques inhabituels, souvent sans relations apparentes à des roches ultramafiques ou mafiques (Hulbert, 1987 et communication personnelle, 1987; Macdonald, sous presse).

La partie de la bande de Serpentine située à l'ouest de la zone d'ophiolites massive de Thetford Mines et à laquelle nous avons rattaché quelques anomalies d'EGP, pourrait effectivement être favorable à la présence de ces métaux. Oshin et Crocket (1982) signalent que certaines de ces roches (région du lac de l'Est) sont plus riches en soufre et considérablement plus riches en Pd et Pt que des roches lithologiquement semblables ailleurs. Notons que les concentrations élevées au sud-est du mont Saint-Sébastien sont situées sur le flanc sud-ouest de la traînée de chrome de Thetford Mines et qu'en remontant le trajet glaciaire à partir de ces anomalies, on aboutit dans le secteur du lac de l'Est. Il demeure néanmoins étrange qu'aucune valeur élevée n'a été trouvée entre le lac de l'Est et le mont Saint-Sébastien, ce qui nous incite à relier la traînée d'EGP à l'intrusif felsique ou à la minéralisation qui lui est associée, plutôt qu'à la bande de Serpentine.

Le dyke de Pennington devrait faire l'objet de recherches plus approfondies de même que les petites zones d'ophiolites démembrées. Laurent (1975) indique que ces petits amas sont plus altérés que les zones massives. Il est possible que cette altération ait remobilisé et possiblement concentré les EGP. Toutefois, Oshin et Crocket (1982) disent n'avoir aucune preuve que la serpentinisation des dunites à cumulat ou des harzburgites ait provoqué des changements dans les teneurs d'EGP dans la région de Thetford Mines. Cependant, Crocket et Chyi (1972) ont remarqué une plus grande dispersion du Pd dans les roches serpentinisées du massif ophiolitique du mont Albert en Gaspésie, comparé aux roches fraîches. Ils suggèrent que ce métal pourrait être mobile durant la serpentinisation.

CONCLUSION

Gauthier et Trottier (1987) ont évoqué la possibilité qu'il pourrait y avoir un certain potentiel en EGP associé aux roches ophiolitiques alpines de l'Estrie-Beauce, en particulier dans la région de Thetford Mines. Nos données suggèrent que ce potentiel, s'il s'avère réel, pourrait être encore plus élevé ailleurs, comme dans la région de Saint-Joseph où les roches ultramafiques contiennent des chromites qui renferment, en moyenne, trois fois plus de Pd que celles de Thetford Mines. D'autres part, le segment de la bande de Serpentine d'une trentaine de kilomètres de long, situé au sud-ouest de Thetford Mines, pourrait être intéressant, à en juger par la présence d'un certain nombre d'anomalies d'EGP, relativement intenses, localisées en aval glaciaire de cette zone. Ce segment renferme des masses d'ophiolites démembrées, possiblement plus altérées et plus riches en sulfures, qui pourraient contenir des concentrations non-négligeables d'EGP. Le dyke de Pennington offre également un intérêt, principalement à cause des sulfures de nickel que l'on croit lui être associés.

Quant aux anomalies localisées au sud-est du mont Saint-Sébastien, elles ont un intérêt particulier dû à leur forte intensité. Elles pourraient être reliées à un type de minéralisation d'EGP (e.g. hydrothermal) dont on ne soupçonnne pas l'existence en Estrie-Beauce.

REMERCIEMENTS

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ERRATA

Notez l'ordre des colonnes de l'appendice A

No. déch	Cr %	Pd ppb	Pt ppb	Ir ppb

APPENDICE A

(échantillons numérotés à la figure 2)

No. d'éch.Cr	Pd %	Pt ppb	Ir ppb	ppb
a) Traînée de Thetford Mines				
1	36.0	10	–	–
2	36.5	16	–	–
3	30.8	10	–	–
4	32.4	21	–	–
5	30.7	16	–	–
6	30.4	10	–	–
7	22.4	13	–	–
8	27.9	17	–	–
9	22.1	11	–	–
10	21.8	11	–	–
11	22.5	10	–	–
12	25.9	10	–	–
13	28.3	15	54	–
14	22.7	15	–	–
15	24.5	18	–	–
16	25.0	12	–	–
17	31.4	17	–	–
18	26.3	22	–	–
19	30.8	15	–	–
20	31.2	20	–	–
21	37.1	12	–	–
22	29.1	15	–	–
23	23.2	12	–	–
24	26.4	12	–	–
25	22.4	11	–	–
26	24.7	14	–	–
b) Traînée nord-est de riv. Chaudière				
27	25.3	27	–	–
28	24.6	27	–	–
29	24.5	39	–	–
30	23.8	64	–	–
31	21.2	22	–	100
32	17.1	15	–	–
33	18.7	14	–	–
34	17.2	18	–	–
35	16.6	20	–	–
36	13.7	18	–	–
37	14.6	23	–	–
38	15.6	12	–	–
39	17.1	14	–	–
40	14.9	10	–	–
41	18.8	20	–	–
42	27.4	10	–	–
43	17.6	18	–	–
44	8.6	14	–	–
45	10.3	15	–	–
46	8.8	12	–	–
47	10.4	17	–	–
c) Traînée du mont Saint-Sébastien				
48	21.8	359	156	–
49	19.8	431	438	–
50	10.9	–	54	–
51	10.6	–	–	720
52	10.0	–	–	820
53	13.7	20	–	–
54	11.4	–	–	460
55	9.6	19	–	–
56	13.8	14	–	360
57	15.0	11	–	–

No. d'éch.Cr	Pd %	Pt ppb	Ir ppb	ppb
d) Sud-ouest de riv. Chaudière et région de St. Robert				
58	0.2	–	–	170
59	0.4	–	164	–
60	5.5	27	–	–
61	16.3	10	–	–
62	0.9	11	–	–
63	3.9	12	–	–
64	2.5	15	–	–
65	1.1	26	–	–
66	1.2	25	–	–
67	1.6	12	–	–
68	2.9	10	–	–
69	10.5	12	–	–
70	5.9	30	–	–
71	9.7	11	–	–
72	14.6	24	–	–
73	9.1	12	–	–
74	7.4	10	–	–
75	6.5	21	–	–
76	8.7	11	–	–
77	6.3	12	–	–
78	6.9	–	–	110
79	17.4	12	–	–
80	21.4	11	52	–
81	23.9	57	–	–
82	15.7	24	–	–
83	19.1	35	–	–
e) Sud-ouest de Thetford Mines et Lac Aylmer				
84	0.2	20	–	–
85	0.5	15	–	–
86	4.9	–	119	–
87	2.5	11	–	–
88	2.3	19	–	–
89	1.3	12	–	–
90	2.4	11	–	–
91	27.3	15	–	–
92	11.3	–	–	120
93	13.6	12	–	–
94	9.2	10	–	–
95	22.3	14	–	–
96	13.9	11	–	130
97	18.7	15	–	–
98	11.8	12	–	–
99	17.2	17	–	–
100	10.6	20	–	–
101	14.8	10	–	–
102	22.5	13	–	–
103	14.9	–	94	–
104	21.6	10	–	–
105	16.7	10	–	–
106	21.7	–	–	140
107	11.3	15	–	–
108	11.7	14	–	–
109	13.7	–	–	120
110	1.5	–	157	–
111	15.1	–	–	560
112	17.5	11	–	–
113	18.5	14	–	–
114	21.3	10	–	–
115	16.2	15	–	–
116	14.9	10	–	–
117	19.0	11	–	–
118	19.9	10	–	–
119	8.3	–	–	160

Saint George map area: the end of the Avalon zone in southern New Brunswick

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Currie, K.L., Saint George map area: the end of the Avalon zone in southern New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 9-16, 1988.

Abstract

Late Precambrian granitoid rocks extend from the Bay of Fundy north to the Devonian Saint George batholith. These rocks are cut by the bimodal Kingston complex which marks the site of Late Precambrian ductile transcurrent motion of uncertain but large scale. Old mylonite zones bounding the Kingston complex have been used by younger normal fault complexes (Wheaton Brook, Belleisle, Kennebecasis faults) of relatively minor displacement. Silurian rocks rest unconformably on Precambrian (Avalonian) basement, and none of the faults marks a boundary between Precambrian and younger rocks, although the Wheaton Brook fault may lie close to a pronounced thinning of old crust. Mapping suggests major oblique convergent plate motions in Late Precambrian time, followed by episodic transpressional and transtensional motion through the lower Paleozoic.

Résumé

Les roches granitoïdes du Précambrien supérieur s'étendent de la baie de Fundy au nord jusqu'au batholite dévonien de Saint-George. Ces roches sont recoupées par le complexe bimodal de Kingston qui marque le site des mouvements de décrochement horizontal accompagnés de phénomènes de fluage et survenus, au Précambrien supérieur, à une échelle non déterminée mais sans doute importante. Les zones mylonitiques anciennes qui limitent le complexe de Kingston ont été empruntées par des complexes plus récents de failles normales (failles de Wheaton Brook, Belleisle, Kennebecasis), caractérisés par des déplacements relativement mineurs. Les roches siluriennes reposent en discordance sur le socle précambrien (Avalonien), et aucune des failles ne marque la limite entre les roches précambriennes et plus récentes, même si la faille de Wheaton Brook se situe probablement près d'un amincissement prononcé de la croûte ancienne. Les travaux de terrain suggèrent qu'au Précambrien supérieur, ont eu lieu des mouvements obliques de plaques convergentes, suivis de mouvements épisodiques de compression et d'extension transversales durant tout le Paléozoïque inférieur.

INTRODUCTION

Many compilers considered the Bellisle fault to be a fundamental feature of the geology of southern New Brunswick, separating Precambrian rocks of the Avalon zone from younger rocks to the north (Williams, 1979; Hatcher and Williams, 1983; Keppie, 1984). Alcock and Perry (1960) introduced the fault as follows: "The abrupt change from comparatively fresh, dark volcanic rocks of Silurian age on the northwest to more altered volcanic rocks cut by an abundance of dykes to the southeast is very striking. This contact... is believed to be a fault zone." However, Helmtaedt (1968) found a varied succession of rocks (rather than Silurian volcanics) north of the fault in the Beaver Harbour area, and concluded that movement on the fault was minor and normal. Farther northeast, Garnett (1973) also found brittle movement on the fault to be small and normal, and noted that Precambrian rocks occurred north of the fault. Rast and Currie (1976) and Rast and Dickson (1982) concluded that major motion was of Precambrian age, overprinted by Carbonifer-

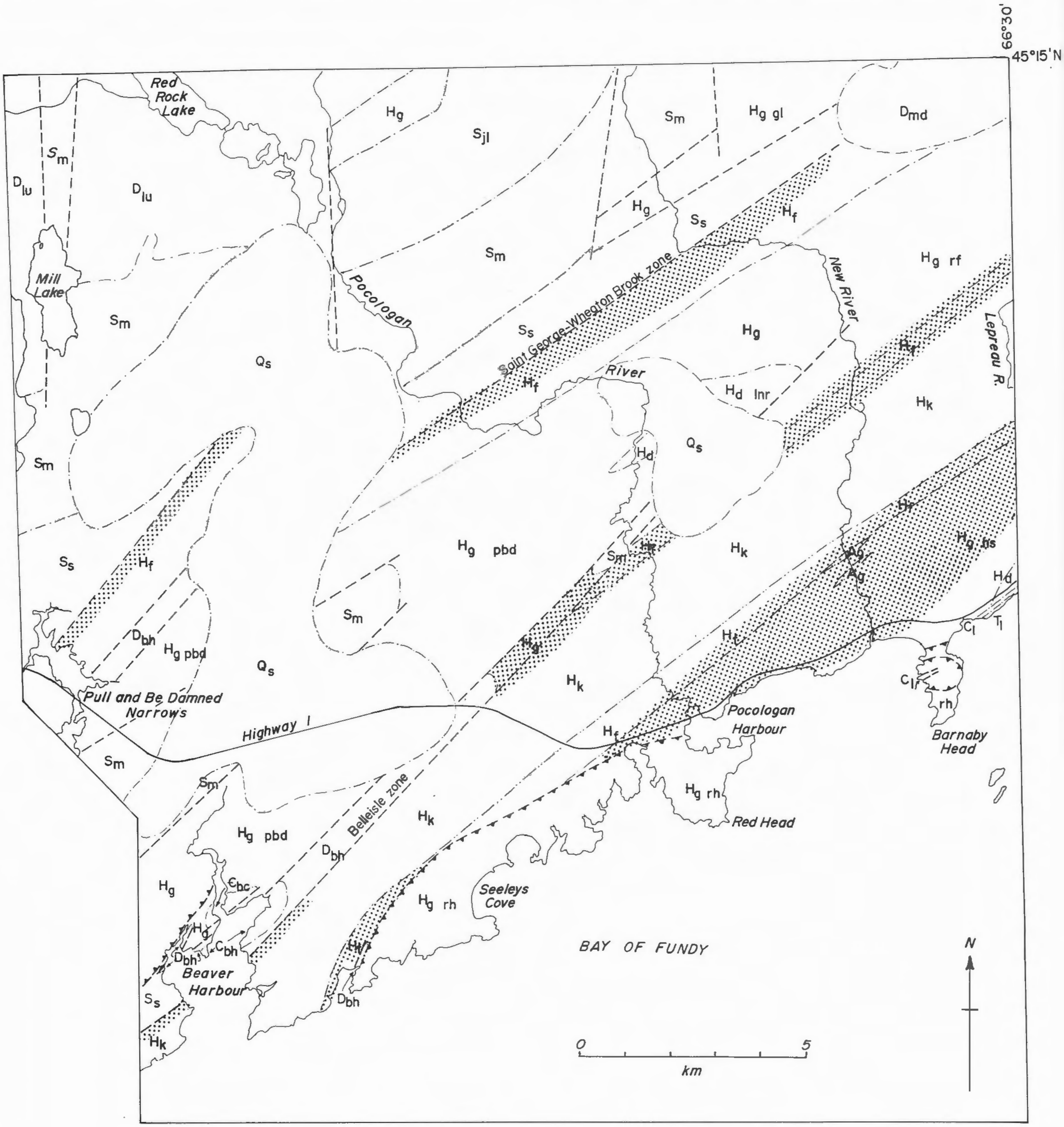

Figure 1. Geological sketch of Saint George map area (east 1/2)

LEGEND

QUATERNARY
Q_S stratified sand and gravel, moraine, till, boulder clay
-- unconformity --

TRIASSIC
T_l LEPREAU FORMATION; brown conglomerate, red sandstone, siltstone
-- unconformity --

CARBONIFEROUS
C_l LANCASTER FORMATION; grey lithic arenite, pebble conglomerate, shale
-- unconformity (on H_g) --

C_{bh} BEAVER HARBOUR FORMATION; green to tan siltstone with cobble beds
-- unconformity --

DEVONIAN
SAINT GEORGE BATHOLITH (units D_{md} and D_{lu})

D_{md} MOUNT DOUGLAS PHASE; pink biotite granite and porphyry with rapakivi feldspar; aplite dykes
-- intrusive to gradational contact --

D_{lu} LAKE UTOPIA PHASE; pink to red seriate biotite granite; tuffisite veinlets
-- intrusive contact (to S_{jl} and S_m) --

D_{bh} BLACKS HARBOUR BEDS; cleaved red siltstone and sandstone, conglomerate and fanglomerate; numerous caliche horizons
-- unconformity (on S_m and older) --

SILURIAN OR DEVONIAN
S_{jl} JAKE LEE MOUNTAIN COMPLEX; grey to purple riebeckite granite and aplite; granitic to dioritic porphyry; mafic dykes
-- intrusive contact --

SILURIAN
S_m MASCARENE GROUP; amygdaloidal basalt, mafic tuff, brown rhyolite, siltstone, chert; gabrroic dykes and sills
-- conformable contact (?) --

S_s FISH PLANT BEDS; fine banded black shale and white siltstone with massive siltstone beds
-- relations unknown --

CAMBRIAN
C_{bc} BUCKMAN CREEK BEDS; basalt flows and tuff, rhyolitic tuff, siltstone, calcareous siltstone, conglomerate
-- unconformity (on H_g) --

LATE PROTEROZOIC
H_k KINGSTON COMPLEX; sheeted felsic and mafic dykes; foliated amphibolite, felsite, minor granophyre and massive diorite
-- intrusive contact --

mylonite, blastomylonite, schist, minor massive enclaves
-- tectonic contact --

H_f massive to foliated felsite, rhyolite quartz porphyry
-- relations uncertain, probably intrusive --

GOLDEN GROVE SUITE (units H_g and H_d)
H_g hornblende granodiorite, biotite granite, megacrystic granite; rocks commonly fractured, epidotized, chloritized. Hs-Hansen Stream pluton; rf-Ragged Falls pluton; rh-Red Head pluton; gl-Goose Lake pluton; pbd-Pull and Be Damned complex
-- intrusive to gradational contact --

H_d tonalite, diorite, amphibolite; lnr-Little New River complex
-- intrusive contact --

?EARLY PROTEROZOIC
A_g mylonitized dioritic to tonalitic gneiss and migmatite

geological contact, approximate, assumed

trend of dykes

high angle fault

thrust fault

Geology by K.L. Currie 1987 with additional information from Carroll, 1984; and McLeod, 1986.

ous thrusting. In view of the importance attached to the Belleisle fault in regional interpretation, the type area for this fault (NTS 21G/2E) has been remapped (Fig. 1).

Outcrop is excellent along the coast but generally poor inland. Logging roads and all-terrian vehicle tracks provide not only access but also much of the outcrop. Regions in the centre of the area are covered by glacial outwash and boulder ridges.

DESCRIPTION OF FORMATIONS

The oldest rocks occur as enclaves of mylonitized migmatitic gneiss up to a few tens of metres in length along the lower New River and adjacent Fundy coast (unit ?A_g). The gneiss consists of chloritized biotite amphibolite and granite laminated on a millimetre to centimetre scale with many ptygmatically folded granite veins and boudined, transposed mafic dykes. The rocks resemble the middle to lower Proterozoic Brookville gneiss of the Saint John area (Currie et al., 1981).

Granitoid rocks are massive but altered and fractured (units H_d, H_g, Golden Grove suite of Hayes and Howell, 1937). Mafic rocks (unit H_d) consist of plagioclase and hornblende. The Little New River complex (lnr) is coarse grained and massive, but sufficiently metamorphosed that metamorphic amphibole visibly overgrows primary amphibole. The rocks are riddled with fine grained, foliated amphibolite dykes. On its northern side the complex forms an agmatite with slightly younger granodiorite. The western tip of the complex on the Pocologan River appears very fresh, and displays relict diabasic texture. This portion may be Silurian and related to gabbroic sills of the Mascarene Group (unit S_m). Small amounts of dioritic to tonalitic rocks of the Talbot Road pluton (tr) (Currie, 1987) occur on the eastern boundary of the map area.

In addition to projections of the Hansen Stream (hs) and Ragged Falls (rf) plutons (Currie, 1987), three major bodies of Precambrian granitoid rocks occur, namely the Red Head and Goose Lake plutons and the Pull and Be Damned complex. The Red Head pluton consists mainly of coarse grained, grey hornblende granodiorite with rounded, partially digested inclusions. This phase resembles the Hansen Stream pluton, and may be continuous with it. Red Head itself consists of a leucocratic variant of this phase containing ellipsoidal red feldspar and chloritized biotite. A more calcic phase of the pluton, essentially lacking potassium feldspar is faulted against the granodiorite in some exposures, but grades to it in others. All of the Red Head pluton exhibits a moderate to strong cataclastic overprint.

The Pull and Be Damned complex (pbd) can be examined in a large roadcut on Highway 1. The matrix consists of pink, fractured medium grained hornblende granite, variously epidotized and chloritized. The granite exhibits numerous abrupt transitions to fine grained material, either mylonitic or foliated to unfoliated felsite. The complex contains about 5-20 % of fine grained foliated amphibolitic dykes trending about 060°, thought to be correlative to the Kingston complex (unit Hk). Massive but altered north-trending gabbro dykes, petrographically identical to sills in the Mascarene

Group, form a further 15-20 % of the exposure. At its eastern end the Pull and Be Damned complex appears to pass gradationally by loss of dykes into known plutons.

Much of the Goose Lake pluton consists of coarse grained chloritized hornblende granodiorite like the Ragged Falls and Hansen Stream plutons. However the pluton displays abrupt transitions to fine grained porphyritic or felsitic phases over distances of 30 cm or less. The pluton contains numerous gabbroic intercalations in its western part, thought to be correlative to petrographically identical sills in the nearby Mascarene Group.

In the Saint George area the granitic rocks of unit H_g exhibit a degree of deformation and metamorphism not seen in nearby Paleozoic rocks, and are cut by foliated amphibolite dykes whose age has very generally been taken to be Precambrian (Helmstaedt, 1968; Garnett, 1973; Rast, 1979; Rast and Dickson, 1982; Currie, 1984, 1987). Where (tectonic) enclaves of Paleozoic rocks are found in the granite the contrast between the unmetamorphosed, fossiliferous sedimentary rocks and the brecciated, greenschist grade granite is particularly striking. A good Pb-U zircon age of 555 Ma was obtained from the Ragged Falls pluton on the west boundary of the map area, identical to the age of the younger granitoid rocks south of the Belleisle fault reported last year (Currie, 1987).

High level felsic rocks ranging from (predominant) pink felsite through quartz-feldspar porphyry and aplite to rhyolite (unit H_f) fringe the Kingston complex, occur as a separate belt farther north, and form patches within the Pull and Be Damned complex and Goose Lake pluton. Considerable parts of these belts are mylonitized, but original igneous texture is locally preserved. Textural changes from felsite to porphyry to aplite to true granitic textures are abundant and abrupt but transitional over a few centimetres. No unequivocally volcanic textures were seen by me although Rast and Dixon (1982) described such textures.

A distinctive features of the Saint George area is the abundant development of mylonite (indicated on the map by a stippled pattern). The Pocologan zone has been described by Rast and Dickson (1982) and their comments apply to the zones farther north. Mylonite zones fringe both sides of the Kingston complex, but affect the complex only slightly and marginally, suggesting that mylonitization is older than, or contemporaneous with, dyke emplacement (Rast and Dickson, 1982; Currie, 1987). Mylonites exhibit greenschist grade metamorphism like the Kingston complex, with grade decreasing slightly from the core of the Kingston complex to the edge of the mylonites. Mylonites affect a variety of Precambrian rocks (mainly units H_g and H_f), but no Paleozoic rocks, although Paleozoic rocks overlie a mylonite zone east of Beaver Harbour. All mylonite zones are associated with younger brittle deformation, as shown by fault slices of Paleozoic rocks in mylonite zones.

The Kingston dyke complex (unit H_k) extends more than 100 km from the edge of the Carboniferous cover near Hampton to Campobello Island. Throughout its length the complex, which varies in width from 2 to 6 km, consists almost entirely of mafic and felsic dykes, now at greenschist grade of metamorphism but so little deformed that igneous layering and chilled margins can commonly be recognized. Dyke widths vary from 15 cm to 200 m but commonly fall in the 1-10 m range. Dyke contacts commonly are mutually chilled, but crosscutting by both acid and mafic phases occurs, and some contacts are sheared. East of Beaver Harbour a few of the felsic dykes are mylonitic. Mafic dykes range from fine grained homogeneous through feldspar-phyric to appinitic while felsic dykes are commonly felsitic, but locally spectacularly glomeroporphyritic, or flow banded. Small elongate plutons of gabbro (on Lepreau River) and granophyre (in New River) also occur. Where not obscured by cataclasis, the margins of the complex are gradational with progressively fewer dykes intruding a felsic host. The age of the Kingston complex is fixed by stratigraphic evidence as late Precambrian (Currie, 1984, 1987).

Helmstaedt (1968) discovered a fossiliferous Cambrian sequence in Buckman Creek at the head of Beaver Harbour. As the stratigraphy of this sequence has not been formally described, I use the name Buckman Creek beds (unit C_{bc}). The sequence can be examined on both sides of a syncline. To the south, a variably welded quartz-feldspar tuff lies on older granite, and is overlain by basaltic flows and tuff. To the north vesicular basalt lies directly on granite which is cut by similar basalt dykes. The middle of the sequence consits of 20 to 50 m of greygreen mafic lapilli tuff, tuffaceous siltstone and calcareous siltstone which yielded a trilobite pygidium typical of the base of the Middle Cambrian (P. benneti zone, W.H.Fritz, pers. comm., 1987). All of this zone has been affected by carbonate alteration. The top of the sequence comprises red siltstone with volcanic cobbles. The lack of metamorphism and the relatively weak deformation compared to the nearby Kingston complex and fringing mylonite zones is striking. The Buckman Creek beds occur within a complex faulted block, and similar lithologies have not been recognized elsewhere in the Saint George area. A similar sequence, also in a complex fault block, near Westfield on the Saint John River, was mapped as Eocambrian (Currie, 1987).

The oldest Silurian rocks appear to be rhythmically banded black shale and white siltstone alternating on a scale of a few millimetres, here termed the Fish Plant beds (unit S_s). South of the Saint George batholith, these beds clearly underlie basaltic rocks of the Mascarene Group. On the west shore of Beaver Harbour this sequence exhibits massive siltstone beds up to 50 cm thick at the top of the sequence, and intricate slump folds and load casts. At the north end of the harbour this unit had disintegrated to form the matrix of a tectonic mélange. Primitive fossil plants from this unit were originally thought to be Devonian (Cumming, 1967) but reconsideration suggests a nonspecific Silurian age (D.C.McGregor, pers. comm., 1987).

The Mascarene Group (unit S_m) can be divided into a lower basaltic and upper rhyolitic and sedimentary sequence (Ruitenberg and McCutcheon, 1978). The lower sequence consists of minor vesicular, xenolith-rich basalt flows and abundant tuff, tuffaceous siltstone and related rocks with numerous gabbroic sills up to 50 m thick, which also occur sparsely in unit S_s. The large number of granite xenoliths leaves little doubt the volcanic rocks passed through granitic basement. The rocks are altered, but not significantly metamorphosed or recrystallized, and face monoclinally north.

The upper part of the Mascarene Group is poorly exposed excepts in local knobs of hornfels. The rocks appear to consist mainly of greenish siltstone and cherty siltstone with one or more intercalations of brownish, flow-banded rhyolite. Just west of the mapped area, correlative rocks yielded a Llandovery to Wendlock fauna (Boucot et al., 1966). Rhyolitic rocks possibly correlative to these strata (Letang rhyolite of Donohoe (1978)) form a rib of partially intrusive, strongly cleaved rhyolite, porphyry and minor siltstone stretching east from Pull and Be Damned Narrows some 10 km.

The Jake Lee Mountain complex (unit S_{jl}) forms an elliptical northeast-trending mass cored by grey, coarse riebeckite granite granding to a carapace of purplish aplitic amphibole granite which fines toward the contact. On the northwest side, a sequence of prophyries separating the complex from the Saint George batholith ranges from biotite diorite or gabbro to alaskite, and displays mixing textures ranging from megascopic to armoured xenoxrysts, particularly amphibole-cored plagioclase. Very fresh, fine grained basaltic dykes cut the porphyry but not the Saint George batholith. The Jake Lee Mountain complex hornfelsed the Mascarene Group, but is fractured, discoloured and veined by quartz, against a finer grained marginal phase of the Lake Utopia pluton (unit D_{lu}). The age of the complex is therefore confined between Wenlock and the Middle Devonian age of the Lake Utopia.

A distinctive sequence of red siltstone and sandstone with abundant caliche horizons and numerous intervals of coarse conglomerate (unit D_{bh}) occurs in a number of fault slices. The cobbles include red shale rip-ups and Precambrian granite. These beds are commonly strongly folded and cleaved, and contain no debris from the Devonian granites. Schluger (1973) included them in the Upper Devonian Perry Formation, but Alcock and Perry (1960) described the Perry Formation as slightly deformed and containing debris of the Saint George batholith. I have therefore adopted a suggestion of N. Rast, and term this sequence the Blacks Harbour beds of presumed lower to middle Devonian age. At the northern end of Beaver Harbour, slightly deformed sandstone and siltstone of the Carboniferous Beaver Harbour Formation is thrust over strongly cleaved and folded Blacks Harbour beds.

The Mount Douglas phase of the Saint George batholith (unit D^{md}) underlies a small area in the northeast corner of the map with pink biotite granite and porphyry containing rapakivi feldspars up to 5 cm accross. This boss causes a hornfels aureole in the Fish Plant beds, but merely fines in grain toward chloritized and fractured Goose Lake pluton. The Lake Utopia phase (unit D_{lu}) consists of pink seriate biotite granite with very coarse porphyritic phases containing K-feldspar crystals up to 10 cm long. Biotite is always finer in grain size than the other phases. The Lake Utopia phase is fringed by an extensive hornfels aureole in the Mascarene Group. According to M.J. McLeod (pers. comm., 1987), all granitic phases of the Saint George batholith appear to have been emplaced about 370 Ma.

The Beaver Harbour Formation (Helmsteadt, 1968) consists of less than 100 m of grey-green to tan siltstone and sandstone with nebulous conglomerate lenticles (unit C_bh). Poorly preserved Visean to Namurian plant remains abound on fine grained green siltstone bedding surfaces of this fluvial sequence. Cobbles were derived mainly from the Kingston complex and older granites, but red sandstone fragments may stem from the Blacks Harbour beds. The Beaver Harbour Formation is little deformed. Dips exceed 25° only adjacent to faults, and cleavage is rare.

The Westphalian Lancaster Formation (unit C_l) outcrops on the eastern margin of the area. Half 0.5 m of red calcareous shale rests on Precambrian granite and passes up into 10 m of typical cleaved grey lithic arenite and pebble conglomerate. At the top of the sequence brown, massive conglomerate probably belongs to the Carboniferous Balls Lake Formation (Currie and Nance, 1983). A small fault sliver of Lancaster Formation occurs on the west side of Barnaby Head about on strike with the other occurrence. Brown, gently dipping conglomerate of the Triassic Lepreau Formation occurs on two offshore islands. Helmsteadt (1968) and Garnett (1973) assigned some basalt dykes a Triassic age, but no dykes have been found to cut undoubtedly Carboniferous or younger rocks, and it seems more probable that these dykes are Devonian.

STRUCTURAL GEOLOGY

The region south of the Kingston complex, essentially the Red Head pluton, was examined by Rast and co-workers (Rast et al., 1978; Rast and Dixon, 1982). The pluton is locally affected by intense cataclasis and mylonitization associated with the Pocologan zone, but Carboniferous deformation is more pervasive. Well developed cleavage(s) dips south at 30 to 50°and may exhibit thin mylonitic bands. Strong Carboniferous deformation stops on a northwest-directed thrust exposed at Barnaby Head and on a road west of Red Head where the fault plane contains a metre of foliated carbonate. Weak to moderate, gently south-dipping cleavage occurs north of this fault in a narrow belt.

The Kingston complex is bounded on both sides by mylonite zones. The southern (Pocologan) zone described by Rast and Dixon (1982), and the northern (Seven Mile Lake) zone described by Garnett (1973) and Leger and Williams (1986) are separated by several kilometres of unmylonitized dykes. The association of dykes and mylonite over tens of kilometres strongly suggests related origins. A model of left-handed transtensional faulting and right hand transpressional closing (Fig. 2) can explain the observed attitudes of the dykes and provides an explanation of almost horizontal stretching lineations in the mylonite zone passing gradationally to vertical in the dyke complex. The model assumes the horizontal lineations are due to late transcurrent motion, but the vertical ones due to earlier upwelling of magma. Young steeply dipping brittle deformation occurs on both sides of the Kingston complex. The brittle zone to the north is termed the Belleisle fault zone, while that to the south is the Kennebecasis fault zone. Fault-bounded slivers of Paleozoic sedimentary rocks suggests graben-like features on both faults, with sedimentary rocks down-dropped into, or possibly accumulated in, fault-bounded troughs, and subsequently compressed as evidenced by exposed thrust faults within several sequences. Helmstaedt (1968) and Garnett (1973) found negligible transcurrent brittle motions. The Blacks Har-

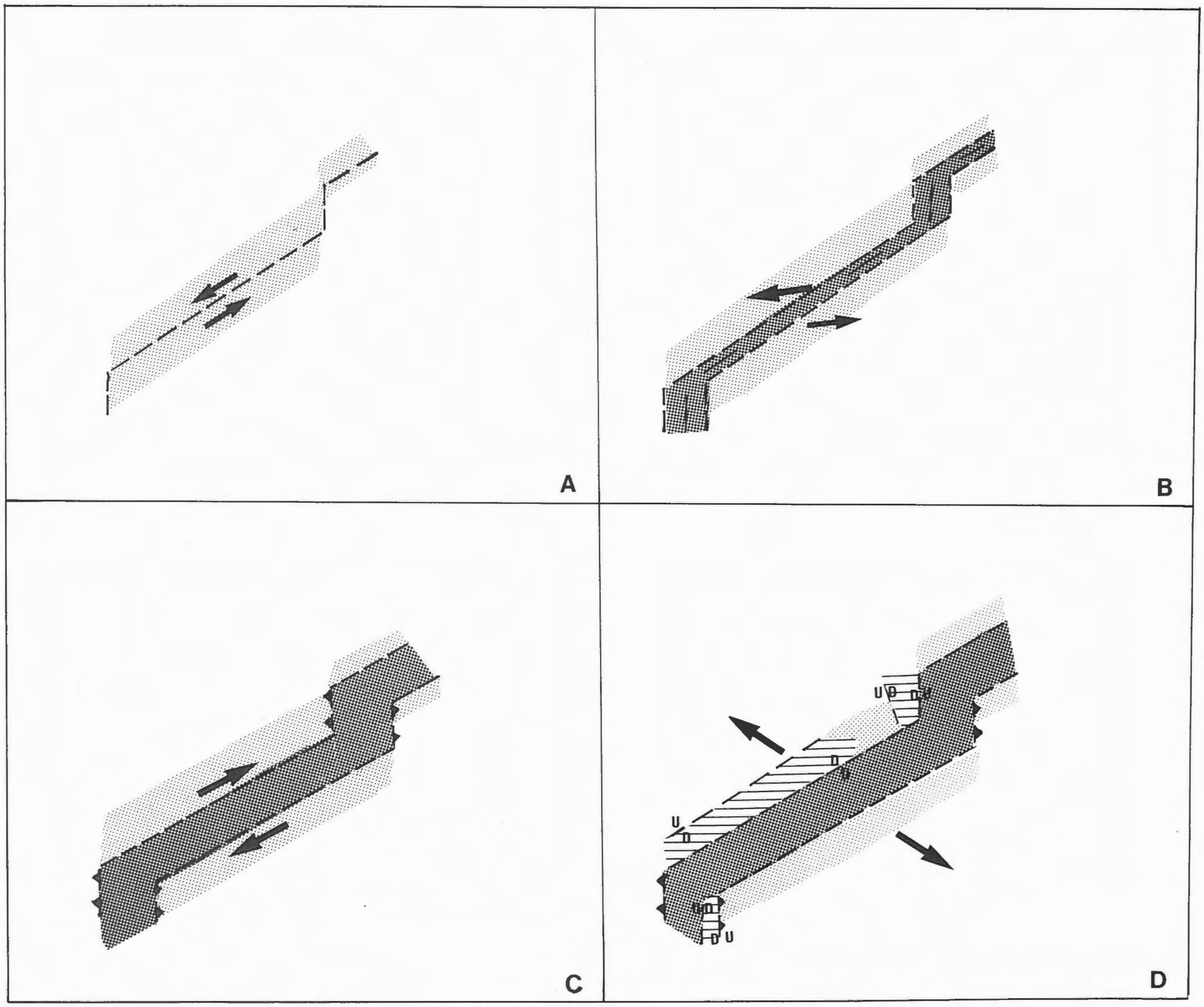

Figure 2. Diagram of development of the Kingston complex

(A) Assumed initial configuration. Fault with offsets or irregularities undergoes sinistral ductile shearing producing a mylonite zone (light stipple). For simplicity a fault plane is shown, but the scale of ductile deformation implies a rather diffuse zone.

(B) The sinistral motion develops a transtensional component, leading to emplacement of dykes (dark stipple). Trend of dykes is shown by lines. Note that the swarm is wider where it trends north, and that the dykes intrude and split the mylonite zone.

(C) Emplacement of dykes is terminated by development of transpressional motion. Dextral structures developed during this phase overprint the dykes. Note that at the changes in dyke strike, thrusting must develop.

(D) Late high-level brittle faulting. Tension as shown by the arrows produces normal faults with movements as shown. These fault troughs trap sedimentary outliers (hachured). An even later phase (not shown) is marked by compression, deforming these sedimentary outliers.

bour beds can be readily correlated across both faults, demonstrating that post-lower (?) Devonian movements must be relatively small.

The vast number of presumably Silurian dykes and sills suggest that the Pull and Be Damned complex was under tension during early to middle Silurian time. However the complex is thrust over Silurian and younger rocks at the Head of Beaver Harbour, and possibly on Highway 1. The rapid alternation of tension and compression resembles that in the sedimentary slivers described above. The northern boundary between Silurian rocks and the Pull and Be Damned complex is a major fracture zone (Saint Georges fault of Donohoe, 1978), in which the granites to the south contian tectonic slivers of Silurian sedimentary rocks. Like the Belleisle and Kennebecasis faults, this brittle deformation is superimposed on older mylonite well exposed in New River. This major fault is truncated by the Saint George batholith to the east, but appears to be the continuation of the Wheaton Brook fault which also separates Silurian and Precambrian rocks (Currie, 1984, 1987).

The Silurian, and possibly younger rocks, are cut by major north-trending faults. A 2 km sinistral offset in the boundary of the Saint George batholith north of Mill Lake is accompagnied by brecciation and alteration of the granite and a strong map linear. The magnetic map and lithological patterns suggest an offset of several kilometres along the line of the upper Pocologan River. The Jake Lee Mountain complex is truncated on this linear, although a continuous marginal fine grained zone suggests they fault controlled, rather than off-set the pluton. The other end of the pluton lies against a major zone of tectonic schist just west of upper New River which offsets the Silurian and appears to truncate the granite rib running southwest from Goose Lake. This rib is likewise cut by various north-trending and northeast-trending faults exposed in brooks southwest of Goose Lake.

DISCUSSION

The Belleisle fault zone is not a significant lithological separator of Phanerozoic rocks. Major ductile motion took place in late Precambrian time following emplacement of a major batholithic belt, probably as a late result of oblique convergence. The scale of motion remains uncertain because reliable marker horizons are lacking, but the general coherence of the geology across the Kington complex suggests that motions did not excede some tens of kilometres. The geometry of the movement zone and preserved small structure (Rast and Dickson, 1982) suggest that the initial motion was sinistral and transtensional. Large scale emplacement of mafic dykes split a major mylonite zone into two parts, and eventually led to crustal anatexis producing the Kingston complex. Dyke emplacement was terminated by destral transpressional movements producing the presently observed geometry of the Kingston complex.

At the end of igneous activity the surface appears to have been close to sea level, as shown by the nature of the Cambrian sections. The total absence of late Arenig to lower Llandovery beds from the Avalon zone of New Brunswick suggests non-deposition, or emergence during this time. In Llandovery time the old granitic crust was being stretched and intruded by dyke (Pull and Be Damned complex), with fluvial sequences collecting in possibly fault bounded troughs (Fish Plant beds), and shallow marine strata just to the west (Back Bay area, Boucot et al., 1966). The record strongly suggests a tensional breakup of the Avalon terrane with initial basaltic volcanism gradually becoming bimodal, and persisting into the Devonian. The lack of major folds in the Silurian section is particularly striking. Abundant minor folds are found only adjacent to faults (compare Donohoe, 1978). Major transcurrent faulting may also be suggested by the peralkaline Jake Lee Mountain complex (Whalen et al., 1987).

The Blacks Harbour beds mark commencement of a regime in which a continental block was repeatedly rifted and compressed along the line of old weaknesses (Wheaton Brook, Belleisle and Kennebecasis faults). The rift-fill of alluvial sediments exhibits spectacular and easily recognized evidence of compression and tension, but the motions involved appear to be minor and superficial.

ACKNOWLEDGMENTS

The generous hospitality of Nick Rast, University of Kentucky and his inimitable panache in describing and demonstrating Fundy geology have not produced complete agreement, but our views on the Saint George area are in reasonable accord. Terminology for the Siluro-Devonian granitoid rocks was adopted after discussions and field trips with Malcolm McLeod, NBDNR. I have extensively used an outcrop compilation (Carroll, 1984) representing fieldwork by several NBDNR geologists. The immense labour of finding and locating outcrop benefits all future workers, whatever the merit of the lithological identifications and interpretations.

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A preliminary report on the sedimentology, tectonic control and resource potential of the Upper Devonian — Lower Carboniferous Horton Group, Cape Breton Island[1]

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Hamblin, A.P., A preliminary report on the sedimentology, tectonic control and resource potential of the Upper Devonian-Lower Carboniferous Horton Group, Cape Breton Island; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 17-22, 1988.

Abstract

Detailed sedimentological study of the Horton Group may allow interpretation of facies present, paleogeography, contemporaneous structural style and resource potential. The lithologically-defined stratigraphic units of previous workers are probably diachronous megafacies representing tectonically-controlled depositional settings arranged laterally in the basin. These megafacies include: a) Fisset Brook continental tholeiitic basalts extruded near the margins of extensional basins, b) Craignish grey to red conglomerate, sandstone and siltstone deposited in basin margin alluvial fans and braidplains, c) Strathlorne grey siltstone, limestone and sandstone in coarsening-upward sequences deposited in basin centre lakes, d) Ainslie red conglomerate or red to grey interbedded sandstone, siltstone and limestone deposited in basin margin alluvial fans and basin centre fluvial systems. Paleocurrent data indicate that Mabou Highlands and parts of Cape Breton Highlands were sediment sources throughout Horton deposition but Creignish Hills were not. The eastern margin of the basin probably lay in the Bras d'Or Lake area. Potential exists for accumulations of petroleum and metallic minerals, with distributions partly controlled by the sedimentary characteristics of the Horton Group.

Résumé

Grâce à une étude sédimentologique détaillée du groupe de Horton, il est possible que l'on puisse y interpréter les faciès présents, la paléogéographie, le style structural contemporain ainsi que le potentiel du point de vue des ressources. Les unités stratigraphiques définies lithologiquement par les chercheurs précédents sont sans doute des mégafaciès diachrones qui représentent des cadres sédimentaires à contrôle tectonique, disposés latéralement dans le bassin. Ces mégafaciès comprennent: a) les basaltes tholéiitique continentaux de Fisset Brook, qui se sont épanchés près de la marge des bassins en extension, b) les conglomérats, grès et sandstone gris à rouges de Craignish, qui se sont déposés en marge du bassin, dans des cônes alluviaux et plaines à réseau hydrographique anastomosé, c) les sandstone, calcaires et grès gris de Strathlorne qui forment des séquences positivement granoclassées déposées dans les lacs du centre du bassin, d) les conglomérats rouges ou les grès, sandstone et calcaires rouges à gris, interstratifiés de la formation d'Ainslie, déposés en marge du bassin dans des cônes alluviaux et des réseaux fluviatiles traversant le centre du bassin. Les données sur les paléocourants indiquent que les hautes-terres de Mabou et des portions des hautes-terres du Cap Breton ont été des sources sédimentaires pendant toute la durée du dépôt de la formation de Horton, contrairement aux collines de Craignish. La marge est du bassin se trouvait probablement dans la région des lacs Bras d'Or. Il est possible que cette région contienne des accumulations de pétrole et des gîtes minéraux métallifères, dont la distribution est partiellement déterminée par les caractéristiques sédimentaires du groupe de Horton.

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INTRODUCTION AND OBJECTIVES

This report summarizes the results of the first field season investigating the sedimentology of the Horton Group on Cape Breton Island, which comprises post-Acadian basal nonmarine clastics beneath the Windsor Group. Horton sediments were deposited in fault-bounded extensional basins in central and western Cape Breton Island. They have significant mineral and petroleum potential both onshore and offshore, but have not been investigated in detail using sedimentological data as a predictive tool.

The initial focus in on the Mabou/Lake Ainslie/Baddeck area of western Cape Breton Island (lat. 45°45′ to 46°30′, long. 60°30′ to 61°30′) (Fig. 1). Stream and coastal outcrops provide most surface data. Outcrop quality is highly variable, and nowhere is the entire Horton Group well exposed. Nevertheless, the facies present can be characterized in detail and vertical/lateral facies distribution can be recognized.

Objectives of the study fall into three categories, all closely interrelated.

a) Structural/stratigraphic — to delineate the geometry and style of the basin(s) and the organization of the enclosed sediment, then decipher the geological history of Horton deposition and its relation to the regional tectonic setting.

b) Sedimentological — to describe and interpret the facies present, the predictability of their distribution and geometry, and the effects of syndepositional tectonic movements.

c) Resource potential — to investigate the ways in which the stratigraphic and sedimentological characteristics of the Horton can be used as predictive tools in exploration, and as a model for other comparable units.

METHOD OF STUDY

An analysis of the sedimentary fill of the basin may help establish predictable relationships among a) tectonic setting, b) syndepositional structural movements, c) facies deposited, d) facies geometries, and e) resource potential. Detailed sedimentological study at each major outcrop may allow interpretation of the facies and paleogeography, as well as identification of exploration targets. A limited amount of subsurface data is available and can be integrated into the surface data to enhance the three-dimensional view of the complex architecture of these depositional basins. Some palynological samples are being analyzed and some paleomagnetic data are available to aid stratigraphic correlation in these poorly fossiliferous sediments.

The study began with one month of literature review and compilation, followed by one month of reconnaissance fieldwork. This was followed by three months of basic fieldwork including detailed measurement of selected sections, sedimentological descriptions, acquisition of paleocurrent data, lithological and palynological sampling.

PREVIOUS WORK

The Early Carboniferous of Cape Breton Island has been studied intermittently since the work of Dawson (1858), with emphasis on stratigraphy before 1968 and on basin structure since then. Much of this work was spurred by the presence of oil seeps and several phases of petroleum exploration drilling from 1869-1983.

1. Tectonic style. Bell (1958) and Kelley (1967) visualized deposition in a broad subsiding downwarp (Fundy Basin) in which semi-connected subbasins were separated by active fault-bounded internal uplands. Intermittent tectonic rejuvenation continued from Late Devonian through Early Carboniferous times. Regional subsidence and low topographic margins produced similar overall stratigraphy throughout Atlantic Canada.

Belt (1968) visualized an active, fault-bounded graben with several major subbasins separated by internal basement horsts. Periods, magnitudes, and sense of dip slip motions would vary throughout the "rift" producing a different local stratigraphy in each subbasin. However the basal volcanics and trend of finer grain size toward basin centre would be typical of all parts.

Bradley (1982) suggested that all Carboniferous basins in Atlantic Canada are small individual pull-apart basins developed within a major post-Acadian dextral oblique-slip fault zone. The resulting basins should each have their own complex local stratigraphy, geometry and structural history.

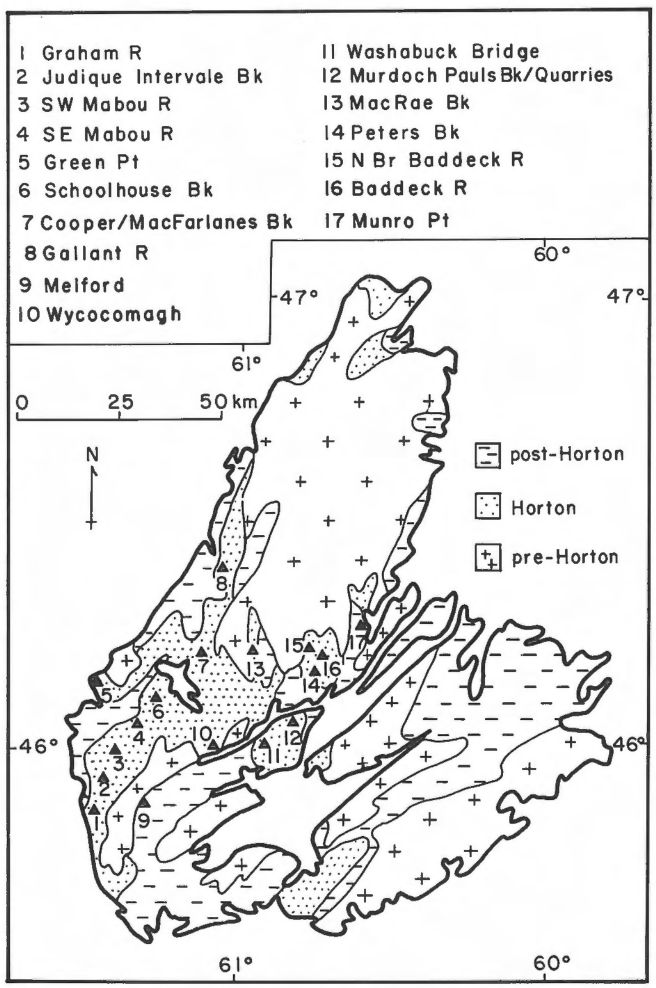

Figure 1. Generalized geological map and measured sections.

2. General stratigraphy. The Horton Group comprises up to 3000 m (but generally 1000 m) of nonmarine clastics, ranging in age from Frasnian? to Tournaisian, deposited in an array of fault-bounded basins between upthrown basement blocks. These strata represent the initial post-Acadian deposits in the complex Maritimes "Basin". The type section for the Horton Group is at Horton Bluff in Minas Basin area (*see* Bell, 1960), but on Cape Breton Island Murray (1960) suggested Southwest Mabou River as a suitable reference section.

Norman (1935), Bell (1958), Murray (1960) and Kelley (1967) identified a general stratigraphic sequence of four units on Cape Breton Island.

a) Fisset Brook Formation ("pre-Horton"): basic to felsic volcanics with interbedded red sediments, up to 500 m thick, unconformably overlie Acadian basement and have localized distribution near basement blocks. This unit grades upward into the Craignish Formation.

b) Craignish Formation ("lower" Horton): red and grey coarse clastics up to 1500 m thick, overlie the Fisset Brook, or more commonly, unconformably overlie Acadian basement. The Craignish is thought to be thickest near basin centre (west) and thinnest near basin margin (east). There is an upward-fining sequence of members: lower Graham River conglomerate and sandstone, middle Skye River sandstone and siltstone, upper Macleod siltstone and sandstone.

c) Strathlorne Formation ("middle" Horton): grey, thinly laminated siltstone or mudstone with some thin sandstone and limestone beds, up to 300 m thick, sharply but conformably overlie the Craignish Formation. This unit thins from basin centre to basin margin and has a gradational top, interfingering with the Ainslie Formation.

d) Ainslie Formation ("upper" Horton): grey and red sandstone, siltstone and conglomerate up to 700 m thick, gradationally overlie the Strathlorne Formation but have a sharp (disconformable) upper contact with the basal Macumber Formation of the Windsor Group. There is an upward-fining sequence of members: lower, McIsaac Point conglomerate, sandstone and siltstone, and upper, Glencoe siltstone and sandstone.

SUMMARY OF OBSERVATIONS

Seventeen sections comprising 11 500 m of Horton strata were measured, evenly distributed throughout the study area. Many other sections, less well exposed or less accessible were checked. A total of 96 samples were collected, of which 30 are for palynological analysis to aid in correlation. A total of 544 paleocurrent measurements were made.

On a gross scale, the tripartite stratigraphic scheme of the literature can be recognized in most sections. These lithologically-defined units are probably diachronous megafacies which represent tectonically-(or perhaps climatically-) controlled depositional settings arranged laterally in fault-bounded basins. Therefore equivalent lithofacies may not correlate precisely in time between sections or all be well developed at each section, depending on the position within the basin. All sections measured display great variability in lithology, grain size, paleocurrents and colour. This appears to be a hallmark of Horton deposition, but renders correlation very tenuous. In addition, the obvious presence of post-depositional faulting and folding complicates correlation, especially in poorly exposed sections where the measured and estimated thicknesses of Horton strata may be greater than the true thicknesses.

HORTON MEGAFACIES

The Craignish megafacies is well developed in the Graham River, McFarlanes Brook and Baddeck River areas. The Strathlorne and Ainslie megafacies are well developed on the various branches of the Mabou and Baddeck rivers. Toward the basin centre (Mabou area) the Horton has a composite measured thickness of about 2700 m, whereas nearer the basin margin (Baddeck area) the composite measured thickness is about 1000 m. Nevertheless, the characteristics of the three megafacies in these two areas can be compared (Fig. 2).

Fisset Brook volcanics were observed only on Cooper Brook where 300 m were measured. The lower part of the section comprises pink fine crystalline rhyolite, generally crudely bedded but with some finely laminated units. This is succeeded by dark green thick-bedded, fine crystalline basalt with a few 1-2 m thick flow breccias. The breccias

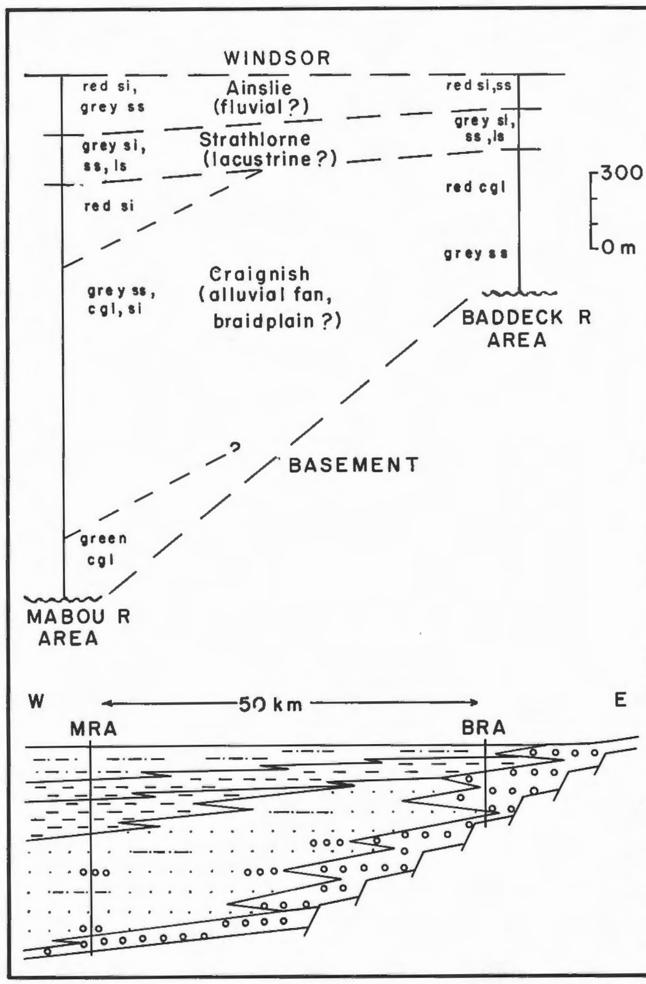

Figure 2. Horton depositional setting and megafacies.

are clast-supported with basalt cobbles set in a reddish glassy matrix. The upper part of the section comprises reddish-brown fine crystalline, highly altered, vesicular basalt. Blanchard et al. (1984) dated these rocks at 370 ± 20 Ma (whole-rock Rb-Sr method) and interpreted them as continental tholeiitic basalts extruded near the margins of extensional basins.

The Craignish megafacies, exposed in a more marginal position in Baddeck River area, is up to 700 m thick. It consists of red polymictic matrix-supported conglomerate and pebbly sandstone in thin fining-upward beds with trough crossbedding. Also present are thick red siltstone units with calcareous nodules and limestone beds which separate coarser units. The red colour gives way upward to grey. These strata can be interpreted as the deposits of alluvial fans near basin margin faults. In the Graham River/Mabou River area, in a more basinal position, the Craignish consists of: a) lower green and grey quartzitic conglomerate with some fine grained sandstone and siltstone, and includes many dioritic sills (Graham River Member), b) middle thin bedded, well-sorted, grey fine- to coarse-grained pebbly sandstone with minor conglomerate and siltstone (Skye River Member), c) upper thin-bedded, red, calcareous siltstone, commonly massive and with desiccation features, interbedded with red or grey, fine- to coarse-grained sandstone in fining-upward units (Macleod Member). The 1700 m Craignish here displays an overall fining-upward trend and is interpreted as the deposits of alluvial fans and braidplains in an overall waning tectonic phase.

The Strathlorne megafacies, well exposed in Baddeck and Washabuck areas, consists of 150-200 m of dark grey siltstone to claystone, commonly organic-rich and burrowed with small horizontal feeding and resting traces. Thin interbeds of sharp based, rippled, very fine grained calcareous sandstone and limestone are abundant. The beds are commonly arranged into 2-20 m thick coarsening- (and shallowing-) upward sequences, and there is a thin but distinct coarsening-upward shoreline sequence (including hummocky cross stratification) between the Craignish and Strathlorne megafacies. This suite of sediments is interpreted to represent shallow but quiet lacustrine deposition, perhaps during a period of relative tectonic quiescence. In the Mabou River area the Strathlorne consists of 200-400 m of dark grey, thinly interbedded siltstone and calcareous, rippled very fine grained sandstone, with thick intervals of green or grey calcareous very fine- to coarse-grained sandstone in thick beds with sharp scoured bases, trough crossbedding and ripples. Here the Strathlorne may represent an area of greater energy level, sediment input and fluvial/shoreline influence within the lacustrine basin. In all areas Strathlorne units may interfinger with uper Craignish and lower Ainslie megafacies in transitional zones.

The Ainslie megafacies also occurs in two distinct forms depending on basin position, but both probably represent tectonic rejuvination and basin infilling by continental sedimentation. Near basement blocks where the Horton is thinner (Whycocomagh, Green Point, Washabuck), the Ainslie commonly consists of 100-500 m of thick coarsening-upward sequences of stratified pebbly coarse sandstone to matrix-supported conglomerate in sharp based fining-upward beds. These beds commonly display trough crossbedding, pebble imbrication, scour pockets, and discontinuous lenses of red medium grained sandstone. These strata can be interpreted as prograding alluvial fan sediments associated with fault margins. Away from active fault margins (Mabou R., Baddeck R., Gallant R.) the Ainslie consists of 100 to 400 m of micaceous sediments in an overall fining-upward sequence. This includes: a) lower interbedded red to grey siltstone with green limestone beds and desiccation features, and red, green or grey calcareous fine to coarse sandstone in sharp based fining-upward units with abundant crossbedding (McIsaac Point Member), and b) upper poorly exposed calcareous red siltstone and fine sandstone with thin limestone beds and concretions (Glencoe Member). The Ainslie in these areas can be interpreted as subaerially exposed fluvial/floodplain deposits in a basin with low topography. In all areas the uppermost Horton is overlain sharply, but apparently conformably, by distinctive dark grey laminated marine limestone of the Macumber Formation (basal Windsor Group), which probably represents a very extensive and nearly synchronous marine transgression (*see* Kirkham, 1978 for discussion).

PALEOCURRENTS AND PALEOGEOGRAPHY

Due partly to variable exposure quality, paleocurrent data are sparse and in some sections inconclusive. The predominant paleocurrent flow directions depicted on Figure 3 represent a preliminary visual estimate, based on data from all types of sedimentary structures lumped together. Further refinement of this information will lead to more sophisticated analysis.

Ninety-seven measurements were made in 7 widely separated exposures of Craignish alluvial fan/braidplain sediments, generally on medium scale trough crossbedding in conglomerate and sandstone. At most sections little relation can be discerned between dominant flow direction and nearby basement blocks. Near the Creignish Hills block most measurements indicate northeast or southwest flow which may suggest that this block was not a major sediment source in early Horton time (an idea first mentioned by Murray, 1960). In contrast, in the Baddeck area the few measurements indicate flow away from the nearby Cape Breton Highlands basement blocks.

The Strathlorne lacustrine megafacies has yielded 124 measurements from 9 sections, mostly as ripple crest trends, ripple cross-lamination and tool marks in sandstone beds within the finer sediments. At any given section these tend to fall into a dominant direction. In the Mabou area many measurements indicating flow directions toward the east, south and west, suggest that the Creignish Hills were not a significant sediment source in middle Horton time. However, in Baddeck area the few measurements indicate flow away from nearby basement blocks.

The Ainslie alluvial fan/fluvial/floodplain megafacies is the best and most extensively exposed, and yielded 323 paleocurrent measurements in 16 sections. Most of these data represent trough crossbedding and ripple cross-lamination in sandstone, or trough crossbedding and pebble imbrication

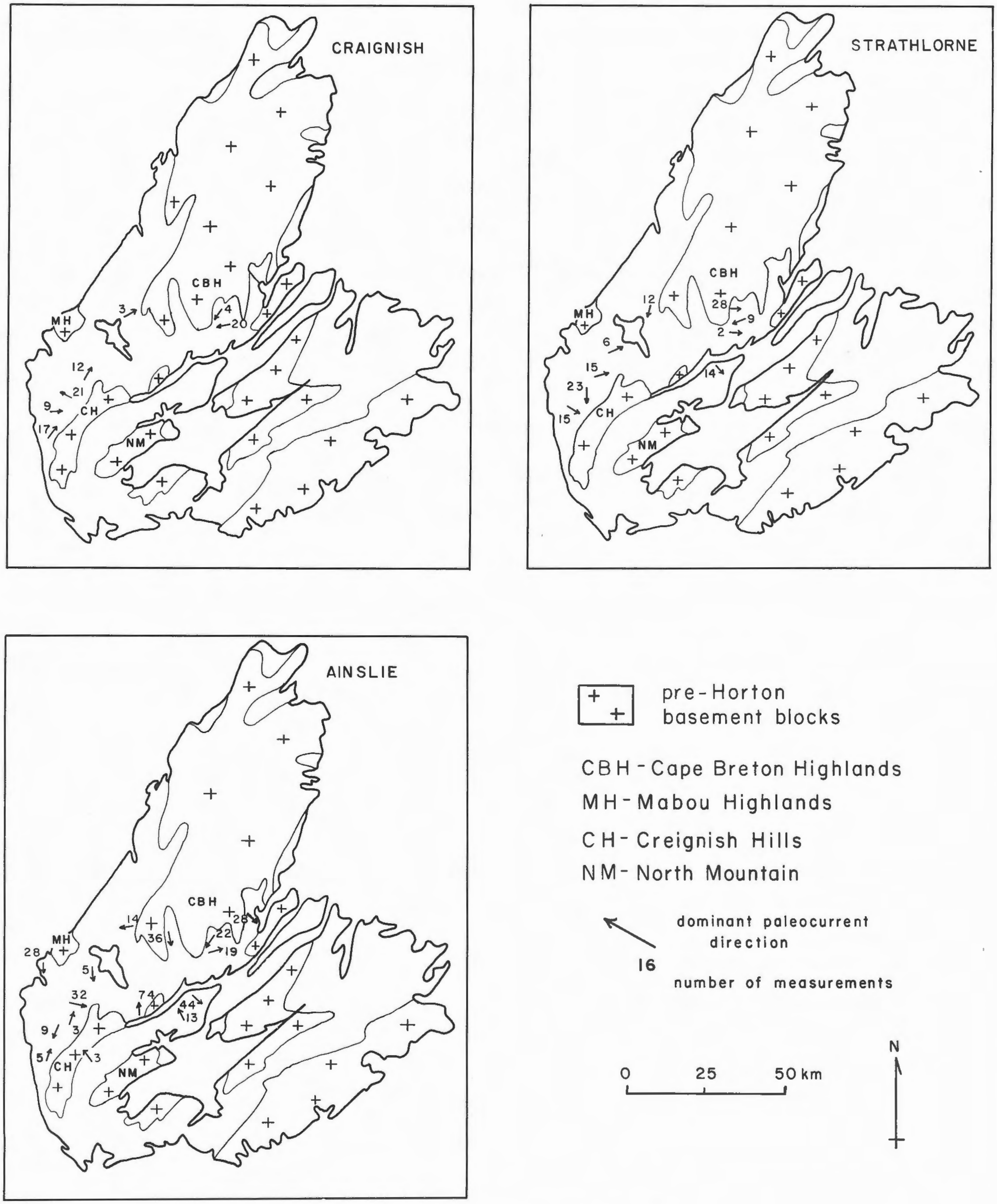

Figure 3. Summary of paleocurrent data.

in conglomerate. In Mabou area there was an overall dominance of Mabou Highlands block rather than Creignish Hills as a sediment source. Flow directions from basin margin conglomerates at Whycocomagh and Washabuck may indicate the basin margin followed the North Mountain block. In Margaree and Baddeck areas the Cape Breton Highlands appear to have been the dominant sediment source throughout Horton time.

RESOURCE POTENTIAL

Assessment of the economic potential of the Horton Group involves clear understanding of the interrelationships of depositional style, facies distribution and tectonic controls. The juxtaposition of dark grey fine grained facies and red coarser grained facies, confined in a localized structural basin and overlain by a regionally continuous carbonate/evaporite unit all represent favourable factors. However, details of the organization, geometry and predictability of the facies are crucial to further evaluation of their potential.

Both Craignish and Ainslie megafacies clastics could provide good aquifers for mineralizing fluids and/or good reservoirs for hydrocarbons. This might include well sorted fining-upward fluvial sandstone sequences, thick coarsening-upward conglomerate sequences, or coarsening-upward lacustrine shoreline sequences. Both the Strathlorne lacustrine magafacies and Windsor marine carbonate/evaporite sequences could act as source rocks (for metallic ions, or hydrocarbons) and as traps (as a redox interface to form a locus for precipitation of metallic ions, or as a seal rock to block further migration of hydrocarbon fluids.) Numerous structural features may also enhance resource potential.

Thus the overall tectonic, stratigraphic and depositional settings are suitable for several types of metallic deposits (Cu, Pb and Zn sulphides; placer Au and U) and petroleum accumulations (oil, natural gas, oil shale). The setting is reminiscent of the classic Kupferschiefer copper deposits of Europe (Kirkham, 1975) and, in fact, the upper few metres of the Horton Group typically display green staining and elevated Cu, Pb, Zn, U contents (Kirkham, 1978). Oil seeps are well known in the area and exploratory drilling has proceeded sporadically since 1869 with limited success (NSDME, 1986). The Stoney Creek oil and gas field in New Brunswick produces from partly correlative units in a similar depositional setting (Greiner, 1974). The setting is also similar to the oil shale-rich Green River Formation of Wyoming (Fouch and Dean, 1977) and the petroleum-rich Rotliegendes of Europe (Glennie, 1972).

ACKNOWLEDGMENTS

The work outlined here forms part of a Ph D study at University of Ottawa. Co-operation with geologists from federal, provincial, university and industry sources has aided the investigations and will ultimately allow compilation of much scattered data on the Horton Group.

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Levé héliporté magnétique et électromagnétique des monts Stoke, Québec, et interprétation géologique préliminaire†

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Schwarz, E.J., Palacky, G.J., Stephens, L.E., Dion, D.-J., Lefebre, D.L., Church, H., et Gravel, C., Levé héliporté magnétique et électromagnétique des monts Stoke, Québec, et interprétation géologique préliminaire; dans Recherche en cours, partie B, Commission géologique du Canada, Étude 88-1B, p. 23-27, 1988.

Résumé

Un levé héliporté détaillé a été exécuté dans la région des monts Stoke au Québec, à l'automne 1986. Cette région offre un potentiel intéressant puisque son milieu géologique suggère la présence de gisements polymétalliques. On dispose de la carte d'anomalies ÉM, de la carte du champ magnétique total et son gradient vertical, de la carte de conductivité électrique, et des profils de très basse fréquence (TBF). La corrélation géophysique-géologie s'exerce surtout du côté structural où on retrouve le grain tectonique appalachien et des axes synclinaux-anticlinaux sur le gradient vertical. La carte de conductivité montre des zones possibles de failles, graphitiques, minéralisées ou d'altération. Une zone de très forte conductivité a été vérifiée et confirmée au sol par des levés MaxMinI, ÉM-34 et des sondages électriques.

Abstract

A detailed helicopter-borne survey was carried out in the Stoke Mountains region of Quebec in fall 1986. This region offers interesting potential, since its geological environment suggests polymetallic deposits. Available information includes maps of EM anomalies, of the total magnetic field and its vertical gradient and of electrical conductivity, as well as very low frequency (VLF) profiles. The geophysics-geology correlation chiefly shows the structure, revealing the Appalachian tectonic grain and the synclinal-anticlinal axes on the vertical gradient map. The conductivity map shows possible fault, graphitic, mineralized or weathered zones. A zone of very high conductivity was checked and confirmed on the ground by MaxMinI and EM-34 surveys and by electrical well logging.

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INTRODUCTION

Dans le cadre du programme intitulé « Mesures fédérales relatives à l'amiante » (1984-1987) un levé géophysique héliporté détaillé a été effectué dans la région des monts Stoke. La firme AÉRODAT LTÉE a réalisé ce contrat. La région couverte se situe à 25 km environ au nord-est de Sherbrooke.

Géologiquement, la région des monts Stoke s'aligne structuralement dans le cadre général appalachien sur une droite N45°E allant du lac François au sud-ouest, vers la zone de Weedon et Scotstown au nord-est. Ces trois régions sont constituées des mêmes types de roches du Paléozoïque inférieur : des roches volcaniques, des sédiments et des granites. De plus, quelques gisements polymétalliques ont été exploités dans les régions au nord-est et au sud-ouest des monts Stoke (Harron, 1976). Le ministère de l'Énergie et des Ressources du Québec avait déjà couvert la région de Weedon et Scotstown au nord-est des monts Stoke par un levé héliporté MAG et ÉM (ministère de l'Énergie et des Ressources, 1984; ministère de l'Énergie et des Ressources, 1986). De même, la région des monts Stoke constituait une priorité pour un levé visant des cibles d'intérêt minier.

Levé héliporté

Ce levé a été effectué à l'automne 1986 à bord d'un hélicoptère remorquant deux coquilles séparées : une contenant le système ÉM actif et l'autre contenant deux magnétomètres à vapeur de césium et le récepteur TBF. La distance verticale entre les deux coquilles était de 20 m environ et l'altitude de vol du système ÉM était d'environ 30 mètres. Le système ÉM se composait de trois paires de bobines opérant à trois fréquences fixes entre 900 et 4 600 Hz. Une de ces trois paires de bobines est coplanaire, les deux autres sont coaxiales. Les réponses en phase et en quadrature reçues par la configuration coplanaire ont servi au calcul de la conductivité électrique du sol. L'arrangement des deux capteurs magnétiques a permis la mesure du gradient magnétique vertical ainsi que la valeur absolue du champ géomagnétique. L'espacement des lignes de vol était de 100 m.

Résultats du levé héliporté

Les résultats du levé sont présentés sur six cartes disponibles dans le dossier public 1591 (Commission géologique du Canada, 1987) sous les rubriques suivantes :

1. Carte en couleurs à 1/50 000 du champ magnétique total (C21312G).
2. Carte en couleurs à 1/50 000 du gradient magnétique vertical (C41312G).
3. Carte en couleurs à 1/50 000 du gradient magnétique vertical (C25037G).
4. Carte de contours à 1/20 000 du champ magnétique total (21311G).
5. Carte de contours 1/20 000 du gradient magnétique vertical (41311G).
6. Carte de contours 1/20 000 de la conductivité apparente et symboles de conductance (25038G).

Au verso des cartes C21312G et C41312G paraissent les résultats du levé TBF. Les données enregistrées le long des lignes de vol sont aussi disponibles sous forme de microfiches (25039G).

Les cartes montrent une forte variation des paramètres géophysiques mesurés ou calculés. Ces écarts sont causés par les grandes variations géologiques de la région. Le premier objectif est d'établir une corrélation qualitative entre les cartes géophysiques et la géologie connue.

Interprétation

La description de la géologie des monts Stoke la plus détaillée et récente provient de de Römer (1985). Une reproduction de sa carte géologique paraît à la figure 1. Les cartes géophysiques reproduites en noir et blanc et à l'échelle réduite sont illustrées aux figures 2, 3 et 4. Les conclusions suivantes peuvent être tirées d'une comparaison de ces cartes :

1. Il n'y a pas de corrélation positive ou négative des anomalies majeures de conductivité électrique (fig. 2) et de susceptibilité magnétique du sol tel qu'indiqué par les cartes magnétiques (fig. 3 et 4).

2. La carte du gradient vertical (fig. 3) montre des anomalies magnétiques causées par les sources (concentrations de magnétite) moins profondes. Les axes des anomalies suivent le grain structural de la région ce qui prouve que les sources de ces anomalies sont reliées aux structures apppalachiennes. Du côté lithologique, il n'y a pas de corrélation évidente. La masse granitique au sud-est du massif montre peu de variation. Les anomalies plus fortes dans les parties ouest et nord du massif seront causées par les laves basiques de la formation d'Ascot-Weedon. Mais les anomalies fortes du gradient vertical dans la partie nord-ouest de la région n'ont rien à voir avec la lithologie montrée à la figure 1. Puisque les roches sédimentaires affleurantes ne possèdent qu'une susceptibilité magnétique basse ($\sim 5 \times 10^{-4}$SI) les sources des anomalies devront être cherchées à profondeur. Du côté structural, on observe que les anomalies positives du gradient et du champ total (fig. 4) tendent à correspondre aux axes anticlinaux indiqués sur la carte géologique. Cette corrélation suggère un outil possiblement puissant pour une future révision de la géologie structurale de la région. Par exemple, l'anomalie forte du gradient au sud-est de la rivière Stoke suit généralement la bande des grès feldspatiques du groupe de Magog (unité 3b de Römer, 1985) qui est flanqué des deux côtés par des sédiments (unité 4) plus jeunes. L'anomalie pourrait donc indiquer le noyau d'un anticlinal dont l'axe plonge vers le nord-est. Le gradient fort et positif semble indiquer que les laves appartenant à la formation d'Ascot-Weedon datant de l'Ordovicien inférieur à moyen, approchent la surface. Il s'agit des seules roches magnétiques dans la colonne stratigraphique établie par de Römer. Dans la partie sud-est de la région, soit le massif des monts Stoke, ces laves se trouvent à la surface et donnent naissance à des anomalies fortement positives du gradient vertical et du champ magnétique total (fig. 4).

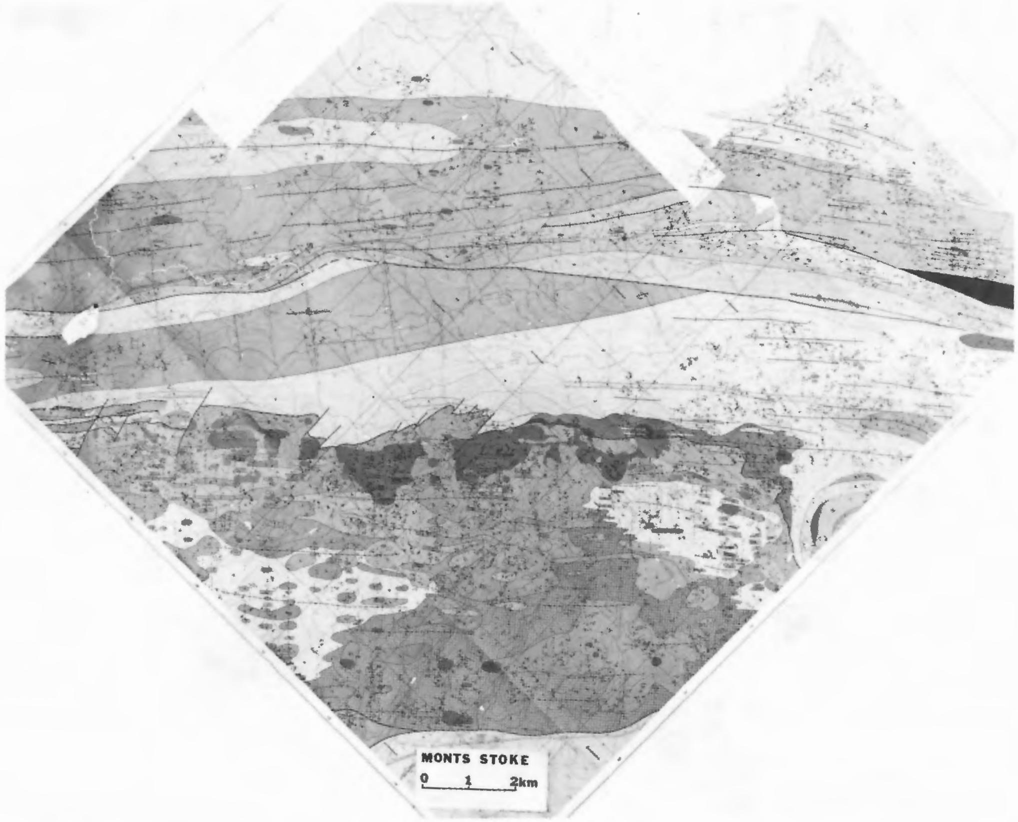

Figure 1. Reproduction de la carte géologique des monts Stoke (de Römer, 1985) originalement publiée à l'échelle de 1/20 000.

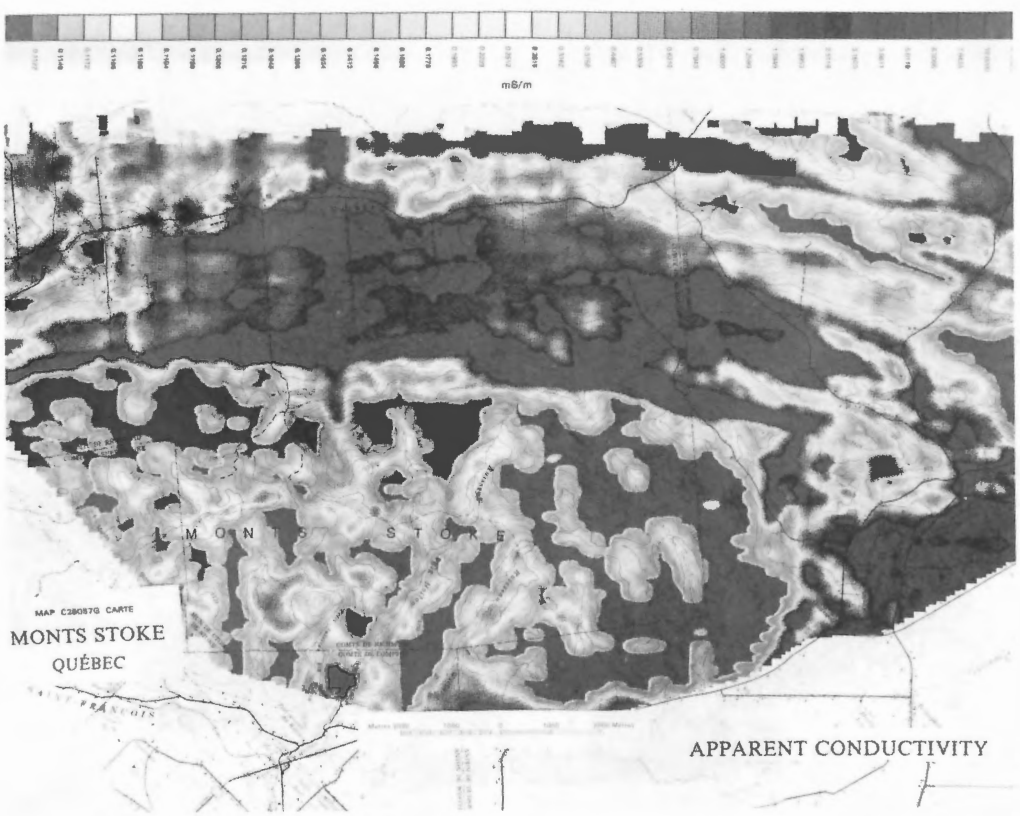

Figure 2. Reproduction de la carte en couleurs, à 1/50 000, de la conductivité apparente.

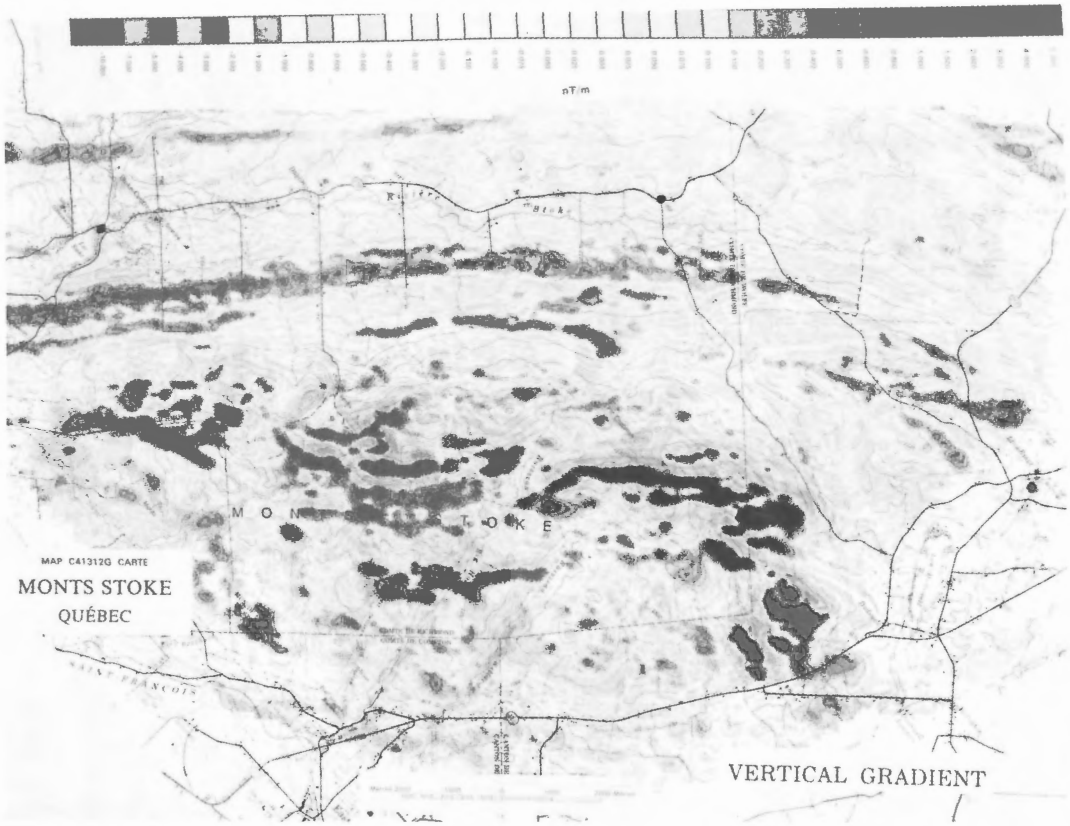

Figure 3. Reproduction de la carte en couleurs, à 1/50 000, du gradient magnétique vertical.

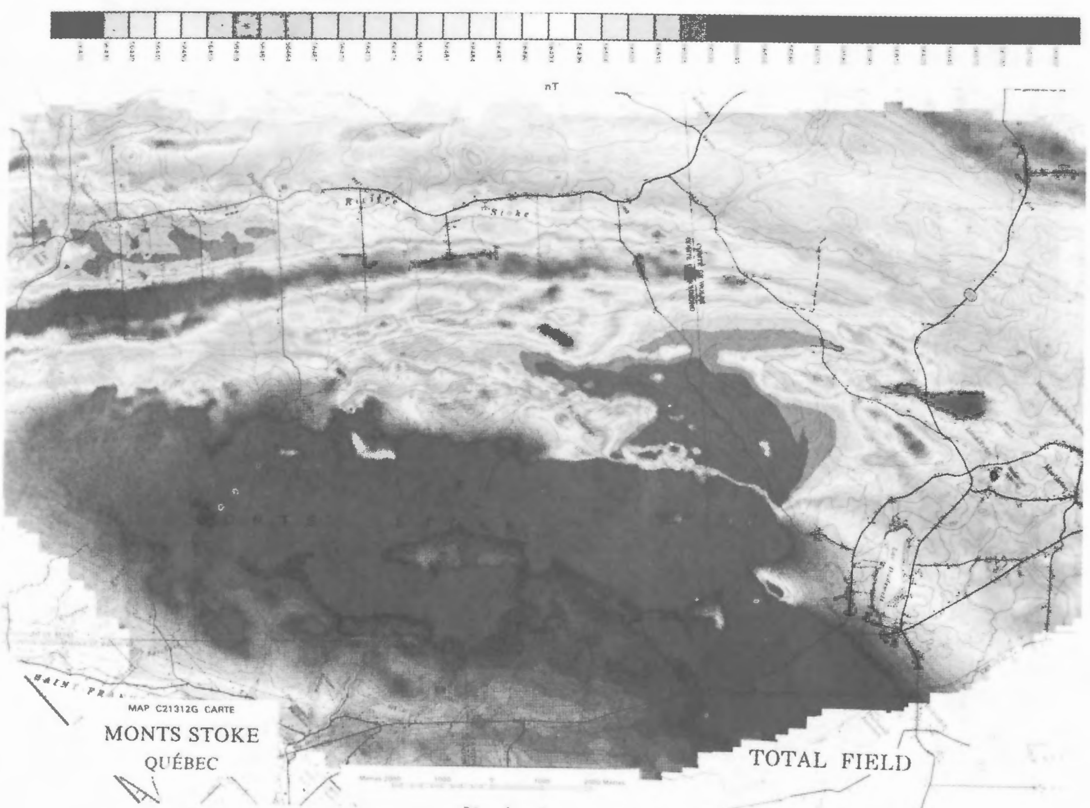

Figure 4. Reproduction de la carte en couleurs, à 1/50 000, du champ magnétique total.

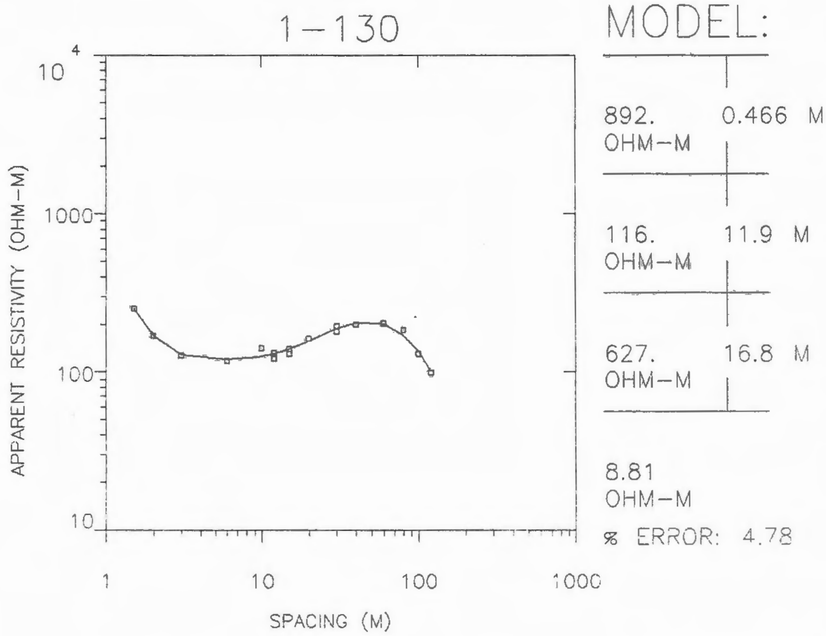

Figure 5. Profil de la résistivité apparente obtenu avec un appareil RSP-6 à 2km au sud-est de Duplin. En dessous de la couverture quaternaire formée par trois couches à résistivité différente, on observe une conductivité beaucoup plus élevée qui coïncide avec la zone conductrice indiquée sur la carte montrée à la figure 2.

3. La conductivité apparente de la région occupée par les roches volcaniques et granitiques de la formation d'Ascot-Weedon dans la partie sud-est est basse par rapport à celle des sédiments au nord et à l'ouest du massif. Mais les autres relations entre la lithologie et la structure d'un côté et la conductivité de l'autre demeurent obscures. Le grain de la conductivité correspond à celui de la géologie ce que indique que le patron majeur de la conductivité n'est pas déterminé par le type et l'épaisseur des dépôts meubles. Les causes possibles sont des failles, des zones graphitiques ou minéralisées inconnues dues au manque d'affleurements, ou la manifestation de couches argileuses, soit dans les dépôts meubles, soit formées sur place par l'altération des roches solides (saprolites).

On doit noter la zone de conductivité et de conductance très élevée sur le flanc nord-ouest et nord du massif. Une étude préliminaire de cette zone a été conduite sur le terrain pour en vérifier la présence, la nature et l'étendue.

Traverses et sondages électriques

On a procédé à une vérification des données électromagnétiques du levé aéroporté à deux endroits sur les flancs nord et nord-est du massif des monts Stoke, où une forte anomalie positive de la conductivité est indiquée. Cette vérification s'est faite à l'aide d'appareils Geonics E-31 et EM-34, Scintrex RSP-6, en utilisant le dispositif Schlumberger, et un appareil APEX MaxMinI à bobines horizontales avec une séparation de 80 m et huit fréquences distribuées entre 110 et 14 000 Hz.

Les traverses et sondages effectués sur le flanc nord, à 2 km au sud-est du Duplin, semblent indiquer la présence d'une couche de 30 m d'épaisseur formée par des sédiments quaternaires avec une conductivité de 2 mS/m (fig. 5). En dessous de cette couche, une conductivité plus élevée de 100 mS/m environ a été détectée, phénomène qui coïncide avec la zone de haute conductivité indiquée sur la carte (25037G, fig. 3). L'existence de la zone conductrice est donc confirmée mais dû au manque d'affleurements, son origine demeure inconnue.

Sur le flanc nord-est du massif, on a effectué des traverses et sondages le long du chemin Lessard où la zone conductrice indiquée par la carte traverse le chemin. Un affleurement presque continu le long du chemin est formé par des schistes contenant de la pyrite disséminée et altérée ainsi que des zones fortement cisaillées. Deux zones de haute conductivité ont été détectées, une sone résistante séparant ces dernières. L'existence de la zone conductrice détectée par le levé héliporté est donc confirmée. Il ne fait aucun doute que la source de haute conductivité se trouve dans les roches solides et il semble qu'elle soit reliée à la présence des sulfures disséminés et altérés et des zones de cisaillement et, possiblement, d'altération.

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Preliminary report on petrographic, palynological, and geochemical studies of coals from the Pictou Coalfield, Nova Scotia[1]

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Abstract

Detailed investigations of the petrography and palynology of coal seams from the Pictou Coalfield refine understanding of their depositional environment and local chronostratigraphy. A scheme for coal facies analysis based on macerals, rather than lithotypes, developed for Permian Gondwana coals, is applicable to these late Carboniferous coals. Peats were deposited on lake margins in subenvironments ranging from marsh to wet forest swamp, as wetting and drying trends alternated. Identification of these subenvironments is corroborated by shifts in miospore associations. The coals were deposited between early Westphalian C and early Westphalian D time. Since the coal-bearing strata fill the youngest graben known in the Maritimes Carboniferous Basin, this suggests an upper age limit for significant Alleghenian fault-movement in the northern Appalachians. Preliminary geochemical results suggest that the Pictou coals have locally anomalous As, B, Cd, Cu, Ga, Sb, Sc, and Zn contents.

Résumé

L'étude détaillée de la pétrographie et de la palynologie des filons houillers du champ houiller de Pictou nous a permis de mieux comprendre les milieux sédimentaires dans lesquels sont apparus ces filons et la chronostratigraphie locale. Un mode d'analyse des faciès houillers qui est basé sur l'examen des macéraux au lieu des lithotypes, et a été élaboré pour l'étude des charbons permiens du Gondwana, convient aussi à l'étude de ces charbons du Carbonifère supérieur. De la tourbe s'est déposée à la périphérie des lacs, dans des sous-environnements de caractère variable, allant de marais à des forêts marécageuses, durant des alternances de périodes d'humidité et de périodes de sécheresse. L'identification de ces sous-environnements est confirmée par des déplacements des associations de miospores. Les charbons se sont déposés entre le Westphalien inférieur C et le Westphalien inférieur D. Étant donné que les strates carbonifères comblent le plus récent graben connu du bassin carbonifère des Maritimes, il est possible qu'elles correspondent à la limite chronologique supérieure des failles relativement importantes de l'Alléghanien survenues dans le nord des Appalaches. Les résultats préliminaires des analyses géochimiques suggèrent que les charbons de Pictou sont caractérisés par des anomalies locales des concentrations de As, B, Cd, Cu, Ga, Sb, Sc et Zn.

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INTRODUCTION

The Pictou Coalfield lies in north-central Nova Scotia (Fig. 1, 2). Although current production is very small, estimated reserves are considerable, and recent deep drilling suggests even more potential. Because of their low sulphur content and relatively high rank compared to other Nova Scotia deposits, coals from the Pictou Coalfield have potential both for thermoelectric power generation and some metallurgical applications (Hacquebard 1979).

The geological setting of the Pictou Coalfield has been reviewed by Bell (1940), Hacquebard and Donaldson (1969), Yeo (1985a), and Yeo and Gao (1987). The Pictou Coalfield is restricted to Stellarton Graben, a small synsedimentary basin (Fig. 2). The coal is hosted by the "Stellarton Formation" (Stellarton Series of Bell, 1940). This is coeval with the lower Pictou Group (informally named the "Merigomish Formation" by Yeo, 1985b) north of the graben. The stratigraphy of these units is summarized in Table 1. Four orders of rhythmic sedimentation can be recognized in the "Stellarton Formation": members, beds, lithotypes/mesobands, and microlithotypes/laminations. The studies outlined here were done at the bed and lithotype level.

Hacquebard and Donaldson (1969), in a classic paper on coal depositional environments, described the lower "Stellarton" coals and presented a detailed microlithotype and palynological profile of the Scott seam. The petrography of the overlying Acadia seam was described by Hacquebard and Avery (unpublished report, 1976). The Foord seam was studied by Calder (1979). The petrography of "Stellarton" oil shales has been described by Kalkreuth and Macauley (1987).

Pioneering regional studies of late Carboniferous palynology were made by Barss et al. (1963) and Barss and Hacquebard (1967). An investigation of palynological variation among coal seams in the the Pictou Coalfield (Barss, unpublished report, 1966) proved that the Westville Member coals were older than the Albion Member coals.

Only proximate analyses have been done on Pictou coals before. No previous work has been done their trace metal content.

Studies of vertical and lateral petrographic and palynologic variation of selected coal seams in the "Stellarton" and "Merigomish" formations were undertaken, as part of the Stellarton Basin Analysis Project, to learn more about paleoenvironmental controls on coal quality in the Pictou Coalfield and to refine late Carboniferous chronostratigraphy. A preliminary study of the geochemistry of the coals was undertaken to: ascertain the geochemical character of the coals and test the potential of geochemistry for seam correlation, to compare the geochemistry of autochthonous and hypautochthonous coals from the same coalfield, and to assess the potential of the coals as a source for rare metals.

The Acadia, Foord, and MacKay seams respectively, were chosen as representative of the lower, middle, and upper coal-bearing members of the "Stellarton Formation"

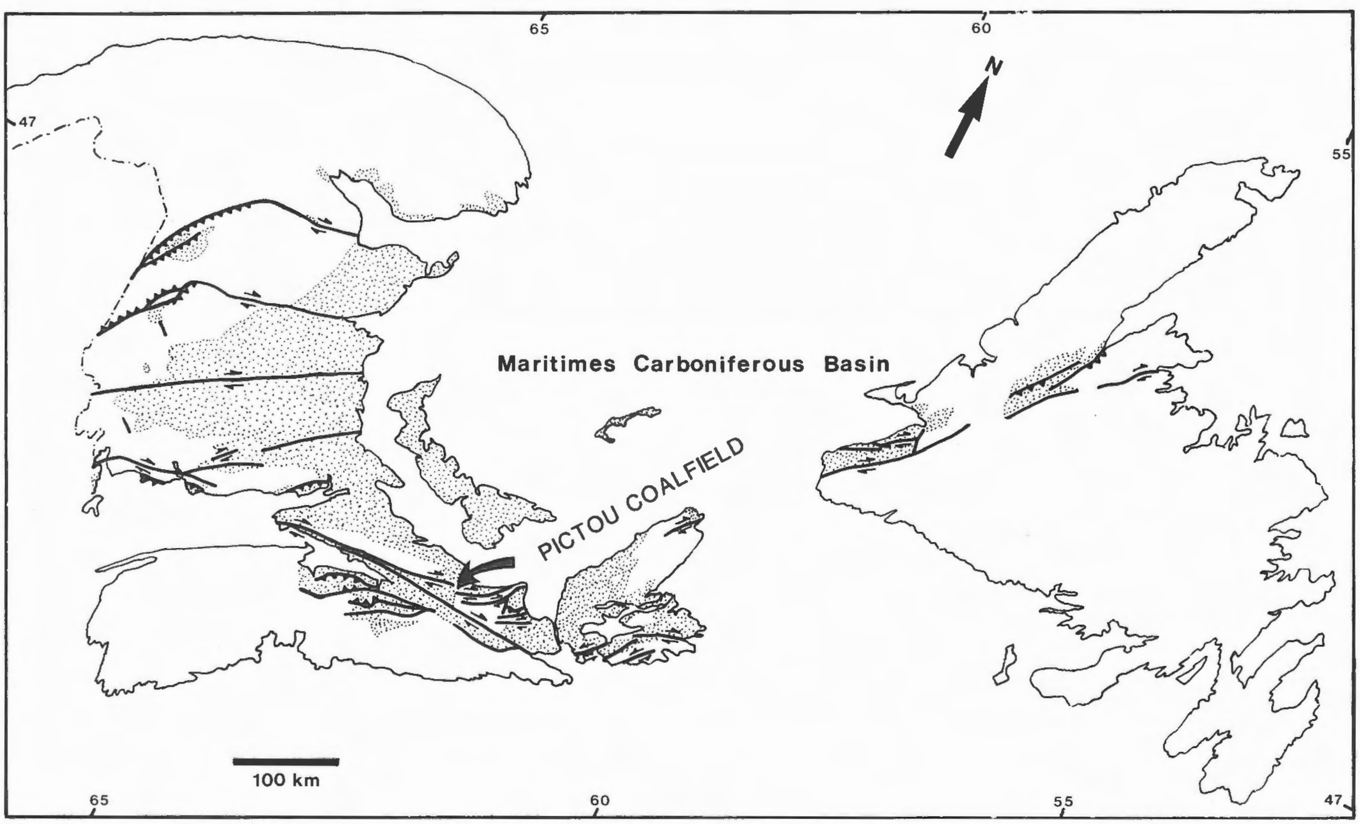


Figure 1. Location of the Pictou Coalfield. Carboniferous strata of the Maritimes Carboniferous Basin are indicated by the stipple pattern. Faults and their sense of displacement are after Keppie (1982).

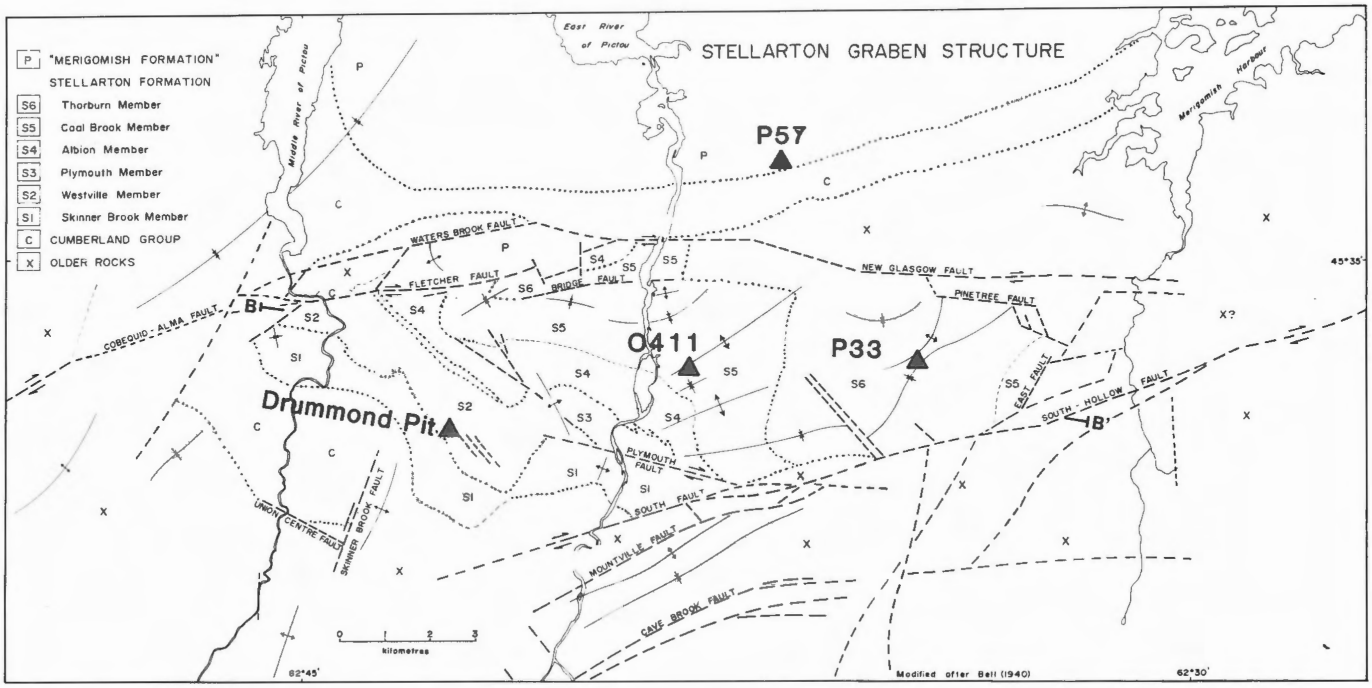


Figure 2. Sketch map of the Pictou Coalfield. Locations of boreholes discussed here are indicated by triangles. Cross-section B-B, (Fig. 3) is also indicated.

Table 1. Stratigraphy of the upper Westphalian Pictou Coalfield area. Thicknesses of "Stellarton Formation" members are maximum thicknesses (Bell, 1940). Thicknesses of "Merigomish Formation" members are from boreholes or estimated from mapping. The ages indicated, eWC, 1Wc, and eWD, refer to early and late Westphalian C and early Westphalian D substages.

AGE	"STELLARTON FORMATION"	"MERIGOMISH FORMATION"
		Caribou Member: −825m grey ss, red & grey sh, minor coal.
eW_D	Thorburn Member: <440 m grey sh, ss, oil sh, coal (includes MacKay seam).	Big Island Member: >350m grey ss, sh, minor oil sh & coal.
	Coal Brook Member: >850m grey & black sh, ss, oil sh.	
$1W_C$	Albion Member: >620m grey & black sh, oil sh, coal (includes Foord seam).	Smalls Brook Member: >110m grey sh, ss, ls, minor oil sh & coal (includes Munro seam).
	Plymouth Member: <295m red sh & ss.	
eW_C	Westville Member: <544m grey & black sh, ss, coal.	
	Skinner Brook Member: <655m red sh, ss, cgl.	

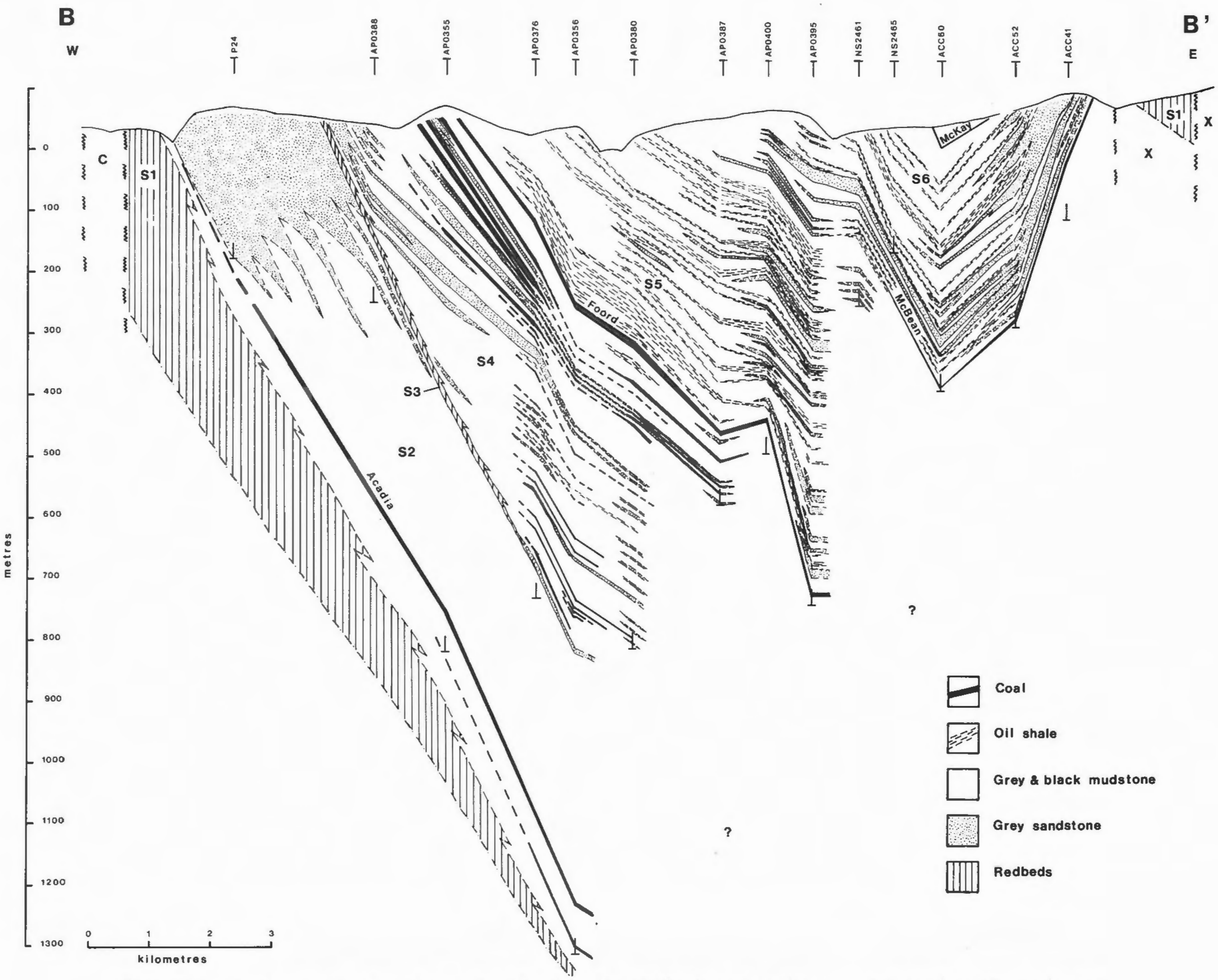
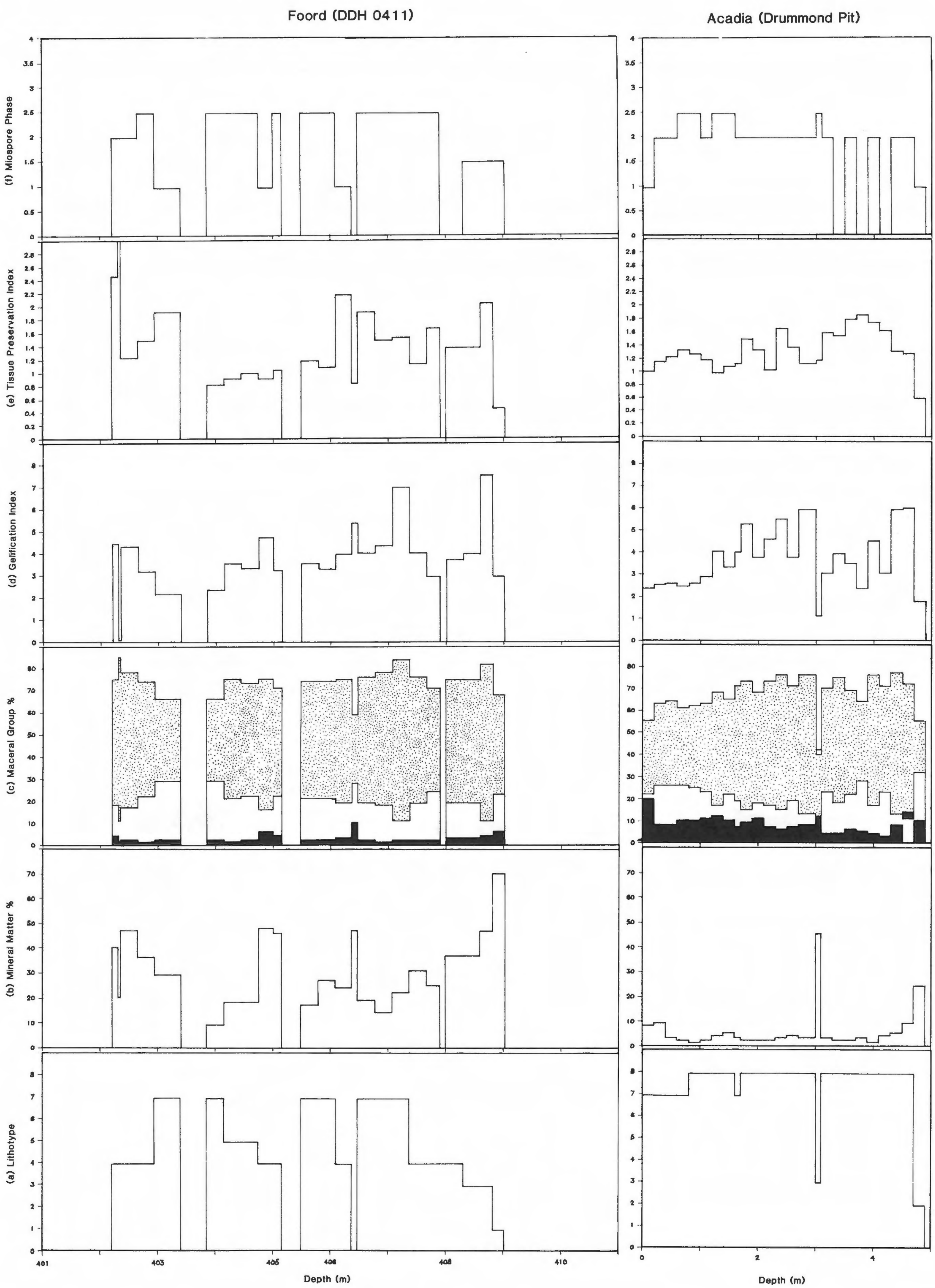


Figure 3. E-W cross-section along the axis of Stellarton Graben showing general stratigraphy and locations of coal seams discussed here. The location of the cross-section is shown in Figure 2. Vertical exaggeration, approx. 10.

(Fig. 3). The Munro seam, found near the base of the "Merigomish Formation" north of Stellarton Graben, was also investigated to contrast a coal seam of comparable age from a predominantly fluvial sequence with the lake-margin/deltaic "Stellarton" coals, and to test the hypothesis that the Munro seam might be correlative with one of the "Stellarton" coals. Except for a channel sample through the Acadia seam at the Drummond pit, all samples were split or whole core from boreholes (Fig. 2). Where possible, samples were collected according to coal lithotype variation following the system currently used by the Nova Scotia Department of Mines (Fig. 4a). Because the studies are still in progress and not all data are available, only one intersection through each of the seams investigated is shown here (Fig. 4 to 6). Lateral variation within seams is described but not illustrated.

Figure 4. Graphic petrographic logs through the Acadia (Drummond Pit) and Foord (borehole 0411) seams.
(a) Lithotypes are indicated by numerical values: 0) no data, 1) shale, 2) carbonaceous mudstone, 3) coaly shale, 4) shaly coal, 5) dull coal, 6) dull banded coal, 7) bright and dull banded coal, 8) bright banded coal, 9) bright coal.
(b) Mineral Matter. The proportion of clastic to organic material in the coal is indicated by the percent of mineral matter.
(c) Maceral groups are indicated by patterns: vitrinite by stipple; liptinite by solid shading; inertinite by no pattern. Their abundance is indicated as a percent of the organic component of the sample interval.
(d-e) Gelification (GI) and tissue preservation (TPI) indices (Diessel, 1986) are discussed in the text. Increased TPI indicates drier conditions and increased proportion of woody tissue (arborescent floral), while increased GI indicates wetter conditions. GI and TPI conditions for each seam are summarized in Figure 6.
(f) Miospore Phase. The dominant miospore phase (Smith, 1962) for each sample interval is indicated by numerical values: 1 and 4 both represent the Lycospore Phase, 2 represents the Transition Phase, and 3 represents the Incursion Phase. Intermediate or mixed conditions are indicated by intermediate values.


Foord (DDH 0411)
Acadia (Drummond Pit)
(f) Miospore Phase
(e) Tissue Preservation Index
(d) Gelification Index
(c) Maceral Group %
(b) Mineral Matter %
(a) Lithotype
Depth (m)
Depth (m)

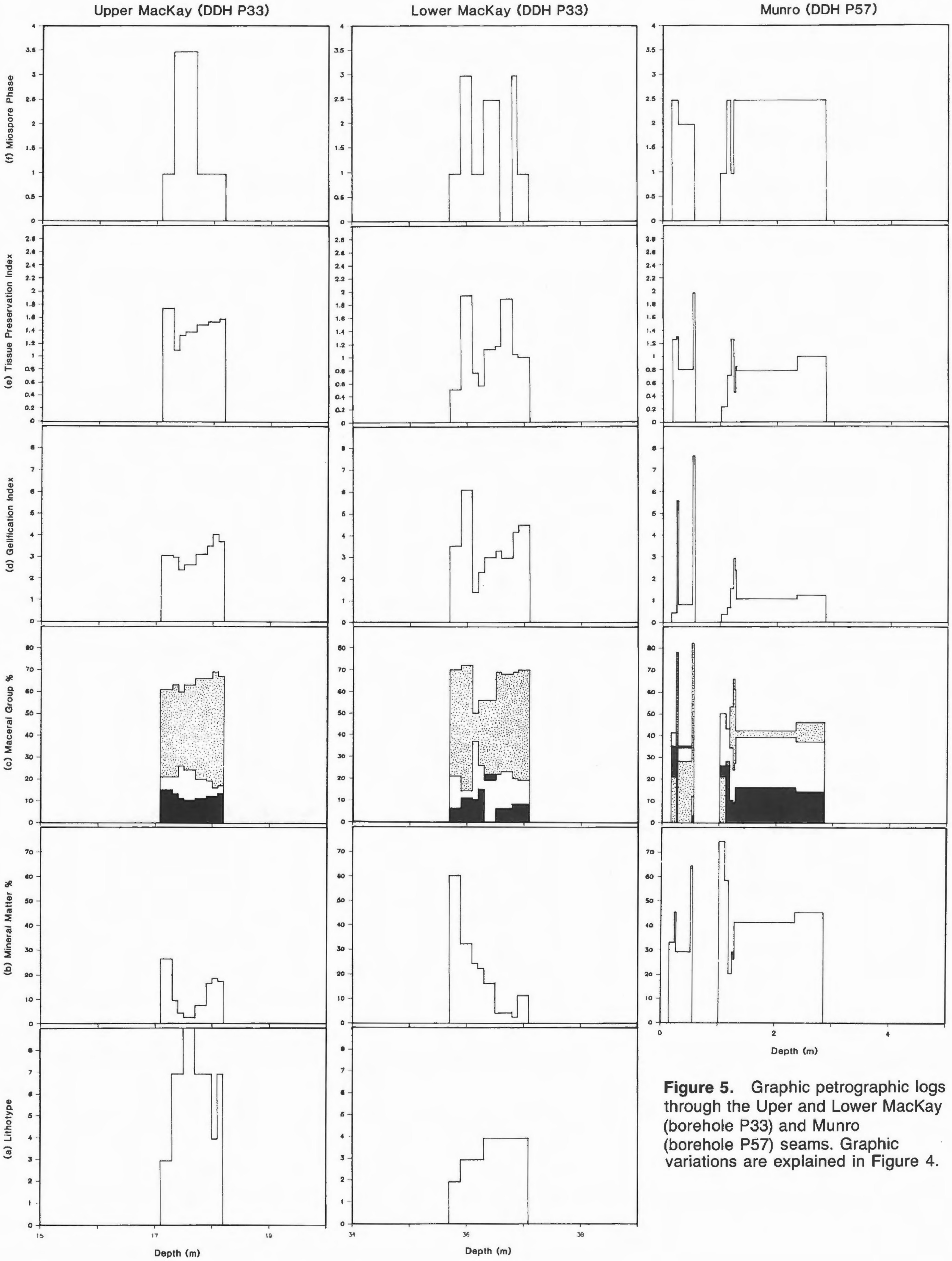


Figure 5. Graphic petrographic logs through the Uper and Lower MacKay (borehole P33) and Munro (borehole P57) seams. Graphic variations are explained in Figure 4.

COAL PETROGRAPHY

Sample preparation and petrographic examination were done by M. Tomica at the Institute of Sedimentary and Petroleum Geology. The maceral terminology of Bustin et al. (1985) is followed here except that vitrinite is subdivided into vitrinite A (clean banded vitrinite) and vitrinite B (matrix vitrinite). Petrographic variation diagrams (Fig. 4, 5) and coal facies diagrams (Fig. 6; Diessel, 1986) were constructed from the maceral data. These were used to interpret the depositional environments of the coals and to document changes during peat accumulation. Vitrinite reflectance measurements were also made on many of the samples. Vertical variation of specific macerals is duscussed, but not shown here. A detailed report on petrographic variation of the Pictou coals is in preparation (Yeo and Kalkreuth, in prep.).

Diessel (1986) showed how ratios of specific maceral combinations (see below), used as indicators of plant material, and depositional and diagenetic conditions, define particular peat-forming environments (eg. limnic, forest swamp, etc.). He defined two indices: the gelification index (GI), the ratio of gelified to non-gelified maceral material, and the tissue preservation index (TPI), the ratio of tissue-bearing macerals in which structure is preserved to those in which it is not preserved. Because of the way in which vitrinite was subdivided in this study, Diessel's (1986) indices have been slightly modified here:

GI = (vitrinite + macrinite)/(semifusinite + funisite + inertodetrinite)

TPI = (vitrinite A + semifusinite + fusinite)/ (vitrinite B + macrinite + inertodetrinite)

An advantage of this sort of paleoenvironmental indicator is that maceral data, on which they are based, are more widely available and easily quantified than the more traditional microlithotype analyses done to interpret coal depositional environment (eg. Hacquebard and Donaldson, 1969).

Vitrinite is the dominant maceral group in all of the seams sampled. This suggests that the original peat accumulated under predominantly wet forest swamp conditions (Bustin et al, 1985).

Acadia seam

In the Acadia seam, the predominant form of vitrinite varies laterally between the section shown here (Fig. 4) and a borehole to the north. Inertinite varies inversely with vitrinite (Fig. 4c). Inertinite macerals show some positive correspondence with shale abundance (Fig. 4b). Liptinite is a minor component except in the lower part of the seam. Sporinite is the dominant liptinite maceral, but alginite and liptodetrinite are high in the lower part of the seam.

Coal facies variation (Fig. 4d, 4e, 6) suggests that alternating limnic and swampy conditions in the lower part of the seam gave way to mixed fen and forest swamp conditions (ie. increased drying and tree density). Peat accumulation was interrupted by a major flood, represented by a shaly band. Wet forest swamp conditions were resumed. These gave way to fen, and finally limnic conditions (ie. increased wetting and reduced tree density).

Foord seam

In the Foord seam, as in the Acadia seam, inertinite varies inversely with vitrinite, and liptinite is a minor maceral group (Fig. 4c). In two boreholes (not shown here) drilled in the central part of the basin, vitrinite B is the dominant vitrinite form, fusinite and inertodetrinite are the main inertinites, and sporinite and cutinite are the chief liptinite macerals. In contrast, in the borehole (DDH 411) illustrated here, drilled towards the southern edge of the basin, vitrinite A is the predominant vitrinite form, fusinite and inertodetrinite are the chief inertinite macerals, and liptodetrinite is the main liptinite maceral. The Foord seam in this borehole is a shaly coal.

Coal facies variation (Fig. 4d, 4e, 6) suggests alternating limnic and wet forest swamp conditions (ie. flooding and drying). Six episodes of peat accumulation can be recognized. Somewhat drier conditions towards the basin edge are suggested by lower liptinite content.

MacKay seam

In the lower MacKay seam, the dominant form of vitrinite is variable. Vitrinite B is the main form found in boreholes (not shown here) drilled towards the basin margin. Vitrinite A is more abundant toward the basin centre, where the borehole P33 (Fig. 5) was drilled. Liptinite displaces inertinite as the second most abundant maceral group towards the basin margin. The chief inertinite macerals are fusinite and inertodetrinite. The chief liptinite maceral is sporinite. Mineral matter is highest towards the northwest.

Coal facies variation (Fig. 5d, 5e, 6) suggests alternate drying and flooding under forest swamp, marsh, and lake margin conditions.

In the upper MacKay seam (Fig. 5), vitrinite A is the dominant maceral. The liptinite maceral group (mainly sporinite and cutinite) is less important than the inertinite group (mainly fusinite and inertodetrinite). Mineral matter is most abundant towards the northeast.

Coal facies analysis (Fig. 5d, 5e, 6) suggests that the upper Munro Seam was deposited under conditions ranging from marsh to forest swamp.

Munro seam

Subordinate inertinite and liptinite co-vary inversely with vitrinite (Fig. 5c). Sporinite and cutinite and the minor liptinite macerals co-vary inversely with mineral matter. Alginite is relatively high. Mineral matter content (Fig. 5b) is high compared to the "Stellarton" coals.

Coal facies variation (Fig. 5d, 5e, 6) suggests that nearly continuous flooding conditions must have prevailed. The upper part of the seam is highly variable. Coal facies variation suggests alternating wet forest swamp and limnic conditions. The Munro seam cannot be correlated petrographically with any of the "Stellarton" coals examined.

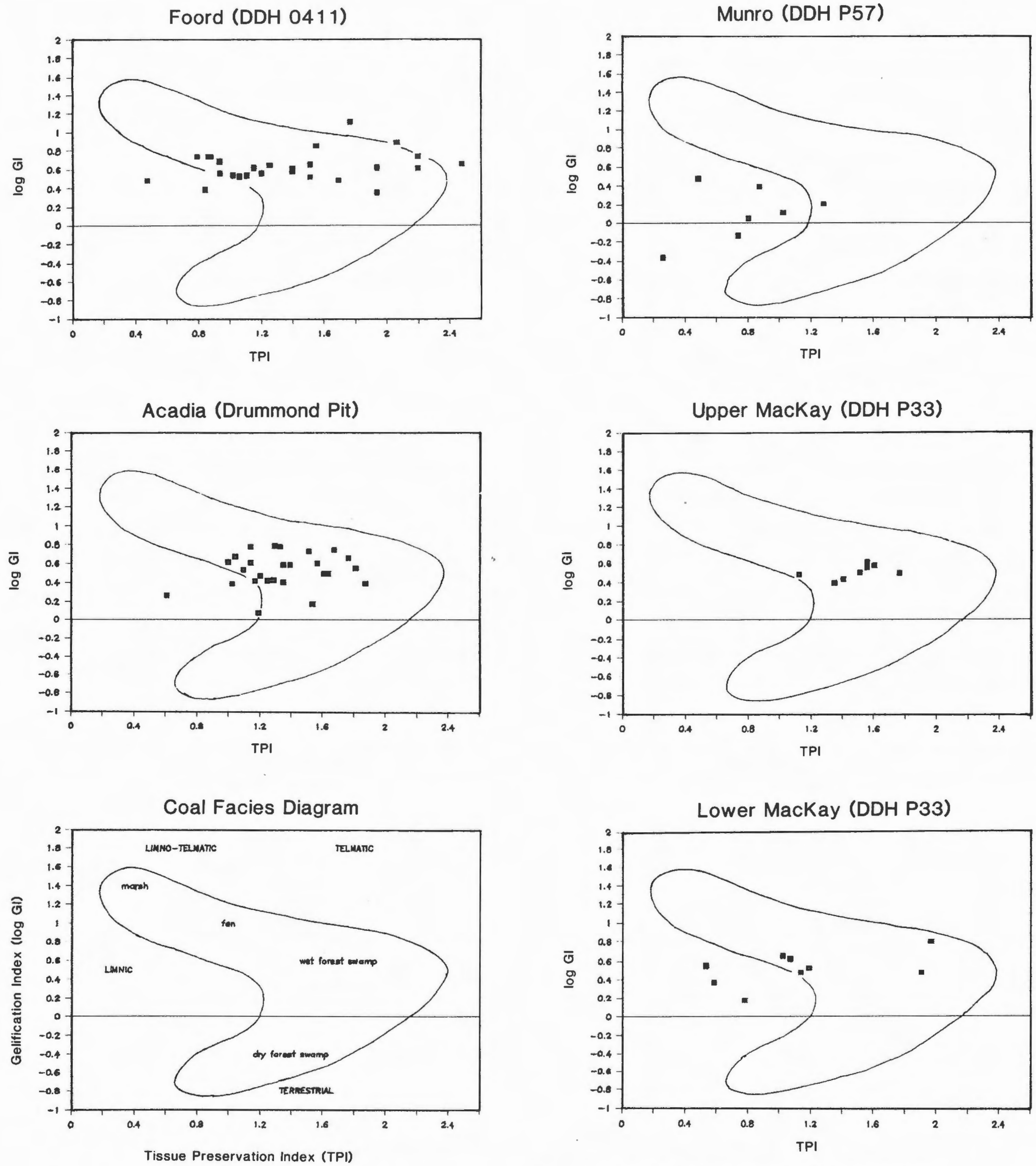


Figure 6. Coal facies diagrams (log GI vs TPI) for 5 seams from the Pictou Coalfield. For reference, the depositional environments from Diessel's (1986) coal facies diagram for Permian Gondwana coals (slightly modified) from the Sydney Basin of Australia is shown.

Coal rank

The high rank of "Stellarton" coals compared to other Maritime coals has been attributed to burial of the "Stellarton Formation" beneath thick cover, subsequently stripped away (Hacquebard, 1979; 1984). Coalification gradients for coals from the "Stellarton Formation" (Fig. 7) are much higher than one based on coals and shales in a deep borehole drilled through age-equivalent "Merigomish Formation" strata about 9 km north of Stellarton Graben. Since there is no evidence that the "Stellarton Formation" was more deeply buried than nearby "Merigomish" strata of about the same age, the higher rank of the "Stellarton" coals is more likely due to anomalous heating within the graben (Yeo, 1985a). This might be caused by deeply circulating ground water or frictional heating in the fault zones.

PALYNOLOGY

A total of 185 samples of coals and shales from the "Stellarton" and "Merigomish" formations were examined (Dolby, unpublished reports, 1986; 1987). Standard palynological preparation methods were followed. A modified specimen counting procedure was employed because: the predominance of *Lycospora* spp. in many samples masks the proportions of minor microflora, and the great variability in abundance of *Lycospora* within an individual seam profile limits its usefulness in coal seam correlation. The first 200 specimens

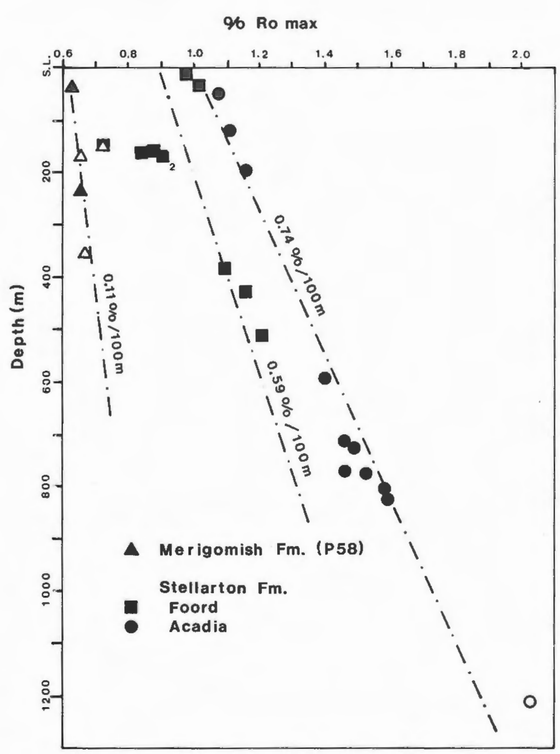


Figure 7. Coalification plots for the Acadia and Foord seams in the Stellarton Graben and coals and shales from a borehole drilled through the "Merigomish Formation" north of Stellarton Graben. Data of Hacquebard (1984) and Kalkreuth and Macauley (1987) are incorporated in this figure.

weere counted to determine the percentage of *Lycospora* present. A further 200 specimens were then counted, excluding *Lycospora*. The remainder of the slide was then scanned for trace species not already identified.

A provisional chart for the range of "index species" for the late Westphalian of the Pictou Coalfield has been developed (Figure 8). The preliminary nature of this scheme is emphasized. Further studies will probably result in revisions to it. Four zones are tentatively recognized. Zones I and III are approximately equivalent to Zone IX and X of Smith and Butterworth (1967), while Zone II is transitional. Zones I to III are correlative with Zone A, the *Vestispora* Zone, of Barss and Hacquebard (1967). Zone IV corresponds to Barss and Hacquebards' (1967) Zone B or *Torispora* Zone. Zones I to III are assigned to the Westphalian C, while Zone IV is Westpahlian D. This age assignment, based on species ranges from western Europe and central North America, indicates a slightly younger age range for the coal-bearing strata than that of Barss and Hacquebard (1967), which was based on the macroflora zonation of Bell (1938).

Smith (1962) recognized four miospore associations in the Lower Westphalian of Yorkshire, which he correlated with coal lithotypes and environments of deposition. The four associations are: 1) the Incursion Phase, developed during flooding (generally associated with a clastic influx), 2) the Densospore Phase, developed during dry conditions (peat raised above the water table), 3) the Lycospore Phase, developped in very wet conditions, and 4) the Transition Phase, intermediate between Densospore and Lycospore phases. Each phase may be repeated in a single seam, but the Lycospore Phase is always at the base. The Lycospore and Incursion phases are almost never adjacent to the Densospore Phase. The Transition Phase may separate two Lycospore phases. Although this scheme was developed on Westphalian A and B coals, it can be applied to the Pictou coals with certain limitations. Some species, such as *Densosporites*, are rare above the Westphalian B, while others, such as *Torispora*, common in the late Westphalian (Fig. 8), were not considered in Smith's (1962) sheme. The Densopore Phase is not recognized in the Pictou coals. Miospore phase variation for the coals discussed here is shown in Figures 4f and 5f.

The Acadia seam is assigned to Zone I ("subzones" C5-C6 of Fig. 8), and is therefore upper Early Westphalian C. The Lycospore Phase is restricted to the bottom and top of the seam (Fig. 4f). Transition and Incursion phases alternate through the central part of the seam. This supports the petrographic evidence for alternating drier (forest swamp) and wetter (fen) conditions during peat accumulation, except very wet (lake-margin) conditions at the beginning and end of the episode.

The Foord seam, assigned to Zone II ("subzones" C9-C10 of Fig. 8), is of lower Late Westphalian C age. As noted by Barss (1966) and others' *Torispora* is common in this seam. Lycospore, Transition and Incursion phases alternate throughout (Fig. 4f). This also supports the petrographic evidence for alternating wet (lake margin) and dry (forest swamp) conditions during peat accumulation.

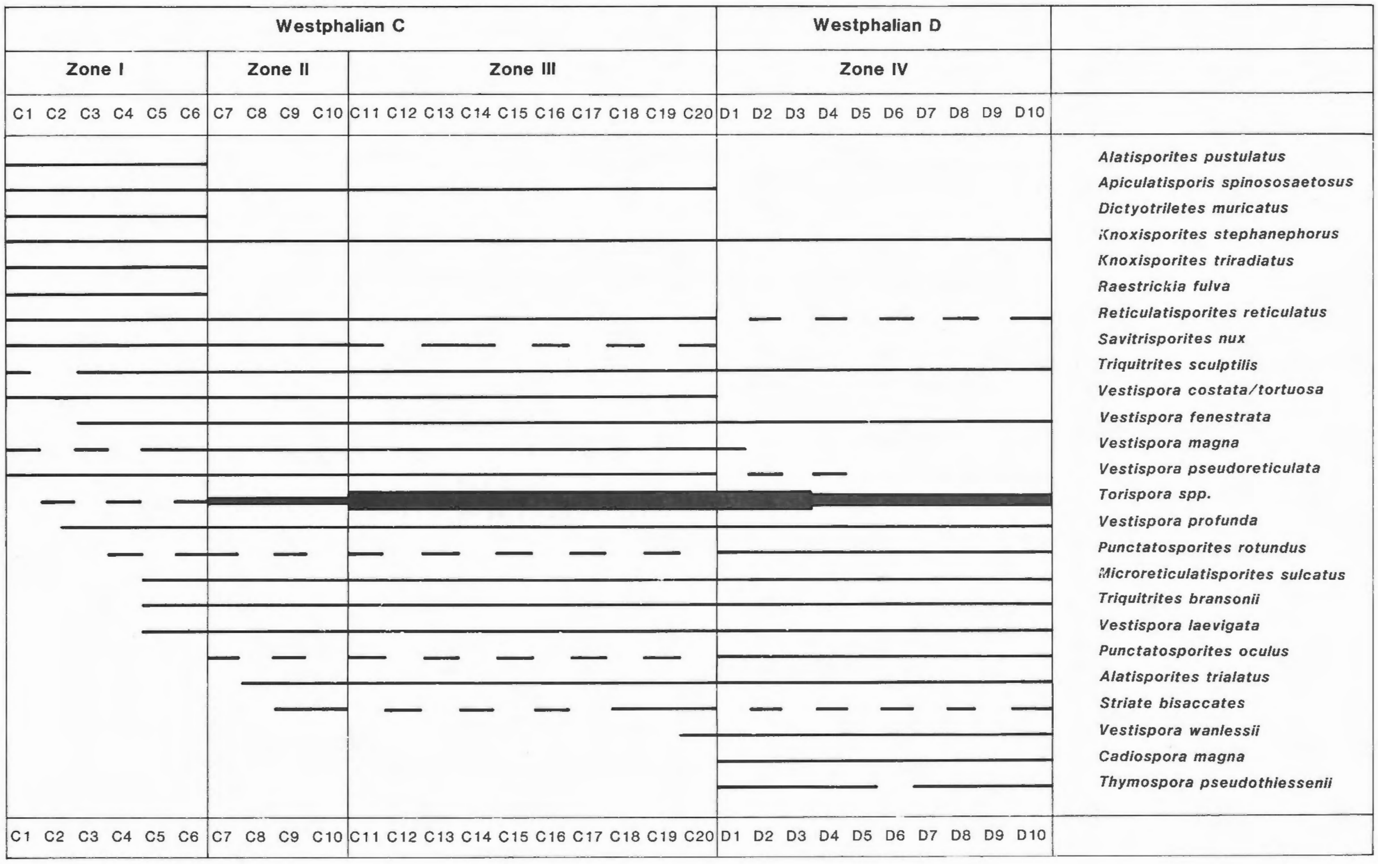


Figure 8. Preliminary range zones for Westphalian C and D miospores from north central Nova Scotia (Dolby, 1987). Relative abundance is indicated by line thickness.

The upper and lower MacKay seams are assigned to Zone IV (''subzone D1'' of Fig. 8) and are therefore earliest Westphalian D. In upper MacKay coals the Lycospore Phase predominates with some Incursion Phase influence (Fig. 5f). The lower MacKay seam shows more variability. Incursion, Lycospore, and Transition phases were recognized (Fig. 5f). This supports coal facies evidence for the broader range of depositional conditions under which lower MacKay peats accumulated.

The Munro seam, assigned to Zone II (''subzones'' C7-C9 of Fig. 8), is of middle Westphalian C age. In contrast to the Foord seam, which is about the same age, *Torispora* is uncommon. It is abundant, however, in a thin coal above the Munro seam. Incursion or Transition Phases alternate with the Lycospore Phase in the Munro seam (Fig. 5f). This supports petrographic evidence for alternating wet and dry conditions.

COAL GEOCHEMISTRY

Preliminary geochemical and SEM studies have been done on 16 samples of coals from the ''Stellarton Formation'' and 6 from the ''Merigomish Formation'' (Birk and Pilgrim, 1986). Only the minor element results are discussed here.

Trace metal concentrations were found to be extremely variable both among and within seams (Table 2). The concentrations of some metals (As, B, Cd, Cu, Ga, Sb, Sc, and Zn) were found to be well above their clarke values. Zn, As, Cd, Ga, and Sc are particularly concentrated, although still within the known ranges for coals (Bouska, 1981, Table 39). In general, metal contents were lower than those reported for Sydney Basin coals (Birk et al, 1986). Zn, As, Cd, and Ga were generally higher in the fluvial, ''Merigomish'' coals than in the limnic, ''Stellarton'' coals. In the ''Stellarton'' coals, As, B, Ga, Sc, and Zn abundances show inverse correspondence with ash content (ie. they generally increase with organic content). Ge, which is typically concentrated in limnic coals (Bouska, 1981), was not analyzed. Trace metal variation within the sections whose petrography and palynology have been summarized here is currently being investigated.

SUMMARY

Earlier coal studies showed that ''Stellarton'' coals were deposited under very wet, lake margin conditions. The coal maceral study outlined here complements this work and documents shifts between subenvironments within the lake margin setting as water levels and plant assemblages changed.

Table 2. Selected trace element data for coals from the Pictou Coalfield (from Birk and Pilgrim, 1986). Analyses were done on coal ash, except those indicated by *, which were done on whole coals. Analysis was done by neutron activation on ash left after heating at 500°C for 24 hours, except Cd and Cu, which were done by DC plasma spectrometry.

Seam:	Sample:	Ash %	As ppm	B ppm	Cd ppm	Cu ppm	Ga ppm	Sb ppm	Sc ppm	Zn ppm
CLARKE VALUES			5	10	0.18	70	19	1	5	40
"Merigomish Formation" coals:										
CARIBOU	YCM001	22.0	180	53	15	350	34	46	35.8	240
	YCM011	13.1	1400	180	380	800	27	150	14.5	58000
COAL PT.	YCM002	5.56	1100	210	180	400	58	98	16.7	21000
TONEY R.	YC23	58.0	380	16	1400	>4000	120	24	2.21	150000
MUNRO	YCM033	39.9	130	58	5.2	130	46	9.1	16.5	420
	4YMUN	41.7	46	82	0.6	24	47	3.9	23.5	180
"Stellarton Formation" coals:										
U. MACKAY	T81-10	20.1	19	100	5.0	210	30	3.0	27.6	1700
	T81-12	46.5	87	89	0.2	64	35	4.2	28.9	340
	T81-85	22.9	190	100	1.0	58	44	3.7	25.7	970
	T81-88	37.2	190	89	1.6	120	45	15	28.5	400
L. MACKAY	T81-13	62.1	6.6	71	4.6	95	20	2.7	28.0	1000
	T81-15	13.8	940	100	3.0	190	<16	36	18.2	1700
	T81-91	20.6	290	89	1.6	95	28	3.3	15.2	1100
	T81-92*	9.0	72	16	na	na	<8	0.3	1.1	340
FOORD	P83-5	16.5	25	120	<0.2	90	27	3.4	21.8	210
	P83-17	15.3	32	150	<0.2	91	<34	3.8	23.2	140
	P883-26	23.0	14	100	<0.2	96	57	4.3	33.8	210
	PE254-74*	20.8	4.2	25	na	na	<8	0.8	6.4	57
ACADIA	PE9-81	43.2	150	65	180	160	30	8.3	29.6	29000
	PE1-81*	14.2	0.4	16	na	na	<12	0.4	2.2	490

Diessel's (1986) coal facies variation scheme, developed for Permian coals in eastern Australia, can be applied to late Carboniferous coals in eastern Canada, although some allowances must be made for possible differences in coal facies ranges, due mainly to different plant assemblages.

The new palynological data also refine earlier work. The "Stellarton Formation" records nearly continuous deposition from early Westphalian C to earliest Westphalian D time. Deposition of the "Merigomish Formation" (Pictou Group) began before late Westphalian C. This confirms the supposition that deposition of the predominantly fluvial "Merigomish Formation" was coincident with development of the Stellarton Graben (Yeo and Gao, 1987) and its infilling by predominantly lacustrine "Stellarton" strata. The "Stellarton Formation" is the youngest deposit which can clearly be related to late Paleozoic faulting in the northern Appalachians. It is therefore reasonable to suppose that the latter is the last significant (ie. basin-forming) Alleghenian deformation event in the northern Appalachians. This provides an upper age constraint on the timing of the Alleghenian Orogeny in this part of the Appalachians (see Suppe 1985, p. 450).

Many metals of growing importance to advanced industrial technology are concentrated in coals. Of these, Ge and Ga are of particular interest as coal ash is, or could become a major source for them. Ge is believed to be bound to the organic component of coals, while Ga is associated with ash (Bouska, 1981). Both show marked concentration locally in the Sydney coals (up to 830 and 2 600 ppm respectively). In spite of generally higher ash contents in the Pictou coals, Ga abundance is relatively low compared to the Sydney coals. The concentrations of metals in coal ash are also a potential environmental hazard.

ACKNOWLEDGMENTS

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Mafic-ultramafic occurrences in metasedimentary rocks of southwestern Newfoundland[1]

D. Fox[2] and J.T. Van Berkel[3]

Fox, D. and Van Berkel, J.T., Mafic-ultramafic occurrences in metasedimentary rocks of southwestern Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 41-48, 1988.

Abstract

In southwestern Newfoundland, thin layers and inclusions of mafic-ultramafic rocks occur within a quartzitic and semipelitic unit of a larger metasedimentary belt. The ultramafic rocks are generally altered to assemblages of serpentine, talc, carbonate, chlorite and tremolite-actinolite. Thin chromite layers and relict (igneous?) olivine are locally observed. The mafic rocks are metagabbros consisting of both massive and layered types and contain plagioclase, hornblende, minor quartz and secondary chlorite and zoisite.

Tectonic disruption of the mafic-ultramafic rocks is interpreted to occur in a wide shear zone which locally develops the "block in matrix" feature typically ascribed to mélange zones. The previous history and origin of the mafic-ultramafic rocks is uncertain. Regionally, however, they lie along strike of mafic-ultramafic rocks of the Glover Island complex to the north and the Long Range mafic-ultramafic complex to the south, which are both interpreted as being part of an ophiolite suite.

Résumé

Dans le sud-ouest de Terre-Neuve, on rencontre de minces couches d'inclusions de roches mafiques et ultramafiques à l'intérieur d'une unité quartzitique et semi-pélitique qui fait partie d'une vaste zone métasédimentaire. Généralement, les roches ultramafiques ont été altérées et ont donné naissance à des assemblages de serpentine, talc, carbonates, chlorite et de trémolite et actinolite. On observe localement de minces couches de chromite et des cristaux relictes (ignés?) d'olivine. Les roches mafiques sont des métagabbros tantôt massifs, tantôt stratifiés, et contiennent des plagioclases, de la hornblende, des quantités accessoires de quartz et de la chlorite et zoïsite secondaires.

On a interprété les dislocations tectoniques des roches mafiques et ultramafiques comme ayant eu lieu dans une vaste zone de cisaillement qui localemnent présente une structure de « blocs dans une matrice », que l'on attribue généralement aux zones de mélange. On connaît mal l'évolution et l'origine des roches mafiques et ultramafiques. Cependant, à l'échelle régionale, celles-ci suivent la direction des roches mafiques et ultramafiques du complexe de Glover Island au nord et du complexe mafique et ultramafique de Long Range au sud, interprétés tous deux comme faisant partie d'une suite ophiolitique.

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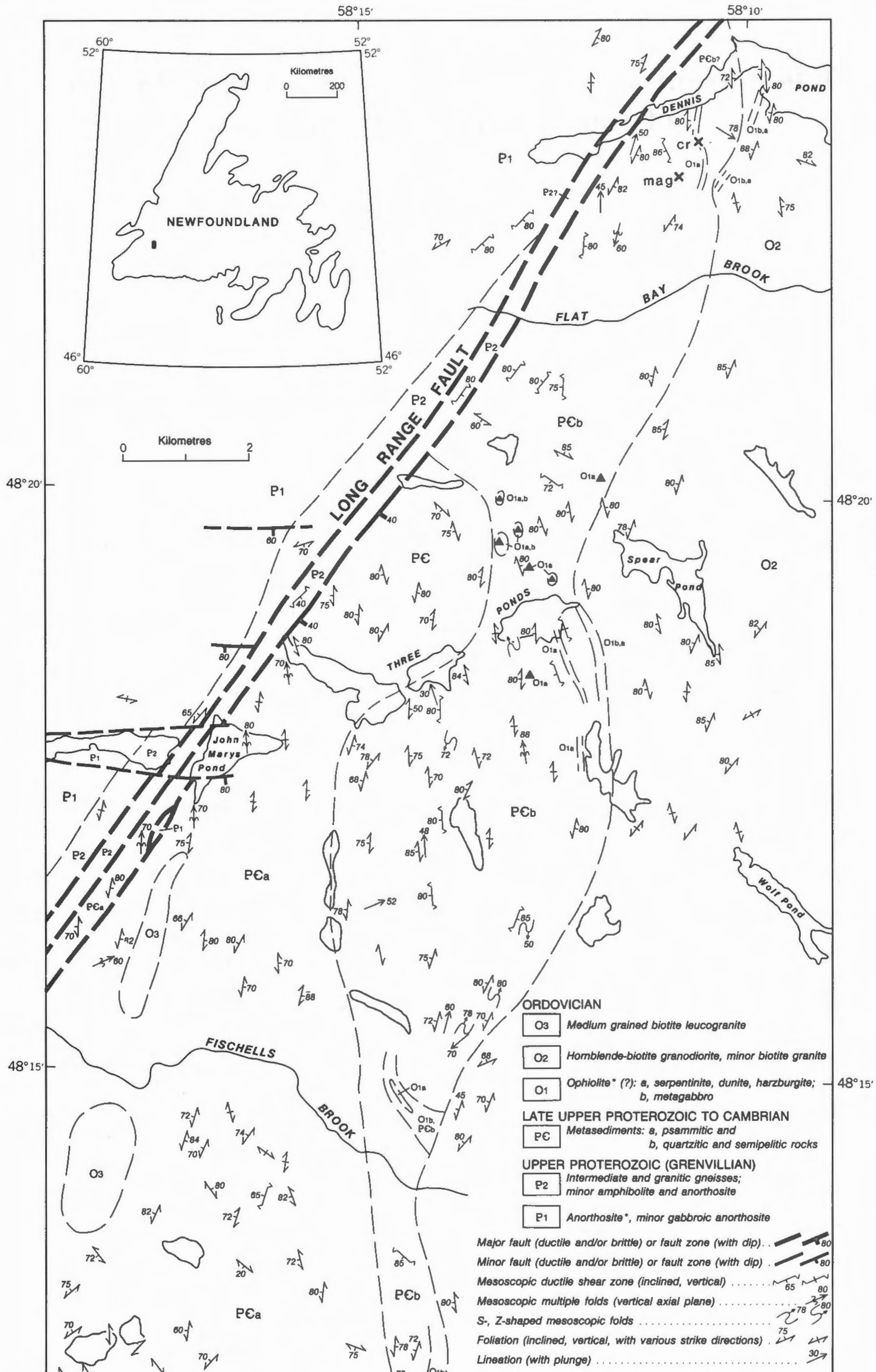

Figure 1. General geology of the study area, southwest Newfoundland.

INTRODUCTION

Reconnaissance mapping at 1:100 000 scale during the summer of 1985 outlined several outcrops of mafic-ultramafic rocks associated with metasediments in southwest Newfoundland (van Berkel et al., 1986). It was believed these rocks could have significant tectonic implications to the area, and therefore a more detailed study was undertaken in the 1986 and 1987 field seasons.

The map area is bounded by 58°10′ and 58°20′ west longitude and 48°10′ and 48°30′ north latitude. Field work was accomplished by traversing from helicopter supported fly camps. Information was plotted on 1:12 500 scale colour aerial photographs and later transferred to 1:50 000 scale base maps. Outcrop is near 90 % in upland areas but is poor to nonexistent in wooded valleys which transect the area in an east-west direction.

REGIONAL GEOLOGY

The map area (Fig. 1), has been divided into two distinctly different terranes separated by the Long Range Fault (Van Berkel et al., 1986; Van Berkel, 1987). To the west, the Steel Mountain Terrane (part of Williams' (1979) Humber Zone) consists of anorthosite with minor gabbroic phases (unit P1). To the east, the Central Gneiss Terrane (part of Williams' (1979) Dunnage Zone) is characterized by metasedimentary rocks (units PCa and b) intruded by voluminous amounst of granitoids (units 02 and 03). The metasedimentary rocks are further subdivided into two main belts, a western mainly psammitic portion (PCa) which continues southwest of the map area, and a less extensive eastern unit comprised mainly of quartzitic and semipelitic rocks (PCb). It is this unit (PCb) that hosts the mafic-ultramafic rocks which are the main focus of this study. The boundary between the metasedimentary units PCa and PCb is not well established but appears to be gradational. The exact boundary between the metasedimentary rocks (PCb) and the granitoids (02) to the east is extremely difficult to map because of the gradational intrusive nature of the granitoids. Generally, mapping is based on a percentage of dominant lithology.

The mafic-ultramafic rocks outcrop most extensively in three main areas: Dennis Pond, Three Ponds and North Fischells Brook (Fig. 1). They were also observed south of the Burgeo Road, east of the Long Range Fault (Van Berkel, 1987) but in this area outcrop is poor. The rocks occur in disrupted north-south trending belts which can be traced for over one kilometre. The occurence south of Fischells Brook is a body of metapyroxenite. This outcrop is not well studied and will not be dealt with in detail here.

DETAILED DESCRIPTIONS OF THE MAFIC-ULTRAMAFIC ROCKS

Dennis Pond area

In the Dennis Pond area (Fig. 2) mafic-ultramafic rocks form two lithologically distinct belts, a western belt consisting entirely of rusty brown to yellow weathered serpentinized ultramafic rocks (unit 01a), and an eastern belt composed mainly of metagabbro (unit 01b) and minor ultramafic rocks (unit 01a). In the western belt, serpentinite (unit 01a) either occurs as large (less than 20 m across) blocks within the metasedimentary unit (PCb), or forms a continuous 20-m-thick layer. It is interpreted that these blocks or layers were formed by tectonic disruption or thinning respectively of a serpentinite layer. Layering in the serpentinites is evident in some places, and varies from a few centimetres to one half metre in scale. In one sample observed in thin section, layers of mesh-textured serpentinized olivine alternate with layers of near 100 % serpentine. The origin of the layering is uncertain but the serpentine-rich layers may have been conduits for fluid movement through the rock so that the original mineralogy has been almost completely serpentinized, whilst the layers containing relict olivine received less exposure to fluids. Lesser amounts of talc and carbonate are also present. Layers of chromite (Fig. 3) occur in places as well as cross-cutting veins of chlorite and serpentine (chrysotile).

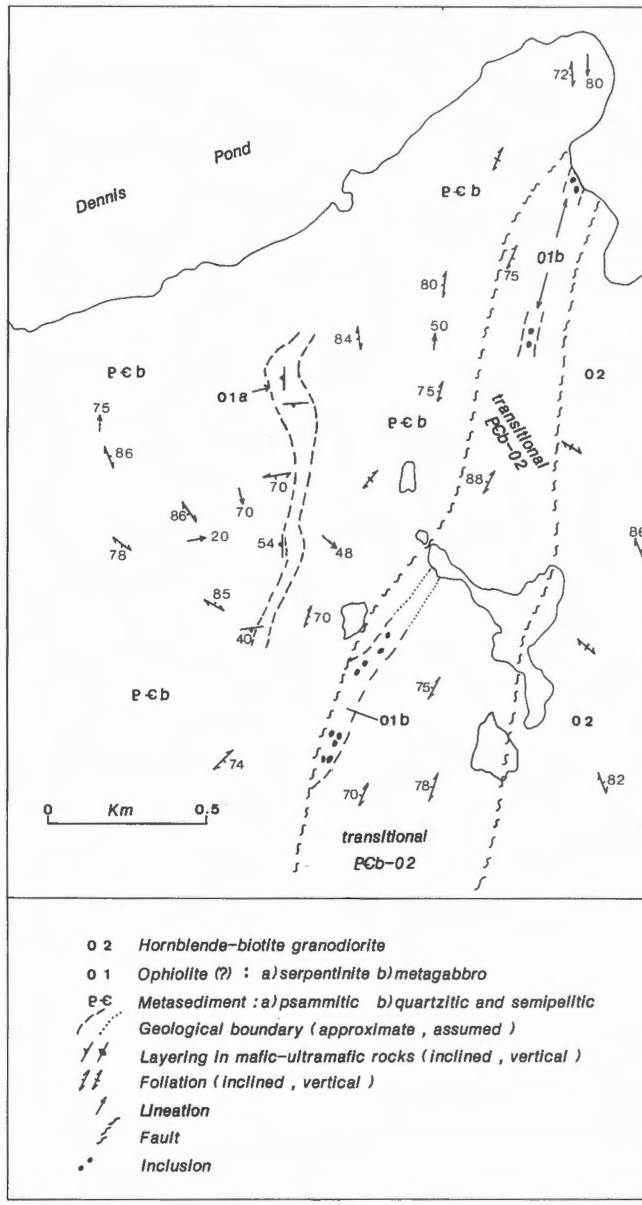

Figure 2. Detailed map of Dennis Pond area showing distribution of mafic-ultramafic rocks (units 01a and b).

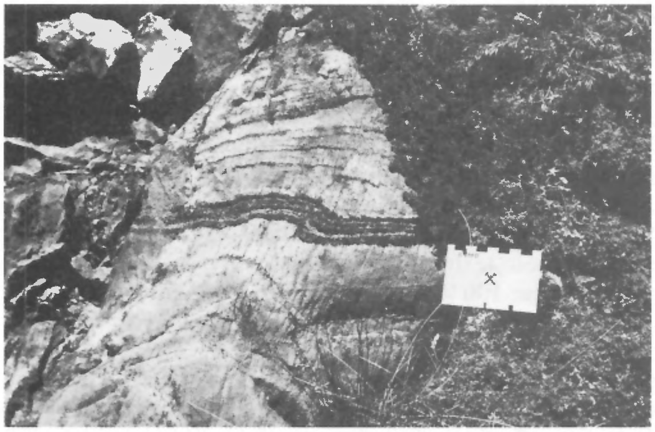

Figure 3. Chromite layers developed in serpentinite (unit 01a) from Dennis Pond area. Scale is 9 cm long. GSC-204460-I.

The extent of the eastern belt is poorly defined due to limited exposure. It lies near the transitional boundary between metasediments (unit PCb) and granitoids (unit 02) and the relations are further obscured by late-stage brittle faulting. The belt is mainly composed of small (less than 0.5 m across) metagabbro (unit 01b) inclusions intruded by late stage granitoids (unit 02). The metagabbro inclusions have a green and white mottled appearance on surface and consist of hornblende, zoisite, epidote and chlorite.

Three Ponds area

Immediately southeast of Three Ponds (Fig. 4), two belts of mafic-ultramafic rocks (unit 01) can be traced for over one kilometre southward. These occurrences are similar to those in the Dennis Pond area. The western belt comprises mainly disrupted serpentinite (unit 01a) layers and inclusions with lesser amounts of metagabbro (unit 01b) inclusions within the metasediments (unit PCb). Layering in the serpentinites is common (Fig. 5), generally strikes approximately north-south and is steeply dipping to vertical. Boudinaged mafic dykes have been observed in the metasediments but have not been seen to intrude the mafic-ultramafic rocks. Inclusions of the mafic dykes occur in several places.

The eastern belt is composed of gabbro-leucogabbro (unit 01b), layered gabbro (unit 01b) and minor serpentinite (unit 01a). These rocks occur as rounded to angular inclusions in a late granitoid which veins the area. The gabbroic rocks are generally composed of hornblende with smaller amounts of quartz, zoisite and epidote. Also occurring in this belt are distinct, dark green inclusions similar to those described below for the area north of Fischells Brook (Fig. 11).

North of Three Ponds, mafic and ultramafic rocks (unit 01) only occur as inclusions which in some areas are chiefly serpentinite while other areas consist mainly of gabbroic rocks. Figure 6 shows a large serpentinite inclusion within metasediments. An inner rim of talc and an outer tim of chlorite which together are a few centimetres thick occur around the serpentinite inclusion, and carbonate occurs throughout the inclusion.

North Fischells Brook area

In the North Fischells Brook area (Fig. 7), layered ultramafic rocks (unit 01a) form a thin disrupted layer trending in a northwest-southeast direction. Figure 8 shows the nature of the layering which varies from 2 cm to less than 0.5 m. Surrounding this unit are inclusions of leucogabbro, layered gabbro and small zoned serpentinite inclusions (unit 01) all of which are pervasively veined by later granitoid (unit 01). Figure 9 shows a small block of the layered gabbro. The layering is on a centimetre scale and is interpreted as being initially a compositional igneous layering which has been transposed during regional deformation. Zoning of serpentinite inclusions is common in this area and is believed to represent metasomatic reaction zones between the ultramafic and country rocks (Curtis and Brown, 1969; Fowler et al., 1981; Sanford, 1982). Almost monomineralic rims of chlorite, tremolite-actinolite, and talc have been observed.

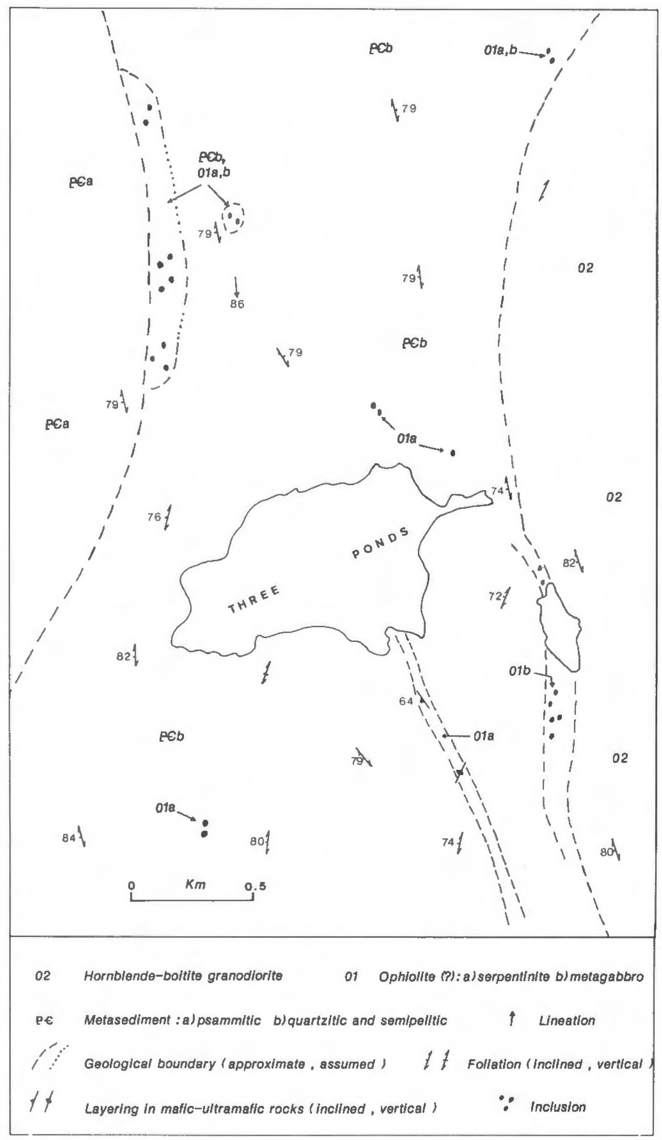

Figure 4. Detailed geological map of Three Ponds area.

Leucogabbro inclusions are composed mainly of plagioclase and hornblende with lesser amounts of quartz (Fig. 10). Also occurring in this area are distinct dark green inclusions veined by granitoid (Fig. 11). In places, these inclusions have the appearance of pillow lavas but on closer observation layering can be found in some of the better preserved samples. It is believed that these inclusions were once part of the layered ultramafic sequence which has been broken apart and partly assimilated by the granitoid. Mineralogically the inclusions consist of green amphibole and chlorite. Plagioclase is absent. These inclusions are similar to those found in the Three Ponds area.

To the west of the mafic-ultramafic belt, small ultramafic inclusions have been observed which are entirely enclosed within the quartzitic and semipeltic metasediments (unit PCb). Figure 12 shows a small inclusion composed mainly of talc in which the fabric of the metasediment wraps around the inclusion. Similar occurrences have been observed in other places throughout the metasedimentary unit (PCb) but are generally rare.

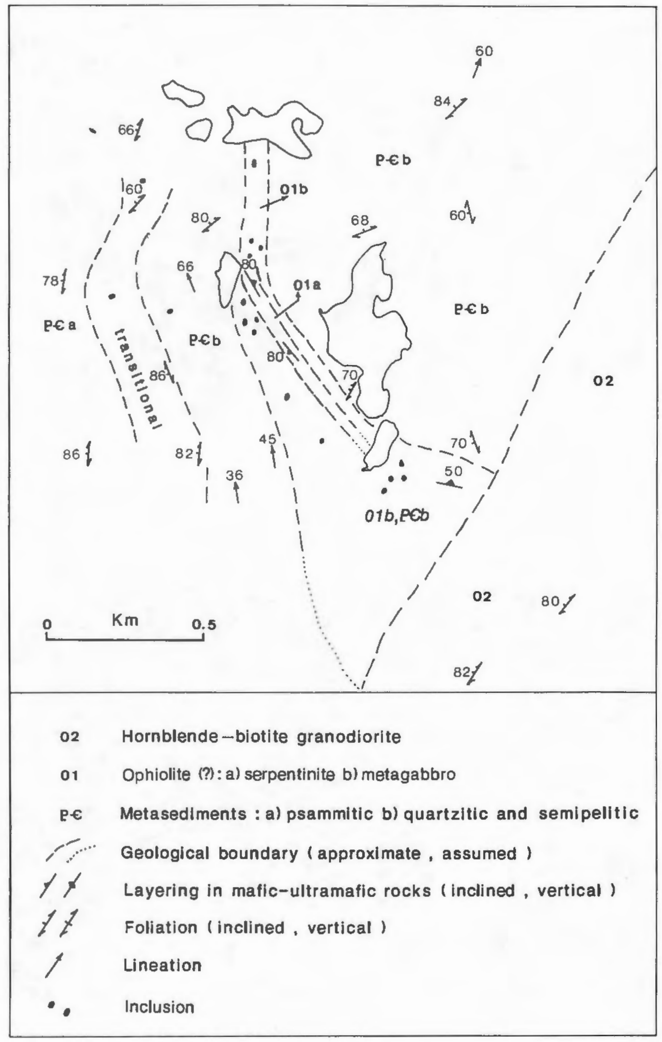

Figure 7. Detailed geological map of North Fischells Brook area.

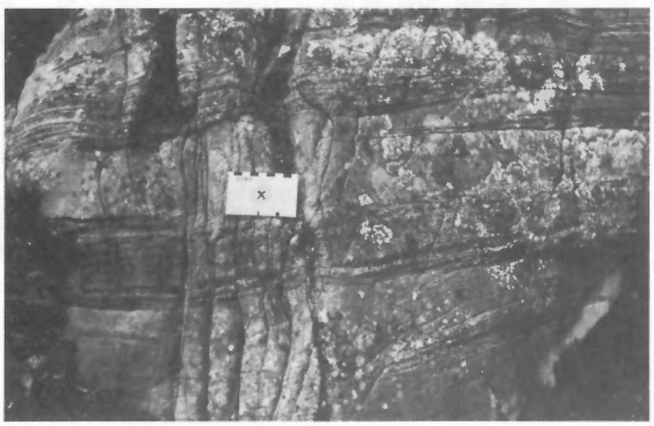

Figure 5. Layering developed in serpentinite (unit 01a) in Three Ponds area. Scale is 9 cm long. GSC-204460-K.

Figure 6. Large serpentinite inclusion (unit 01a) in metasediments (unit PCb). North of Three Ponds area. Hammer for scale. GSC-204460-F.

Figure 8. Layering developed in ultramafic rocks (unit 01a). Area north of Fischells Brook. Hammer for scale. GSC-204460-A.

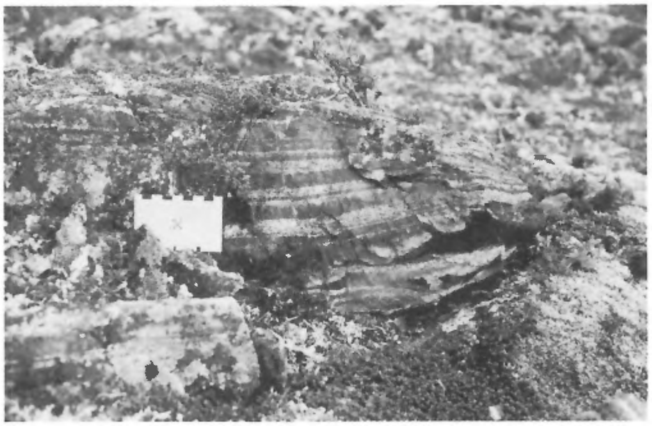

Figure 9. Small block of layered metagabbro (unit 01b). North of Fischells Brook. Scale is 9 cm. GSC-204460-C.

Figure 11. Distinct dark green inclusions (unit 01) composed of actinolite and veined by granitoid. Note angular to rounded nature. Scale is 9 cm long. GSC-204460-G.

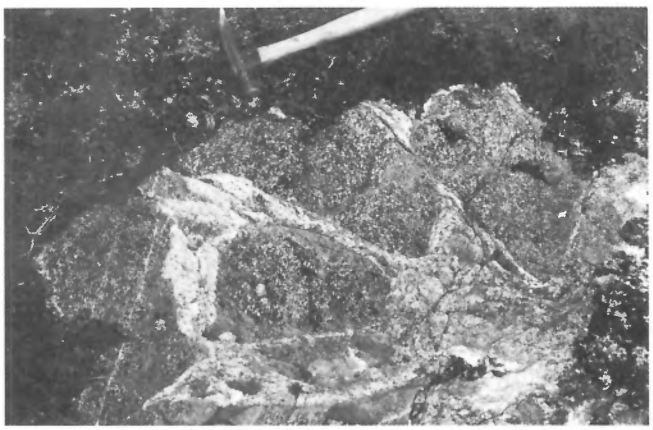

Figure 10. Gabbro-leucogabbro inclusions (unit 01b) veined by granitoids. Area north of Fischels Brook. Hammer for scale. GSC-204460-D.

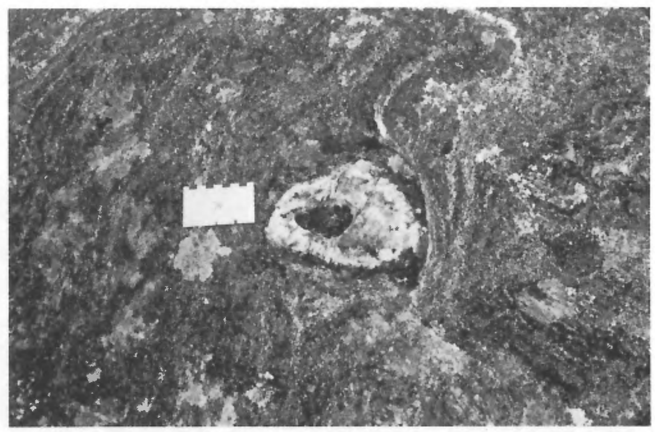

Figure 12. Ultramafic inclusion (unit 01a) in metasediment (unit PCb). Fabric in metasediment wraps around inclusion. Area north of Fischells Brook. Scale is 9 cm long. GSC-204460-J.

ECONOMIC GEOLOGY

An increase in gold exploration in recent years has resulted in a great deal of activity in ultramafic rocks in Newfoundland. As described by Buisson and LeBlanc (1985), carbonatized ultramafic rocks (listwaenites) may prove to be significant in terms of gold mineralization. The ultramafic rocks of this area have been subjected to extensive alteration and fluid movement as indicated by assemblages of talc-carbonate, chlorite and metasomatic reaction rims with the country rocks. Small scale, foliation-parallel quartz veins are locally abundant in the metasediments. Equally as important is the development of ductile shear zones in the metasediments. Archean lode gold deposits are commonly developed in such environments especially where mafic volcanic rocks and ultramafic komatiites are associated with sedimentary rocks (Roberts, 1987).

Analyses of several rock samples in this study gave a highest Au value of 40 ppb in a talc-carbonate rock containing pyrite. A chlorite-rich sample associated with ultramafics gave highest values of 36 ppb Pd and 54 ppb Pt. Only one sample of the metasediment was analyzed yielding 31 ppb Au. Further study of the gold potential in the map area is warranted.

Also associated with the mafic-ultramafic rocks is chromite, magnetite, Ni-Fe sulfides and asbestos. These occurrences however, are generally only small scale, local features and are believed to be of little economic value.

DISCUSSION

Van Berkel et al. (1986) interpreted the mafic-ultramafic rocks to be ophiolitic in nature and to mark ancient thrust surfaces along which ophiolitic rocks have been incorporated into older shelf-continental margin type sedimentary rocks. Similar occurrences are found in the northern part of Newfoundland where ultramafic bodies (serpentinites) occur in the Fleur de Lys Belt of the Baie Verte Peninsula. It is believed these ultramafic bodies represent remnants of oceanic crust which have been tectonically emplaced into rift facies clastics of the Fleur de Lys Supergroup (Bursnall, 1975; Hibbard, 1983). An alternative hypothesis was proposed by Chorlton (1983) for rocks in southwestern Newfoundland (south of the present study area). There, ophiolitic rocks of the Long Range Mafic-Ultramafic Complex are associated with semi-pelitic metasedimentary rocks. The stratigraphic interpretation however, is that the mafic-ultramafic complex is the oldest unit in the area and that the sedimentary rocks were deposited upon it.

Field relationships in the map area are difficult to ascertain due to the tectonically disrupted nature of the mafic-ultramafic rocks and the pervasive intrusion of granitoids. The mafic-ultramafic rocks are typically associated with high strain zones in metasediments. These zones are indicated by recrystallized ribbon quartz grains, small-scale tight to isoclinal folding and grain-size reduction which are interpreted as shear zone fabrics. Locally, shear bands have been observed which indicate a dextral sense of movement.

The latest disruption of the mafic-ultramafic rocks is interpreted to have resulted from extension and boudinage due to net strike-slip movement in the shear zones. Further disruption resulted from intrusion of the granitoids. Whether or not the shear zones had an earlier dip-slip history is difficult to ascertain. Steeply plunging down-dip mineral stretching lineations are locally developed but the importance of these is uncertain due to shearing and several stages of folding in the metasediments. Therefore, at this stage a thrusting model for incorporation of the mafic-ultramafic rocks into the metasediments is speculative.

Certain aspects of the field relationships in the map area may be described in terms of a mélange zone. Clearly, Figures 6 and 12 show the "block in matrix" style typically associated with such zones. However, other aspects of the area indicate the rocks are not totally chaotic but some continuity of layering is still maintained. Cowan's (1985) type 1 mélange zone is believed to have formed by progressive deformation of stratified sequences of sandstone and mudstone. Although the mechanism of deformation is controversial, progressive noncoaxial simple shear may play an important role in its formation. The rocks of the present study may best fit the type 1 mélange zone of Cowan (1985) with the exception that a previous history of incorporation of the mafic-ultramafic rocks into the metasediments has to be explained.

Regionnaly, the map area lies along strike of the Glover Island complex to the north (Knapp, 1980) and the Long Range mafic-ultramafic complex to the south (Chorlton, 1984). Both of these complexes are interpreted to be remnants of dismembered ophiolites. The Annieopsquotch ophiolite complex lies approximately 40 km to the east (Dunning, 1984; Dunning and Chorlton, 1985). The trace of the Baie Verte-Brompton Line as defined by Williams and St-Julien (1982) was shown to include the Glover Island complex, overstep eastward through the Annieopsquotch ophiolite complex and continue southwest through the Long Range mafic-ultramafic complex. Alternatively, the Baie Verte-Brompton Line may widen out south of Glover Island to include the metasediments and mafic-ultramafic rocks of the study area.

ACKNOWLEDGMENTS

The authors are indebted to the following persons for field mapping assistance and stimulating discussions: K.L. Currie and M.A.J. Piasecki in 1985, 1986 and 1987, T. van Nostrand and T. Rivers in 1986, and D. Machin in 1986 and 1987. The manuscript was reviewed by K.L. Currie and A. Nutman. Viking Helicopter of Pasadena supplied and moved the field parties efficiently. The Forest Research Station, Environment Canada in Pasadena supplied base camp space and logistical help. We thank Kruger Inc. and Abitibi-Price Ltd. for permission to use woods roads.

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Glacial dispersal of Deboullie syenite, northern Maine, into western New Brunswick[1]

Martin Rappol

Rappol, M., Glacial dispersal of Deboullie syenite, northern Maine, into western New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 49-53, 1988.

Abstract

Rare syenite and granodiorite cobbles of the Deboullie igneous complex in northern Maine are present in glacial depostis south of Grand Falls, New Brunswick, approximately 80-100 km due east of their source. Deboullie indicators seem to have been introduced to the study area during an older glacial event of eastward flow, that is recorded by the presence of a lower till in the Saint John River valley. They are found reworked in younger deposits associated with south or southeastward ice movement, but seem absent in the Anfield area where a nonerosional glaciolacustrine unit separates upper and lower till.

Résumé

On rencontre, dans des dépôts glaciaires situés au sud de Grand Falls au Nouveau-Brunswick, à environ 80-100 km à l'est de leur source, de rares galets syénitiques et granodioritiques appartenant au complexe igné de Deboullie situé dans le nord du Maine. Il semble que ces roches erratiques de Deboullie aient été amenées dans le secteur étudié, durant un épisode glaciaire ancien au cours duquel les glaces se sont écoulées vers l'est, et qui est indiqué par la présence d'un till inférieur dans la vallée de la rivière Saint-Jean. On trouve ces galets remaniés dans des dépôts plus récents, associés à un déplacement des glaces vers le sud ou le sud-est, mais il ne semble pas qu'ils existent dans la région d'Anfield, où se trouve une unité glaciolacustre non formée par l'érosion qui sépare le till supérieur du till inférieur.

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

INTRODUCTION

This summer, rare syenite boulders and cobbles were first observed in the Saint John River valley south of Grand Falls. On account of their frequency and angularity an Appalachian source for these rocks was suspected, because the majority of Laurentide Precambrian erratics is generally well rounded and polished. After consulting available bedrock maps and colleagues, it was inferred that the most likely source would be the Deboullie igneous complex in northern Maine. Comparison of syenite erratics from New Brunswick with outcrop samples of the Deboullie syenite indeed indicates a high degree of similarity between the two.

The Deboullie igneous complex (Fig. 1) has a diameter of roughly 3 km and consits of two major units: the northern half is mainly composed of alkali syenite, whereas granodiorite forms the southern part of the stock (Boone, 1962; Osberg et al., 1985). The syenite is quite distinctive, and as similar rocks are not known at present from igneous complexes of the Miramichi Highlands of central New Brunswick (J.B. Whalen and G. Watson, pers. comm.), this rock may be of value as an indicator of glacial transport directions and distances over a large area. Glacial dispersal in the immediate surroundings of the pluton has already been investigated by Halter (1986).

The purpose of this paper is to draw attention to the use of the Deboullie syenite as a potential indicator of ice movements in New Brunswick, and to report briefly on its occurrence and distribution in a small area around and south of Grand Falls.

Acknowledgments

I am most grateful to H. Russell and V.K. Prest for field assistance, and to V.K. Prest, W.W. Shilts, and C.R. van Staal for critical reading of the manuscript.

RESULTS

Two tills have been identified in the Saint John River valley south of Grand Falls (Rampton and Paradis, 1981: section 6 in their figure 11; Rappol, 1986b): a lower till with a clast fabric varying around the east-west direction, and an upper till with a north-south or north-northwest-south-southeast fabric. The older till is exposed at many places in the western river bank between the bridge at Limestone (21J/13: 599000, 5197750) and just south of the mouth of the Little River near Morrell (21J/13: 599700, 5188800).

Syenites are commonly found in washed slump and scree deposits on the river bank at places where the lower till is exposed. At these places, most of the erratics are sedimentary and volcanic rocks, whereas Precambrian igneous and metamorphic rocks are extremely rare or absent. However, where scree deposits from younger sediments cover the lower till, Precambrian erratics are relatively abundant (although generally still less than 2 % in boulder counts). Only two syenite cobbles were found in place in till along the Saint John River: one in the top of the lower till, the other in the upper till. It is assumed that the latter represents a reworked cobble.

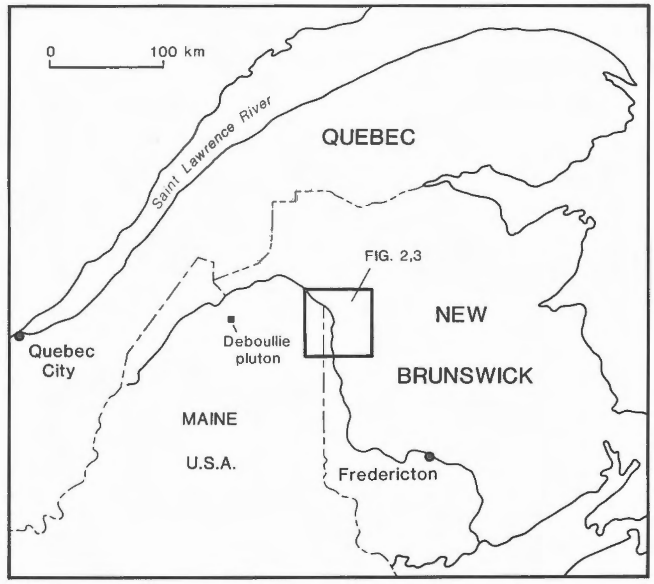

Figure 1. Location map.

Syenites, and also Deboullie granodiorites are abundant, however, in a till exposed along road 109, about 2 km east from the junction with road 105 at Perth-Andover (site at 21J/12: 601300, 5176800; see Fig. 2, 3). This till overlies bedrock with quite variable striae directions due to an irregular bedrock surface morphology. The main striation sets, together with stoss and lee relations, indicate ice movement towards 100° and 130°, and clast fabric in the till aligns with a direction of about 95°-275°. The till is overlain by slope deposits.

The outcrop is in a narrow east-west oriented valley, that is bounded by steep and over 150 m high bedrock slopes. This setting, that may have protected the till from being reworked by younger ice movement in a southern direction, together with striae and fabric orientations suggest that this till is a correlative of the lower till in the Saint John River valley.

Further observations were made at sites indicated in Figure 3. Natural exposures, gravel pits, and boulder piles in fields were visited, and subjected to 15 — 30 minute searches for Deboullie syenites. There are several points of interest in the distribution as shown in Figure 3.

Firstly, Deboullie indicators do not occur north of Grand Falls, including a possible older till occurrence (east-west fabric) in the Little River valley (210/4: 596150, 5212800), just northeast of the town. Also, during three summers of more casual observations during regional investigations, no similar syenites were ever noted north of Grand Falls. In a way this is a fortunate circumstance, as it prevented redistribution of these rocks by the Late Wisconsinan northwestward flow event, that finds its most southern manifestation in bedrock striations at Bellefleur, 13 km northwest of Grand Falls (Rappol, 1986a; see Fig. 2). It may be assumed therefore, that the northern limit of the occurrence of Deboullie indicators marks the northern limit of the original dispersal train.

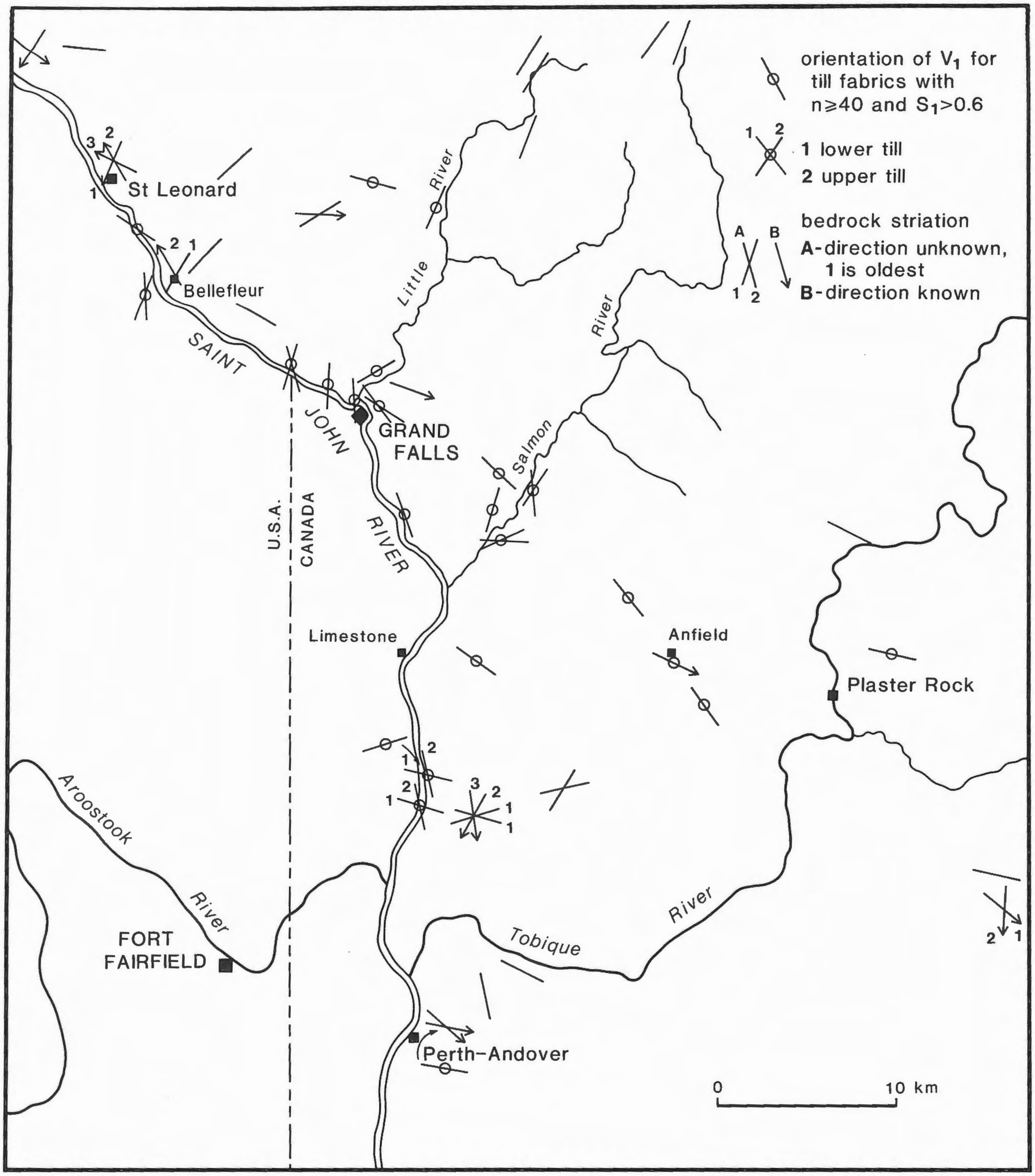

Figure 2. Available striation and till fabric data. Some striations after Thibault (1980).

Secondly, Deboullie indicators appear to be very scarce in the Anfield area, but reappear again farther east in the Plaster Rock area. This supports the assumption that in the investigated area the Deboullie syenites were introduced during the lower-till glacial event, for the following reason. The younger surface till at Anfield was deposited by ice moving towards the southeast (Fig. 2), and overlies deltaic and glaciolacustrine deposits (Rampton et al., 1984). These in turn overly a lower till. The existence of this older till was suspected by Thibault (1980) on the basis of hammer seismics, and was recently confirmed by boreholes at Anfield (21J/13: 612700, 5198200) and Blue Bell (21J/13: 610750, 5201450). Apparently, the non-erosional glaciolacustrine phase burying the lower till, prevented Deboullie indicators from being reworked into the younger deposits.

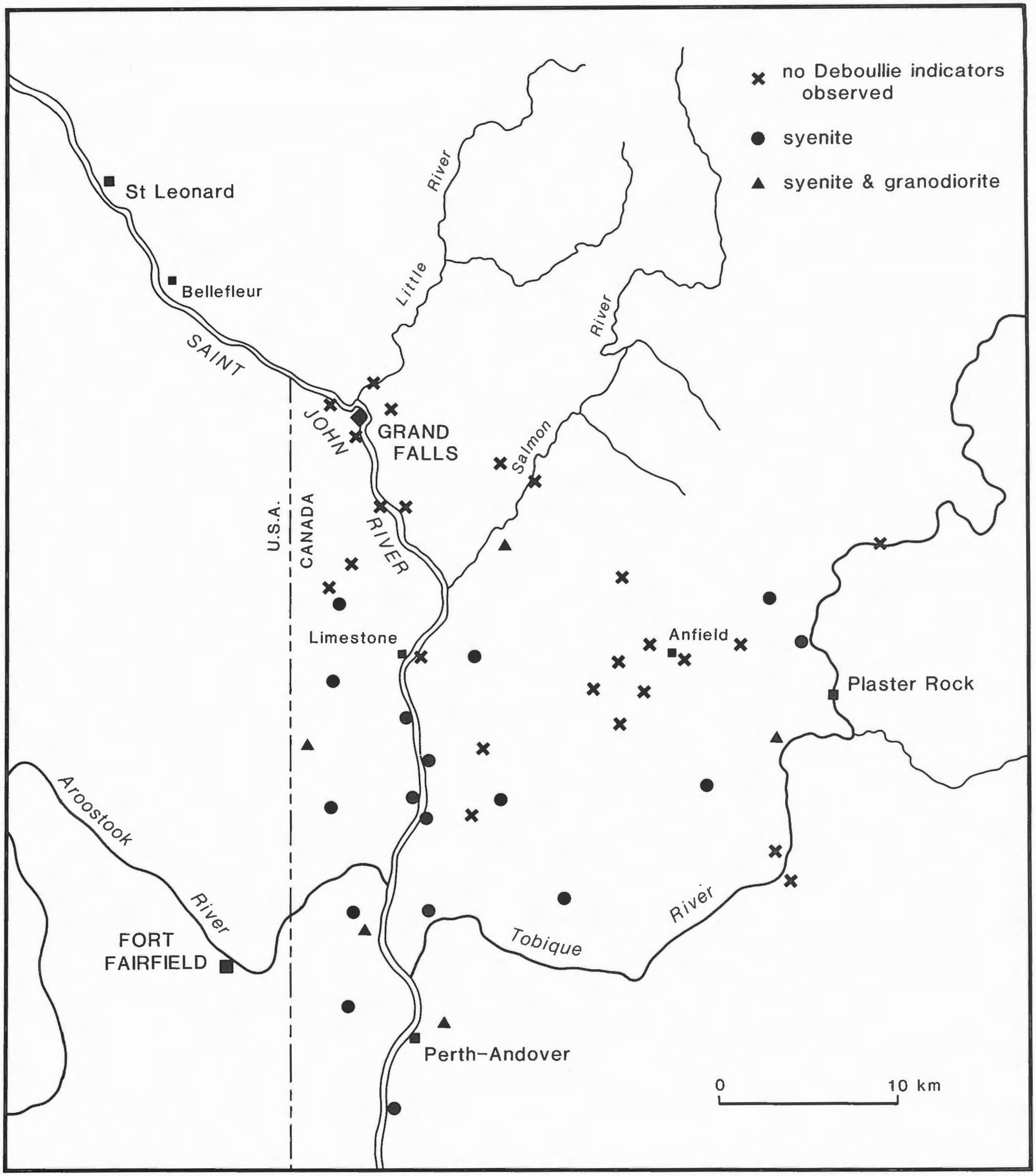

Figure 3. Occurrence of Deboullie indicators in the study area.

Because of the limited outcrop area of the older till, most observations in Figure 3 concern material derived from the surface till and its associated glaciofluvial deposits. As discussed above, it is presently assumed that Deboullie indicators in these deposits have been reworked from older deposits during the younger southward and southeastward flow event. The southern limit of the original dispersal train can therefore not be determined. Moreover, dispersal in a small area around the Deboullie pluton itself as studied by Halter (1986) in conjunction with results from regional investigations by Lowell and Kite (1986), indicates Late Wisconsinan dispersal in a southeastern to east-southeastern direction in the surface till. At present, it is not sure however, whether ice flow lines over the Deboullie pluton during this glacial event reached as far into New Brunswick as suggested by Lowell and Kite (1986: figure 5), or perhaps swung southwards towards the Gulf of Maine after collision with south-southwest moving ice from New Brunswick. Deboullie syenite observed in the Knoxville esker, about 25 km south of Perth-Andover (V.K. Prest, pers. comm.) may therefore have arrived there via two alternative flow paths.

It is further of interest that the occurrence of Deboullie syenite is always accompanied by a high frequency of what may be described as a chlorite-mottled volcanic wacke (F.W. Chandler, pers. comm.), probably also derived from northern Maine. Because of its much higher frequency (at least up to 6 % in boulder counts of the 10-30 cm fraction), and its distinctive character, this rock could be likewise a valuable indicator, provided that its source area is limited and unique, and also located.

CONCLUSION

One implication of the data and interpretations from the middle Saint John River valley is discussed here. Precambrian erratics are extremely rare in the lower till and inderlying gravel, whereas these are easily found in the overlying deposits. A similar relation is found in southeastern Quebec between the Chaudiere and Lennoxville tills, where the lower part of the Chaudiere Till was possibly deposited by an Appalachian ice mass (Shilts, 1981). If the lower till of the Saint John River valley and the Chaudiere Till are correlatives, and indeed deposited by an Appalachian ice mass, then the occurrence of Deboullie indicators in the lower till in New Brunswick suggest that the position of the early Chaudiere ice divide should be located close to the boundary between Maine and Beauce county, Quebec.

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Field relations, petrology, and age of the northeastern Point Wolfe River pluton and associated metavolcanic and metasedimentary rocks, eastern Caledonian Highlands, New Brunswick[1]

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Barr, S.M. and White, C.E., Field relations, petrology, and age of the northeastern Point Wolfe River pluton and associated metavolcanic and metasedimentary rocks, eastern Caledonian Highlands, New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 55-67, 1988.

Abstract

On the basis of field mapping and petrological studies, the northeastern Point Wolfe River pluton is subdivided into six units (in inferred intrusive sequence): quartz diorite-tonalite, quartz monzodiorite-tonalite, Pollett River granodiorite, porphyritic quartz granodiorite, Blueberry Hill granite, and granite porphyry. These probably form a co-magmatic, differentiated "I-type" suite, generated in a volcanic arc environment during the late Precambrian. They intruded a sequence of low grade metavolcanic and metasedimentary rocks, probably also of late Precambrian age. However, the metavolcanic rocks are mainly volcanic-arc tholeiites, generated in a relatively primitive subduction environment, and do not appear to be co-magmatic with the granitoid rocks.

Résumé

En fonction des résultats de travaux de terrain in situ et d'études pétrologiques, on a subdivisé le pluton du nord-est de Point Wolfe River en six unités (selon la séquence intrusive présumée): diorite quartzique et tonalite, monzodiorite quartzique et tonalite, granodiorite de Pollett River, granodiorite quartzique porphyrique, granite de Blueberry Hill et porphyre granitique. Ces unités forment probablement une suite comagmatique, différenciée, « de type I », produite dans un milieu d'arc volcanique durant le Précambrien supérieur. Elles sont intrusives dans une séquence de roches métavolcaniques et métasédimentaires de degré métamorphique faible, probablement aussi d'âge précambrien supérieur. Toutefois, les roches métavolcaniques sont principalement des tholéiites d'arc insulaire, produites dans un milieu de subduction relativement primitif, et ne semblent pas être comagmatiques des roches granitoïdes.

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[1] Contribution to the Canada - New Brunswick Mineral Development Agreement 1984-1989. DSS Contract 20ST. 23233-6-0040.

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INTRODUCTION

The Point Wolfe River pluton, named by Ruitenberg et al. (1979), is the largest granitoid intrusion in the eastern Caledonian Highlands of southern New Brunswick. It intruded metavolcanic and metasedimentary rocks assigned to the Coldbrook Group of probable late Precambrian age (Kindle, 1962; Ruitenberg et al. 1977, 1979; Giles and Ruitenberg, 1977). The Caledonian Highlands are generally considered to form a typical part of the Avalon terrane of the northern Appalachians (e.g. O'Brien et al., 1983).

During the summer of 1986, the northeastern part of the Point Wolfe River pluton was mapped and sampled as part of a continuing study of granitoid and associated metavolcanic and metasedimentary rocks of the eastern Caledonian Highlands. The map area adjoins to the north and west that mapped in 1985 and described by Barr (1987), and includes approximately 300 km^2 (Fig. 1). In addition, some mapping was completed on the southern margin of the 1985 map area. A simplified version of the map and cross-section resulting from this work are presented here (Fig. 2).

A total of 565 samples were collected in the 1986 map area. On the basis of petrographic studies, 28 samples of granitoid rocks and 27 samples of metavolcanic rocks were selected for major and trace element analyses. Six samples from the main unit of the Point Wolfe River pluton were chosen for Rb-Sr age dating.

This report summarizes the results and conclusions from these studies.

POINT WOLFE RIVER PLUTON

The northeastern part of the Point Wolfe River pluton is divided into 6 separate lithologies (Fig. 2). An intrusive sequence from more mafic to more felsic is inferred, although contact relations were rarely observed. None of these lithologies is identical to rocks of the Fortyfive River, Alma, and Goose Creek intrusions described by Barr (1987).

In many exposures, the granitoid rocks exhibit strong shearing and locally protomylonite is developed. Where shearing is less intense, the granitoid rocks lack foliation.

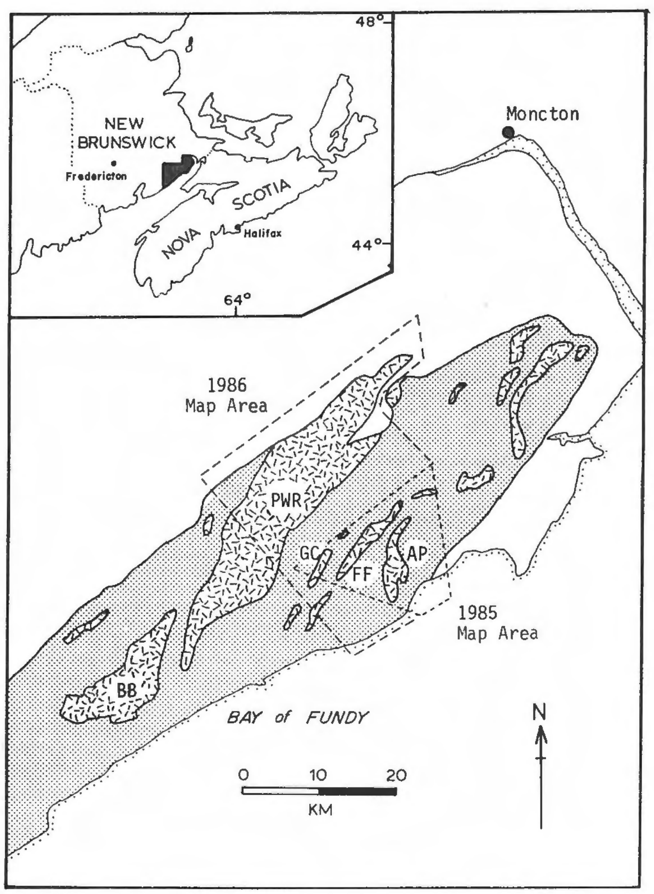

Figure 1. Location map for the eastern Caledonian Highlands. Stippled pattern indicates metavolcanic and metasedimentary rocks; dash pattern indicates granitoid rocks. Dashed lines outline 1985 and 1986 map areas. PWR = Point Wolfe River pluton BB = Bonnell Brook pluton; AP = Alma pluton; FF = Fortyfive River pluton; GC = Goose Creek leucotonalite.

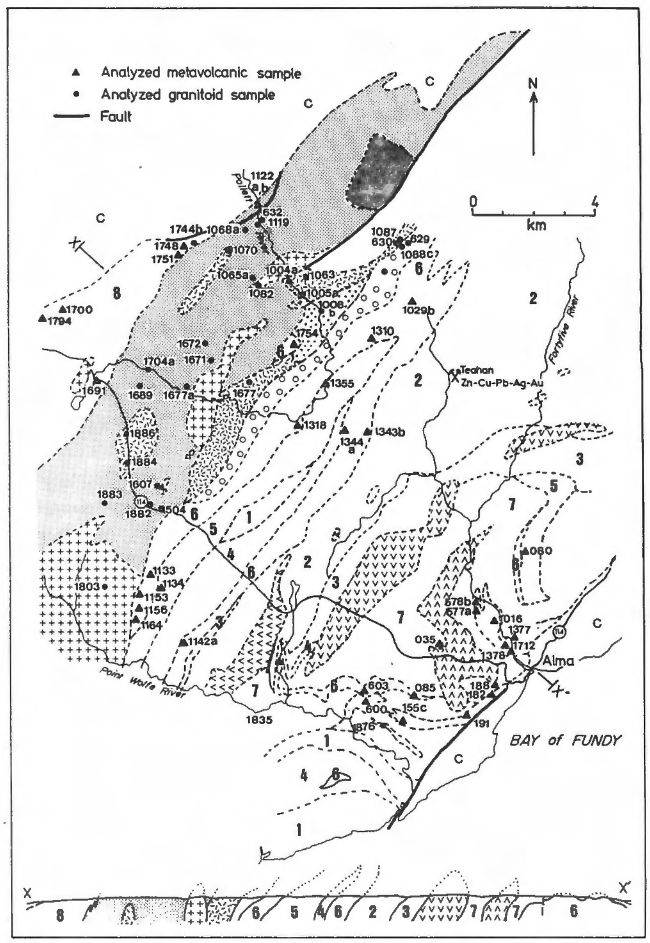

Figure 2. Simplified geological map and cross-section of the combined 1985 and 1986 map areas, showing the distribution of map units described in the text and the locations of analyzed samples. Sample numbers below 100 are volcanic samples from Barr (1987).

Hence they are interpreted to have been emplaced after the D_1 deformational event in the host rocks. Most samples show moderate to intense alteration, including replacement of plagioclase by saussurite, sericite, and albite, and mafic minerals by chlorite, sphene, epidote, and other secondary minerals.

Small bodies of quartz diorite-tonalite are a minor component of the pluton in the map area. They are characterized by 15 to 30 % mafic minerals (amphibole or amphibole and biotite) and only minor amounts of alkali feldspar. Modal compositions range from quartz diorite to tonalite and, marginally, granodiorite (Fig. 3). Texture ranges from subporphyritic (with prominent subhedral plagioclase) to equigranular, and grain size from fine to medium. Nine analyzed samples from locations indicated in Figure 2 have silica contents ranging from 52 to 62 % (Table 1), with most samples

LEGEND (for Figure 2)

CARBONIFEROUS

C Sandstone, conglomerate

LATE PRECAMBRIAN (stratigraphic sequence uncertain)

8 Metavolcanic rocks (mainly metatuffs and mafic to intermediate flows)

7 Crystal metatuff, pyritiferous grey felsite

6 Mainly massive to amygdaloidal metabasalt; minor mafic tuffaceous schists and phyllites

5 Metarhyolite

4 Red, maroon, and grey slate, phyllite, arkosic metasiltstone and metasandstone, metaconglomerate; minor grey to green quartzite

3 Massive meta-arkosic siltstone, sandstone, and conglomerate

2 Tuffaceous phyllite, chloritic schist; minor slate felsite, and meta-arkose

1 Metavolcanic rocks (mainly mafic phyllites, metatuffs, and flows)

LATE PRECAMBRIAN Granitoid Rocks

Point Wolfe River pluton

Granite porphyry

Blueberry Hill granite

Quartz porphyritic granodiorite

Pollett River granodiorite

Quartz monzodiorite/tonalite

Quartz diorite/tonalite

Fortyfive River granodiorite

Goose Creek leucotonalite

Alma diorite

containing about 60 % SiO_2. Hence overall these rocks are less mafic than those of the Alma pluton (Barr, 1987). Some chemical trends are apparent, with TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, and P_2O_5 showing reasonable negative correlations with SiO_2 (Fig. 4). MnO, K_2O, and Na_2O do not show any significant trends, and the latter two display considerable scatter. Among the trace elements (Fig. 5a, b, c), only Sr, Cu, V, and Ga show reasonable negative correlation with SiO_2 and Pb a positive correlation. This variability may be partly a result of alteration, but such scattered data appear to be a feature of dioritic units (Barr, 1987; Barr et al. 1982).

Quartz monzodiorite-tonalite occurs only in one area near the northeastern end of the pluton. The rock is coarse grained, and contains about 20 % prominent subhedral amphibole. Biotite is much less abundant and finer grained than amphibole. The most abundant mineral is plagioclase; interstitial quartz and microcline make up about 15 % and less than 10 %, respectively. On the basis of modal compositions, the unit ranges from tonalite to quartz diorite and quartz monzodiorite, with analyzed sample 1008B transitional between quartz monzodiorite and granodiorite (Fig. 3). Sample 1008B, with about 60 % SiO_2 (Table 1), is chemically similar to samples from the dioritic unit with similar SiO_2 content (Fig. 4, 5).

The Pollett River granodiorite is the main unit of the Point Wolfe River pluton within the map area. It consists of medium grained granodiorite (Fig. 3), containing subporphyritic subhedral plagioclase, amphibole, biotite, interstitial quartz and microcline, and accessory magnetite, apatite, allanite, and sphene. In the area notheast of Pollett River, the granodiorite is mixed with abundant mafic material, apparently mafic metavolcanic rocks, and both are highly altered and sheared. Elsewhere, such xenolithic material is less abundant. Also common in the granodiorite are mafic dykes, typically 1 or 2 m in width. Thirteen analyzed samples from the Pollett River granodiorite show a range in SiO_2 content

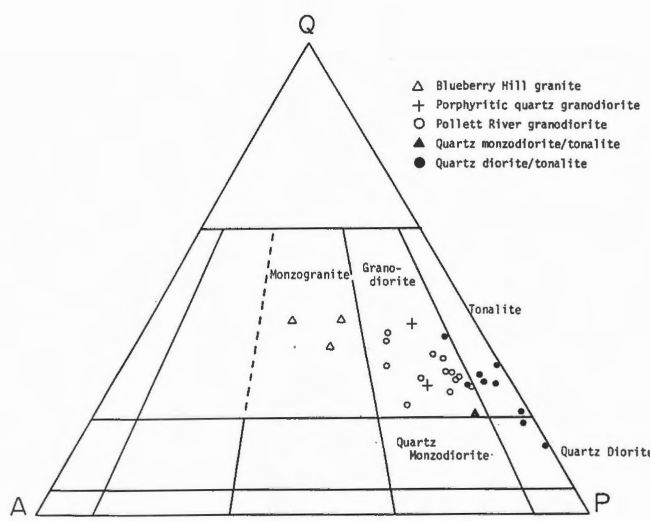

Figure 3. Modal compositions of analyzed granitoid samples plotted on the quartz-alkali feldspar-plagioclase ternary diagram with fields from Streckeisen (1976). Modal analyses were obatined by point counting under a binocular microscope of at least 500 points on slabs stained for potassium feldspar.

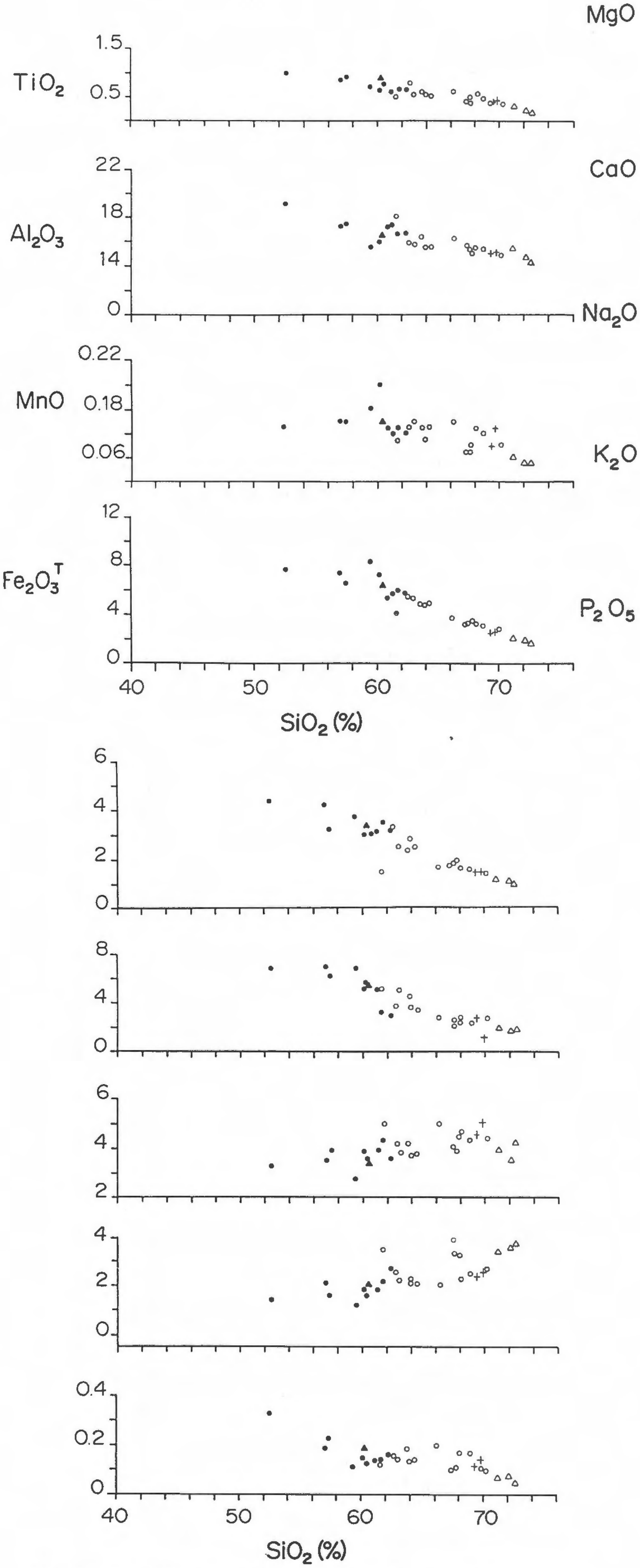

Figure 4. Silica variation diagrams for major element oxides in samples from the northeastern Point Wolfe River pluton. Symbols as in Figure 3.

from about 61 to 70 % (Table 2), which is consistent with the range in proportions of modal mafic minerals, quartz, and feldpsar (Fig. 3). Major elements generally display good negative correlations with SiO_2, especially when viewed in combination with the diorite-tonalite samples (Fig. 4). However, Na_2O and K_2O show considerable scatter and no trends are apparent. Among the trace elements (Fig. 5a, b, c), good negative trends are developed in Cu, Zn, V, Ga, and Sr and positive trends in Ba and Pb. Other elements show less consistency, although overall the Pollett River granodiorite has lower abundances of Ni and Cr and higher Rb, Zr, and Th than the diorite-tonalite samples.

A large body of porphyritic quartz granodiorite forms the southwestern part of the Point Wolfe River pluton, as well as two small intrusions within the Pollett River granodiorite to the northeast (Fig. 2). It is characterized by large (up to 2 cm) ovoid phenocrysts of quartz, which survive as

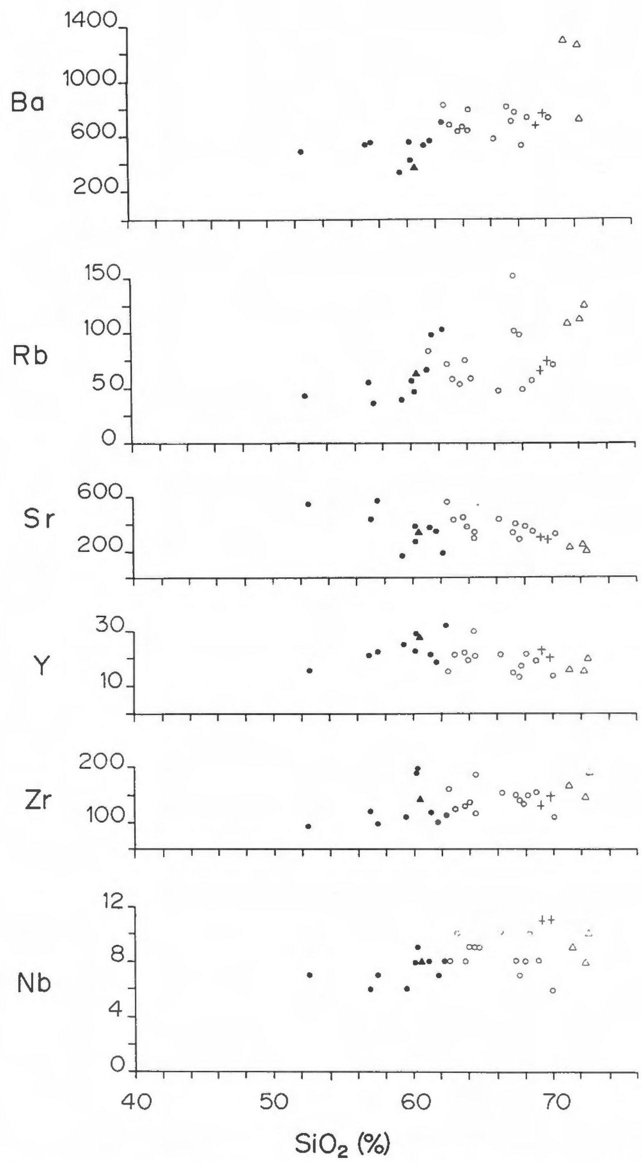

Figure 5a. Silica variation diagrams for Ba, Rb, Sr, Y, Zr, and Nb. Symbols as in Figure 3.

augen in the more highly sheared samples. Plagioclase is subhedral and subporphyritic, and microperthitic microcline is interstitial. Biotite is the main mafic mineral, with minor amphibole in some samples. Allanite is an abundant accessory mineral. The modal composition, as exemplified by the two analyzed samples, is similar to that of the Pollett River granodiorite (Fig. 3). Silica content is relatively high in these samples (Table 2), and they are chemically similar to the Pollett River granodiorite samples with highest SiO_2 (Fig. 4, 5), with the exception of somewhat higher Nb.

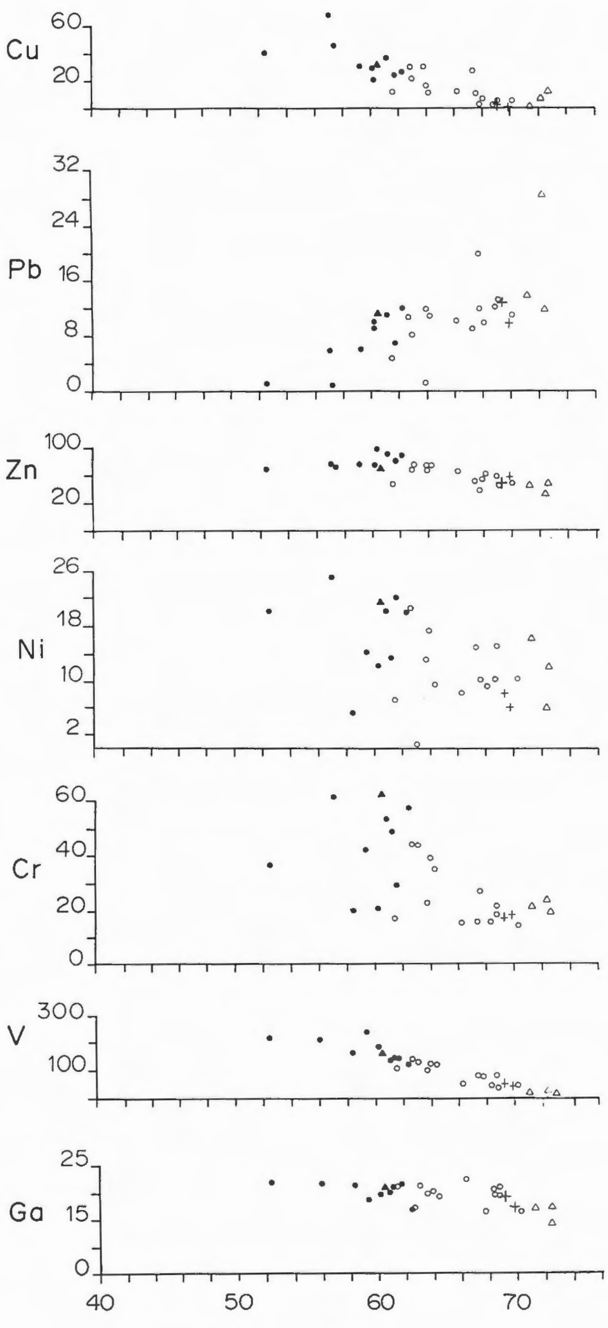

Figure 5b. Silica variation diagrams for Cu, Pb, Zn, Ni, Cr, V, and Ga. Symbols as in Figure 3.

The Blueberry Hill granite extends along the southern margin of the Point Wolfe River pluton from Highway 114 to the northeastern end of the pluton. In many areas it is intensively sheared and reduced to a fine grained protomylonite, but locally the original texture is partially preserved. In those places the rock is coarse grained and consists of approximately equal amounts of quartz, plagioclase, and alkali feldspar with less than 10 % mafic minerals. These minerals form augen in a more granulated matrix, but the original texture appears to have been equigranular. Three samples from the Blueberry Hill granite contain 71-72 % silica, higher than in the granodiorites previously described, but the samples generally plot on the extension of the diorite - granodiorite trend (Fig. 4, 5). Two of the samples are anomalously high in Ba and all three are high in Rb and Th.

A large body of granite porphyry occurs within the Pollett River granodiorite in the northeastern part of the map area (Fig. 2). Most exposures of this unit are intensely deformed and altered, and mafic (metavolcanic?) xenoliths are abundant. The porphyry consists of euhedral plagioclase phenocrysts (up to 0.5 cm in length), less abundant ovoid quartz phenocrysts, and very rare mafic phenocrysts in a fine grained equigranular groundmass of anhedral quartz and alkali feldspar. The granite porphyry appears to have intruded the granodiorite and is probably the youngest (and highest level) unit of the pluton. No samples suitable for chemical analysis were obtained.

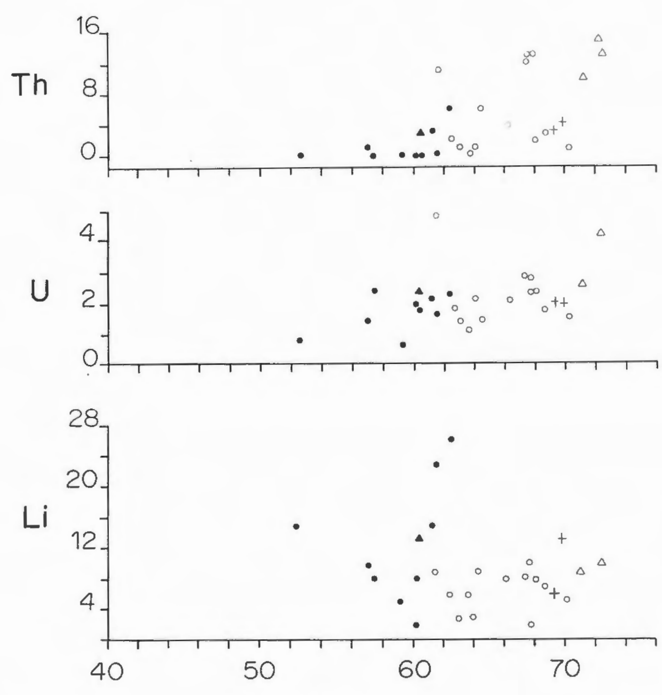

Figure 5c. Silica variation diagrams for Th, U, and Li. Symbols as in Figure 3.

Table 1. Chemical analyses* and CIPW normative mineralogy for samples from the quartz diorite/tonalite and quartz monzodiorite of the Point Wolfe River pluton

	Diorite/Tonalite									Quartz Monzodiorite
%	630	1087	1063	1004A	1005A	1070	1667	1884	1886	1008A
SiO_2	62.36	60.85	61.68	61.23	60.33	56.92	59.44	52.45	57.40	60.42
TiO_2	0.67	0.63	0.67	0.62	0.71	0.80	0.71	0.99	0.89	0.74
Al_2O_3	16.72	17.25	16.55	16.72	16.00	16.63	15.57	19.17	17.53	16.49
$Fe_2O_3^T$	5.86	5.44	6.02	5.74	7.23	7.29	8.14	7.72	6.78	6.34
MnO	0.10	0.11	0.11	0.10	0.18	0.12	0.14	0.11	0.12	0.12
MgO	3.27	3.03	3.47	3.11	3.03	4.17	3.74	4.39	3.23	3.44
CaO	2.48	5.19	3.34	5.25	5.82	6.54	7.07	6.93	6.28	5.35
Na_2O	3.66	3.90	4.37	3.95	3.57	3.46	2.70	3.30	3.90	3.37
K_2O	2.69	1.96	2.24	1.86	1.66	2.17	1.24	1.44	1.57	2.05
P_2O_5	0.16	0.15	0.14	0.14	0.13	0.19	0.12	0.33	0.26	0.18
LOI	2.50	1.40	1.70	1.20	0.80	1.80	1.20	3.00	1.60	1.65
Sum	100.47	99.91	100.29	99.92	99.46	100.09	100.07	99.83	99.56	100.15
Q	18.47	12.99	12.80	13.40	14.09	6.79	16.05	3.48	8.65	14.16
C	3.79	0.00	1.20	0.00	0.00	0.00	0.00	0.41	0.00	0.00
Or	16.27	11.83	13.46	11.14	9.96	13.15	7.48	8.80	9.54	12.33
Ab	31.70	33.67	37.72	34.04	30.75	29.90	23.28	28.97	33.90	29.13
An	11.56	24.26	16.04	22.84	23.28	24.03	27.19	33.53	26.36	24.36
Di	0.00	0.46	0.00	1.31	2.21	3.75	3.31	0.00	1.78	0.74
He	0.00	0.34	0.00	1.01	2.24	2.73	3.08	0.00	1.50	0.56
En	8.33	7.49	8.79	7.25	6.68	8.87	7.93	11.35	7.44	8.39
Fs	7.07	6.38	7.23	6.40	7.76	7.37	8.48	9.24	7.20	7.35
Mt	1.12	1.04	1.15	1.09	1.38	1.40	1.54	1.49	1.30	1.22
Il	1.31	1.20	1.30	1.20	1.38	1.56	1.37	1.94	1.74	1.38
Ap	0.38	0.36	0.31	0.31	0.28	0.45	0.28	0.80	0.60	0.40
ppm										
Ba	696	565	577	538	418	530	338	492	564	559
Rb	104	57	94	64	47	58	37	43	39	63
Sr	187	385	324	377	227	435	170	540	562	332
Y	32	23	19	21	29	21	25	16	22	27
Zr	111	186	98	120	192	118	108	81	95	142
Nb	8	8	7	8	9	6	6	7	7	8
Cu	25	28	23	37	20	67	30	39	45	29
Pb	12	10	7	11	9	6	6	1	1	11
Zn	89	75	79	70	93	77	75	69	75	73
Ni	20	20	22	13	12	25	14	20	5	21
Cr	57	53	29	44	21	61	42	36	20	62
V	123	137	145	143	181	207	238	219	163	161
Ga	17	20	21	23	19	22	18	22	21	20
Th	6	0	0	3	0	1	0	0	0	3
U	2.3	1.9	1.6	2.1	1.7	1.4	0.6	0.8	2.4	2.3
Li	26	2	23	15	8	10	5	15	8	13
Sn	-	1.5	3.1	2.1	2.2	1.3	2.6	4.9	5.4	2.5
Au**	-	<1	<1	<1	<1	<1	2	2	<1	<1

*Analytical methods as in Table 2.
**in ppb

METAVOLCANIC AND METASEDIMENTARY ROCKS

Lithology and field relations

The metavolcanic and metasedimentary rocks in the map area are divided into eight units (Fig. 2). It is expected that these units will be modified as mapping in the region continues, and hence should be regarded as preliminary. Locally, a stratigraphic sequence can be constructed, but it is not yet clear whether or not this can be applied throughout the map area.

The rocks contain mineral assemblages indicative of lower greenschist facies metamorphism. Mafic metavolcanic rocks are dominated by albite, epidote, and chlorite, and more felsic lithologies are characterized by abundant quartz, albite, and sericite. They also are pervasively deformed, with fine grained metasedimentary and metatuffaceous units displaying a strong phyllitic foliation (S_1), and the more massive and/or competent metasedimentary and metavolcanic flow units a less intensely penetrative S_1 foliation. Primary layering and bedding (S_0), are generally subparallel to the

Table 2. Chemical analyses* and CIPW normative mineralogy for samples from the Pollett River granodiorite, quartz porphyritic granodiorite, and Blueberry Hill granite of the Point Wolfe River pluton

	Pollett River Granodiorite													Porphyritic Quartz Granodiorite		Blueberry Hill Granite		
%	504	632	1065A	1068A	1119	1607	1671	1672	1677A	1689	1704A	1882	1883	1082	1803	629	1032A	1088C
SiO_2	64.34	67.62	68.81	61.56	67.42	63.02	68.13	66.29	63.70	62.58	67.77	63.94	70.27	69.86	69.36	72.17	72.49	71.23
TiO_2	0.54	0.43	0.51	0.49	0.43	0.56	0.54	0.59	0.63	0.70	0.47	0.53	0.37	0.44	0.36	0.26	0.23	0.29
Al_2O_3	15.67	15.43	15.40	18.08	15.76	15.97	15.50	16.26	16.20	15.95	15.22	15.52	14.94	15.12	15.26	14.70	14.43	14.76
$Fe_2O_3^T$	4.97	3.35	3.13	4.15	3.34	5.31	3.36	3.83	4.92	5.54	3.57	4.95	2.87	2.86	2.88	2.00	1.88	2.21
MnO	0.11	0.07	0.10	0.09	0.07	0.12	0.11	0.12	0.11	0.10	0.08	0.09	0.08	0.11	0.09	0.05	0.05	0.06
MgO	2.52	1.80	1.53	1.49	1.72	3.03	1.52	1.62	2.34	3.22	1.86	2.75	1.39	1.35	1.40	0.96	0.84	1.11
CaO	3.35	2.63	2.40	4.64	2.32	4.57	2.72	2.98	4.31	3.82	2.35	3.63	2.41	1.53	2.42	1.93	1.86	1.98
Na_2O	3.82	3.97	4.37	5.06	4.14	3.85	4.70	5.06	4.15	4.17	4.49	3.72	4.41	5.01	4.49	3.64	4.22	3.95
K_2O	2.06	3.39	2.46	3.53	3.93	2.25	2.34	2.08	2.10	2.65	3.29	2.24	2.72	2.65	2.42	3.51	3.75	3.42
P_2O_5	0.14	0.11	0.17	0.12	0.11	0.14	0.17	0.20	0.18	0.16	0.11	0.14	0.10	0.13	0.11	0.07	0.06	0.07
LOI	2.20	1.40	0.90	1.20	1.20	1.20	1.20	1.10	1.10	1.60	0.90	2.80	1.00	0.80	1.00	0.70	0.50	1.00
Sum	99.72	100.20	99.78	100.41	100.44	100.02	100.29	100.13	99.74	100.49	100.11	100.31	100.51	99.86	99.79	99.91	100.31	100.08
Q	21.69	22.17	25.80	5.93	19.58	15.94	22.82	18.96	17.22	13.20	19.91	20.27	26.10	24.49	25.83	31.39	27.70	28.63
C	1.45	0.73	1.59	0.00	0.75	0.00	0.72	0.76	0.00	0.00	0.26	0.74	0.64	1.55	1.10	1.57	0.22	1.12
Or	12.56	20.37	14.71	21.13	23.51	13.54	14.02	12.41	12.59	15.94	19.69	13.60	16.16	15.87	14.53	20.96	22.27	20.47
Ab	33.22	34.04	37.52	43.32	35.34	33.05	40.19	43.40	35.77	35.79	38.43	32.36	37.53	42.84	38.57	31.09	35.77	33.82
An	16.20	12.54	11.02	16.38	10.93	20.03	12.56	13.66	19.76	17.28	11.08	17.64	11.34	6.89	11.53	9.22	8.82	9.49
Di	0.00	0.00	0.00	2.36	0.00	1.00	0.00	0.00	0.40	0.42	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
He	0.00	0.00	0.00	2.71	0.00	0.74	0.00	0.00	0.34	0.30	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
En	6.44	4.53	3.87	2.66	4.31	7.21	3.83	4.07	5.75	7.93	4.66	7.06	3.36	3.40	3.52	2.41	2.08	2.77
Fs	6.13	3.94	3.57	3.51	3.92	6.11	3.86	4.46	5.64	6.38	4.16	6.03	3.41	3.31	3.47	2.38	2.23	2.61
Mt	0.96	0.63	0.57	0.78	0.63	1.00	0.62	0.71	0.93	1.05	0.68	0.96	0.54	0.54	0.55	0.37	0.34	0.41
Il	1.04	0.81	0.98	0.94	0.81	1.06	1.02	1.12	1.20	1.33	0.90	1.04	0.69	0.83	0.68	0.48	0.44	0.54
Ap	0.31	0.24	0.38	0.28	0.24	0.31	0.38	0.47	0.40	0.38	0.24	0.31	0.23	0.28	0.24	0.14	0.14	0.14
ppm																		
Ba	634	709	730	805	801	675	710	596	625	781	762	655	712	752	668	1056	704	1071
Rb	57	102	57	82	152	57	49	48	54	71	99	75	70	72	65	112	125	111
Sr	292	295	347	555	317	404	381	412	427	316	289	367	300	280	278	212	181	222
Y	21	13	19	15	15	21	22	21	22	29	17	19	14	20	23	15	20	16
Zr	117	142	152	160	149	124	148	150	127	180	143	131	109	146	130	146	133	167
Nb	9	7	8	8	8	10	10	10	8	9	8	9	6	11	11	8	10	9
Cu	11	11	0	13	26	21	6	13	29	27	3	15	5	0	2	6	13	0
Pb	11	20	12	5	9	8	10	10	1	11	12	12	11	10	13	29	12	14
Zn	74	37	59	42	51	74	59	65	65	70	48	65	47	49	42	36	35	33
Ni	9	10	10	7	15	0	9	8	15	19	15	17	10	6	8	6	12	15
Cr	35	27	19	18	16	44	16	16	23	44	22	39	15	17	18	24	20	21
V	112	70	38	105	74	132	40	47	101	128	76	123	50	36	46	20	18	25
Ga	18	16	18	22	18	21	20	22	19	17	21	20	16	17	20	14	17	17
Th	5	13	3	11	12	1	2	4	0	2	13	1	1	4	3	15	13	10
U	1.4	2.7	1.7	4.9	2.8	1.4	2.3	2.1	1.1	1.8	2.3	2.1	1.5	1.9	2.0	-	4.3	2.6
Li	9	10	7	9	8	3	8	8	6	6	2	3	5	13	6	-	10	9
Sn	-	-	1.1	0.8	2.4	1.6	1.4	3.6	5.4	4.4	5.9	7.5	4.8	1.7	12.0	-	2.9	2.2
Au**	-	-	<1	<1	1	2	<1	<1	<1	4	<1	<1	2	<1	<1	-	3	<1

*Major element and most trace element analyses by X-ray fluorescence using fused disks and pressed powder pellets, respectively, at the Regional XRF Centre, St. Mary's University, Halifax (Chief Analyst, K. Cameron). U by neutron activation analysis (Atomic Energy of Canada, Ottawa); Li by atomic absorption spectrometry and Sn by emission spectroscopy (Minerals Engineering Centre, Technical University of Nova Scotia, Halifax); Au by neutron activation analysis (Chemex Labs Limited, Mississauga). $Fe_2O_3^T$ is total iron expressed as Fe_2O_3. LOI is % weight loss after heating to 1000°C for 1 hour. Normative mineralogy calculated with $Fe^{3+}/Fe^{2+}=0.15$.

**in ppb

S_1 foliation. In the northern part of the map area, the S_1 foliation strikes essentially northeast and dips at moderate angles to the northwest. In the southern part of the area, dips are mainly to the south and locally to the northeast. Large-scale anticlinal folding or doming on an axis trending NE-SW west of Alma (at southeast end of cross-section, Fig. 2) can partly explain these changes in S_0 and S_1 orientations. Related D_2 structural features include crenulation and crenulation cleavage (S_2) in fine grained rocks and inhomogeneously developed shear foliation(s) in granitoid plutons. The sub-parallelism of these later S-planes with the S_1 foliation suggests that later movements essentially along S_1 (transposition) may have occurred, and this may at least partly explain the cataclastic overprint observed in the fabric of many of the metamorphic rocks. Local deviations in the main D_1 and D_2 structural trends can be attributed to other, presumably younger, tectonic events, and include cross-folding and kinking. Faults also contribute to the structural complexity; the dominant trend is northeast but weaker fault systems trend north-northeast and north. Few of these observed or inferred faults can be shown on the scale of Figure 2.

Unit 1 consists mainly of fine grained mafic tuffaceous rocks, many of which are now strongly cleaved. However, the unit is heterogeneous and also includes intermediate to felsic crystal and minor lithic tuffs and flows, as well as mafic flows. Locally grey felsite layers and quartz veins with abundant pyrite are present. Copper, gold, and silver were reportedly mined from quartz veins in unit 1 near the coast about 3 km southwest of the mouth of Point Wolfe River.

Unit 2 is characterized by fine grained, strongly cleaved tuffaceous and volcanogenic sedimentary rocks. The latter are dominantly slates and phyllites which vary in colour from pale grey to green to maroon and black. Rarely, quartz-rich medium- to fine-grained arkosic rocks are present. Rocks of this unit host the Teahan Zn-Cu-Pb-Ag-Au prospect (Ruitenberg et al., 1979). Rocks outcropping in the vicinity of the prospect are a mixed sequence of tuffaceous and phyllitic (phyllonitic) rocks typical of the unit as a whole, but in the area of the old shaft float is dominated by yellowish-white quartz-rich schist, as well as abundant carbonate-quartz-talc rock and vein quartz.

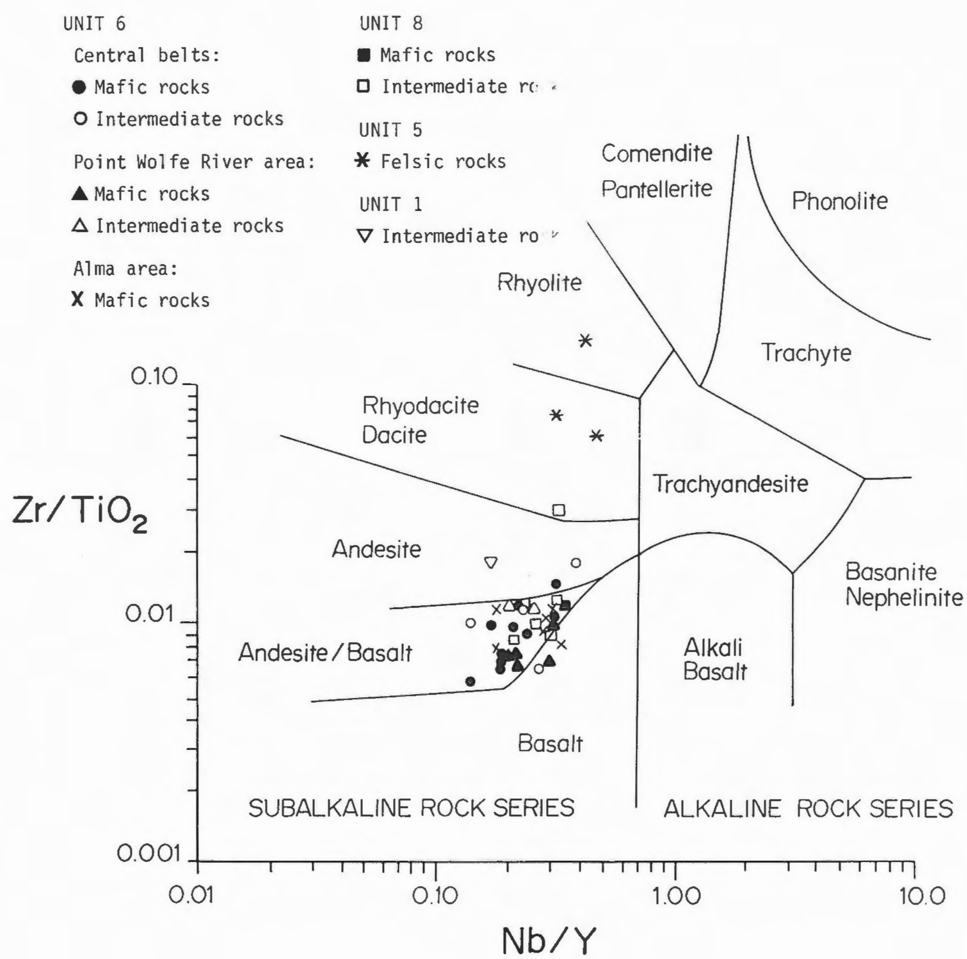

Figure 6. Plot of Zr/TiO2 against Nb/Y for metavolcanic rocks. Fields from Winchester and Floyd (1977).

Unit 3 consists of distinctive arkosic rocks, typically grey to pinkish in colour and varying from coarse-grained and pebbly to medium grained and equigranular. Clasts are generally subangular, poorly sorted, and composed mainly of quartz with less abundant (up to about 15 %) alkali feldspar and albite. Lithic fragments are mainly quartzitic and rarely felsic tuffaceous. No granitoid fragments were observed.

Unit 4 is also a dominantly sedimentary sequence. Consistent younging directions indicate a continuous stratigraphic sequence through the sedimentary rocks of units 3 and 4 into rhyolites and basalts of units 5 and 6, respectively. The most abundant lithology in unit 4 is grey slate with rare layers or lenses of distinctive white-weathering muscovite-bearing quartzite. Also common are red to maroon slate, and grey, red, or maroon phyllite, meta-siltstone and meta-sandstone. Locally conglomerate horizons are present, some with distinctive maroon-coloured matrix and abundant white quartz clasts. Probable tuffaceous layers and various mafic and porphyry intrusions are also present.

Unit 5 consists of rhyolite, pink to grey in colour and typically displaying flow banding. The rhyolite varies from massive to cleaved (phyllonitic), the former appearing to be ignimbritic. Some horizons contain crystal fragments and are clearly tuffaceous. Interbedded mafic tuff or flow layers occur rarely within the sequence.

Unit 6 is dominated by a series of basaltic flows. Outcrop is insufficient to map the number of flows, but at least three are present in the western part of the map area, as indicated by the exposure of amygdaloidal flow tops. The amygdales are typically large (up to 2 cm or more in maximum diameter) and filled with chlorite and/or epidote. The flows are generally massive, but locally sheared and cleaved. Only minor mafic tuffaceous rocks occur in unit 6.

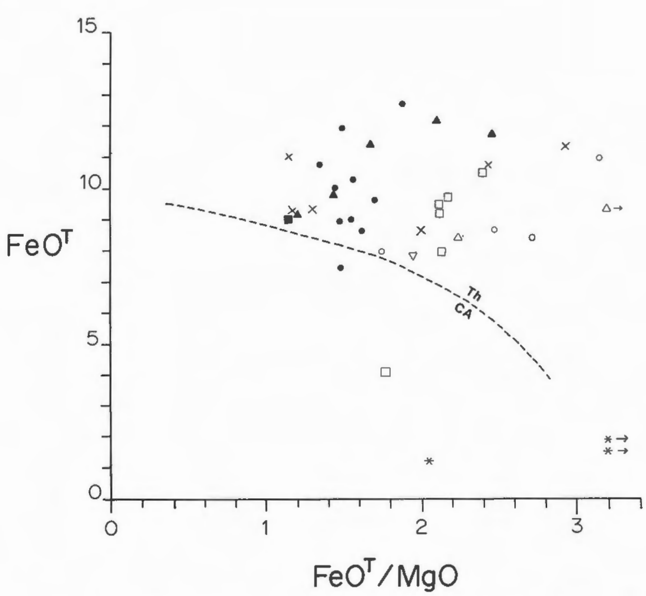

Figure 7. Plot of FeO (total) against FeO/MgO. Symbols as in Figure 6. Tholeiitic – calc-alkalic dividing line from Miyashiro (1974).

Unit 7 appears to overlie basalt in the southern part of the map area. Rocks equivalent to unit 7 are not preserved in the central part of the map area, where they may have been cut off by the intrusion of the Point Wolfe River pluton. The most abundant lithology in unit 7 is medium grained crystal tuff containing 20-50 % plagioclase (now albite) crystals in a fine grained groundmass dominated by sericite, feldspar, and quartz. A second characteristic rock type is grey pyritic felsite consisting of very fine grained quartz and sericite with minor albite. Mafic and felsic porphyry sills and dykes are very common.

Unit 8 occurs on the northern side of the Point Wolfe River pluton (Fig. 2). Mainly tuffaceous rocks are exposed in Pollett River, but farther to the west, mafic schists and metabasalts are dominant. Overall, these rocks appear to be less deformed than elsewhere in the map area.

Mafic sills and dykes are common, both in granitoid and metavolcanic-metasedimentary units. Most are less than 2 m in width and are conformable with layering and/or foliation in the host rocks, if present. Some are porphyritic with plagioclase phenocrysts. Many of these intrusions contain relict clinopyroxene and do not appear to have experienced the regional greenschist facies metamorphism which affected the metavolcanic-metasedimentary units. However, others have mineralogy similar to the host metavolcanic-metasedimentary rocks and may be feeders to the volcanic rocks. Locally larger sill-like bodies of coarser grained gabbroic composition are present. Also present in the area are less abundant felsic sill and dykes including felsite and quartz-feldspar porphyry.

Geochemistry

Analytical work has concentrated on the metabasaltic flow rocks of map unit 6, from which a total of 27 samples have been analyzed. Also analyzed were 7 samples from unit 8, one sample from unit 1, and 3 rhyolitic samples from map unit 5. This data base also includes volcanic samples from the 1985 map area (Barr, 1987). Tuffaceous and amygdaloidal samples were generally avoided in selecting samples for analysis.

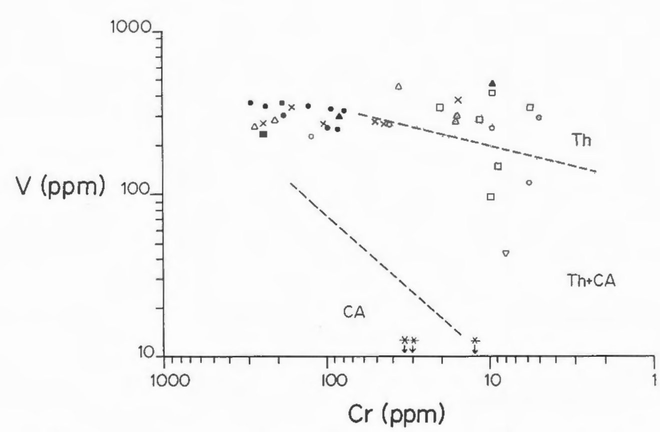

Figure 8. Plot of V against Cr. Symbols as in Figure 6. Tholeiitic (Th) and Calc-alkalic (CA) dividing lines after Miyashiro and Shido (1975).

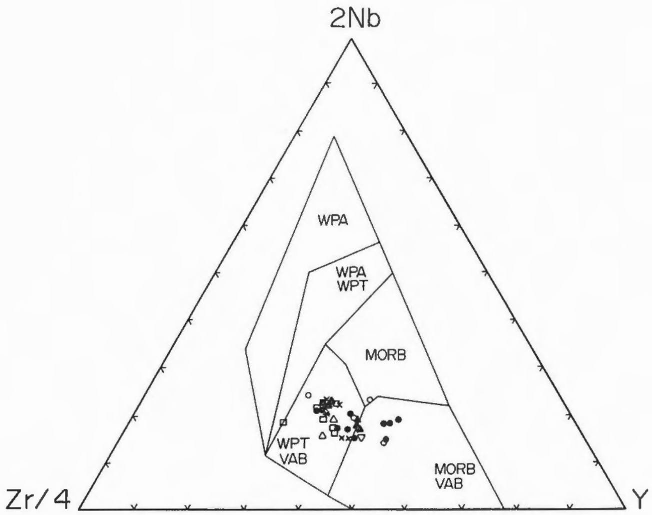

Figure 9. Mafic to intermediate metavolcanic rocks plotted on a Zr-Y-Nb ternary diagram with fields from Meschede (1986). Symbols as in Figure 6.

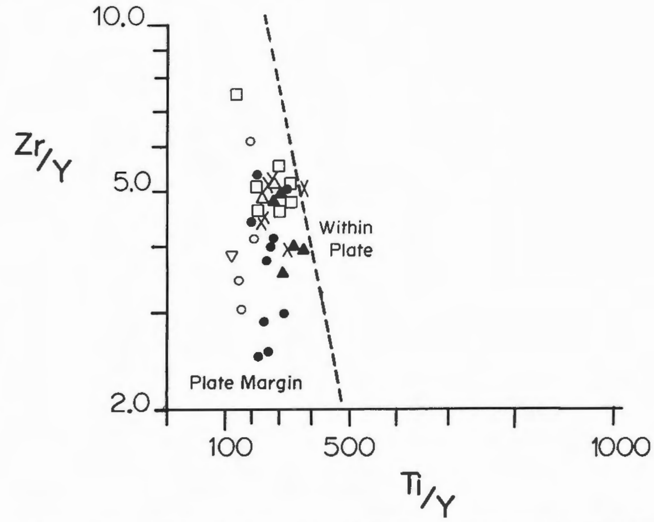

Figure 10. Plot of Zr/Y against Ti/Y. Symbols as in Figure 6. Within-plate and plate-margin dividing line from Pearce and Gale (1977).

Table 3. Chemical analyses* of metavolcanic rocks from map unit 6 in the central and Point Wolfe River areas

	Central Area														Point Wolfe River Area						
%	006	1029B	1054	1133	1142A	1153	1156	1164	1310	1318	1343B	1344A	1355	1754B	085	155C	182	188	191	603	1835
SiO_2	57.38	49.75	50.80	51.13	52.88	55.26	46.14	44.48	49.92	55.13	49.84	56.19	46.91	51.77	51.84	47.50	47.91	47.61	48.09	57.58	54.00
TiO_2	1.18	1.86	1.16	1.23	0.83	1.12	1.00	1.12	1.68	1.98	0.81	1.44	1.27	1.55	2.07	1.59	1.71	1.31	1.79	1.24	1.08
Al_2O_3	15.09	16.35	16.33	16.52	16.87	14.11	17.39	18.86	15.38	14.96	18.48	15.63	16.03	15.44	16.05	16.89	16.64	17.15	16.89	17.02	17.69
$Fe_2O_3^T$	9.24	11.42	9.95	10.73	8.32	12.21	12.04	13.25	11.12	8.90	9.61	9.62	14.15	9.95	13.08	12.67	10.91	10.18	13.51	10.40	9.42
MnO	0.15	0.20	0.21	0.19	0.16	0.20	0.17	0.19	0.21	0.15	0.16	0.20	0.19	0.28	0.22	0.20	0.19	0.19	0.21	0.09	0.16
MgO	3.07	6.57	5.82	5.67	5.04	3.49	8.07	8.11	6.95	4.60	5.37	3.50	6.78	6.07	4.81	6.86	6.92	7.65	5.80	1.63	3.80
CaO	4.53	5.52	8.04	6.31	5.34	8.23	9.13	6.98	8.43	5.87	9.26	6.68	7.28	9.21	2.53	4.14	7.31	7.09	4.14	2.97	4.32
Na_2O	4.57	4.30	4.28	5.44	4.55	3.30	3.05	3.31	3.19	5.12	2.92	3.61	2.89	4.09	5.95	5.31	3.17	3.17	5.45	6.32	6.05
K_2O	0.19	0.31	0.79	0.02	1.27	0.35	0.34	0.56	0.37	0.12	0.36	0.36	0.23	0.46	0.13	0.54	1.76	1.54	0.68	0.91	0.55
P_2O_5	0.38	0.33	0.18	0.32	0.18	0.16	0.13	0.14	0.25	0.24	0.19	0.53	0.21	0.11	0.40	0.34	0.33	0.38	0.35	0.24	0.25
LOI	4.12	3.30	2.40	3.10	5.50	1.40	3.50	4.00	2.80	4.70	3.70	2.50	4.10	0.70	2.91	3.88	3.03	3.42	3.12	1.39	2.70
Sum	99.90	99.91	99.96	100.66	100.94	99.83	100.96	101.00	100.30	100.77	100.70	100.26	100.04	99.63	99.99	99.90	99.88	99.69	100.03	99.79	100.02
ppm																					
Ba	39	60	407	43	318	96	159	166	84	60	461	111	96	129	85	268	516	787	368	314	303
Rb	3	0	14	0	39	5	8	20	5	0	6	4	4	10	1	10	41	32	13	28	7
Sr	196	154	190	220	197	212	109	150	224	234	578	312	245	209	194	136	341	336	291	369	411
Y	33	43	29	32	22	35	26	27	37	37	16	39	31	29	40	27	30	26	33	27	23
Zr	204	171	110	145	119	109	68	74	148	125	83	158	91	87	145	104	147	129	127	126	118
Nb	13	9	5	7	7	5	5	5	9	10	5	9	6	4	8	6	9	8	7	7	5
Cu	85	149	79	52	30	17	15	42	101	9	42	4	87	7	30	7	7	15	11	15	52
Pb	5	9	2	1	5	14	13	6	21	10	7	5	7	13	11	7	9	12	10	9	2
Zn	114	144	93	107	77	100	106	125	197	93	98	112	121	118	165	141	131	191	195	69	90
Ni	12	37	42	44	20	17	70	79	53	36	10	13	46	55	10	37	79	59	18	4	5
Cr	10	196	178	99	41	5	87	82	257	129	94	6	133	296	10	88	280	219	37	18	18
V	244	352	290	242	252	285	243	301	334	212	323	111	337	342	449	290	245	275	433	270	291
Ga	20	21	19	22	22	18	24	22	19	18	23	25	21	20	24	18	20	21	22	17	22
Th	13	0	0	0	0	1	0	0	0	0	0	0	1	0	0	0	0	2	2	1	0
U	-	0.5	0.5	0.3	1.6	0.7	<0.2	<0.2	0.4	0.7	1.2	1.0	0.2	<0.2	-	-	-	-	-	-	1.1
Li	-	30	16	22	20	10	22	40	30	33	14	8	46	4	-	-	-	-	-	-	11
Au**	10	<1	2	<1	<2	<1	<1	<1	3	3	3	<1	<1	<1	<1	3	15	9	-	10	<4

*Analytical methods as in Table 2.
**in ppb

Alteration and low-grade metamorphism are reflected in generally high loss-on-ignition values, which average about 3.5 % in the mafic and intermediate rocks. These processes may also be responsible for much of the chemical variation, especially in the more mobile elements such as Na_2O, K_2O, Rb, and Sr. Hence the following discussion emphasizes those elements which are generally considered to be least mobile during these processes, and also elemental ratios which are also less likely to be changed by secondary processes than are absolute element abundances.

Most of the samples from unit 6 are classified as basalt on the basis of SiO_2 contents less than about 53 % (Tables 3, 4). In contrast, analyzed samples from units 1 and 8 are mainly andesitic. Petrographically, the andesitic samples have higher proportions of plagioclase (now albite) and fewer mafic minerals (mainly chlorite) than the basalts.

Using immobile element ratios, less distinction is apparent between the basaltic and andesitic rocks, most of which plot in the andesite-basalt field (Fig. 6). They consistently display subalkaline chemical characteristics. High iron relative to iron plus magnesia contents indicate that all of these rocks are tholeiitic rather than calc-alkaline (Fig. 7). This is further supported by near-horizontal trends on a V-Cr diagram (Fig. 8), a characteristic feature of tholeiitic magma series (Miyashiro and Shido, 1975). This diagram also more clearly separates the mafic samples with generally higher Cr from the intermediate samples with lower Cr. The three analyzed rhyolite samples are all high in silica, sodium, and potassium (Table 2). Unlike the associated mafic rocks, they are calc-alkalic (Fig. 6, 7, 8).

The magmas were probably generated in a volcanic arc environment, as suggsted by their span of the volcanic arc basalt fields in Figure 9. Although some mid-ocean ridge basalts and within-plate tholeiitic rocks can also plot in these fields, the field assocations are clearly not compatible with the former and low Ti/Y ratios clearly indicate a plate margin setting (Fig. 10).

AGE OF MAP UNITS

The Point Wolfe River pluton was inferred to be Ordovician or older by Ruitenberg et al. (1979), and a late Precambrian age has generally been accepted (e.g. Ferguson and Fyffe, 1985). A Rb-Sr isochron age of 530 $\pm$ 16 Ma (quoted in McLeod, 1986) has been obtained for the Bonnell Brook pluton to the southwest of the Point Wolfe River pluton (Fig. 1), and Barr (1987) reported ages of 597 $\pm$ 18 Ma for the Fortyfive River granodiorite and 598 $\pm$ 27 Ma for the Alma diorite.

Table 4. Chemical analyses* of metavolcanic rocks from map unit 6 in the Alma area and from map units 8, 1, and 5.

	Unit 6						Unit 8							Unit 1	Unit 5		
%	577A	578B	1016	1377	1378	1712	1122A	1122B	1691	1700	1748	1751	1794	1177B	080	600	1134A
SiO_2	50.66	47.15	44.83	52.70	49.38	51.65	67.17	53.66	53.52	49.45	57.05	55.17	53.98	62.93	77.50	74.02	78.84
TiO_2	1.45	1.45	1.52	1.73	1.15	1.56	0.81	1.71	1.67	1.39	1.76	1.84	1.96	1.30	0.23	0.34	0.10
Al_2O_3	17.09	18.95	17.22	14.96	16.04	16.39	14.54	15.06	14.66	16.21	15.31	15.04	15.05	14.03	11.52	14.16	11.99
$Fe_2O_3^T$	9.67	10.43	12.23	12.62	10.17	11.92	4.50	10.23	11.79	10.14	8.82	10.50	10.83	8.67	1.52	1.95	1.37
MnO	0.19	0.20	0.22	0.19	0.18	0.22	0.08	0.14	0.17	0.15	0.17	0.16	0.18	0.16	0.04	0.05	0.03
MgO	4.34	7.25	9.63	3.88	7.96	4.42	2.27	4.34	4.42	7.99	3.73	4.48	4.49	4.00	0.00	0.80	0.61
CaO	8.08	4.14	5.92	5.38	8.09	5.90	2.01	4.91	7.69	7.67	3.57	3.99	5.85	0.95	0.14	0.81	0.14
Na_2O	4.44	4.97	3.68	5.66	3.64	4.95	5.69	5.51	3.05	2.93	5.38	3.92	4.42	3.67	4.59	4.89	3.63
K_2O	0.14	0.47	0.70	0.50	0.62	0.36	1.16	0.89	1.02	0.81	1.41	1.04	0.75	1.49	3.14	2.66	3.05
P_2O_5	0.52	0.55	0.40	0.36	0.30	0.41	0.22	0.23	0.27	0.36	0.55	0.32	0.37	0.46	0.03	0.05	0.02
LOI	3.48	4.16	4.30	2.40	3.00	2.40	1.70	4.20	1.60	3.70	1.90	2.90	2.20	2.80	0.47	0.76	0.70
Sum	100.06	99.69	100.65	100.38	100.53	100.18	100.15	100.88	99.86	100.80	99.65	99.36	100.08	100.46	99.23	99.68	100.48
ppm																	
Ba	133	177	252	307	218	256	296	278	275	246	371	281	169	199	683	828	733
Rb	0	6	10	4	7	1	39	25	40	8	47	36	11	43	58	61	79
Sr	510	232	382	270	369	373	152	171	169	371	217	249	138	23	56	130	30
Y	28	32	24	33	24	39	31	34	33	31	44	34	37	60	39	17	37
Zr	149	165	120	130	109	173	232	186	152	161	202	160	196	228	164	180	156
Nb	8	10	8	6	7	7	10	11	10	10	10	7	10	10	12	8	15
Cu	239	38	19	71	34	32	11	16	27	51	7	8	2	0	0	0	0
Pb	5	10	0	5	0	10	25	8	10	0	10	22	13	10	13	16	7
Zn	126	156	129	131	97	171	63	84	109	103	117	108	144	450	46	42	50
Ni	14	27	44	12	46	20	13	19	17	75	16	14	17	26	14	7	20
Cr	51	105	172	16	255	47	10	6	21	264	9	13	10	8	30	31	14
V	275	258	330	383	261	274	95	330	326	230	144	277	411	42	0	4	3
Ga	24	22	25	26	21	21	23	26	22	20	26	23	27	28	14	16	16
Th	0	0	0	0	0	0	9	2	0	0	2	4	0	1	5	8	11
U	-	-	0.5	1.0	0.6	1.6	2.6	1.6	1.5	1.1	1.7	1.6	1.8	0.9	-	-	1.8
Li	-	-	18	9	19	9	14	23	9	22	14	15	20	11	-	-	4
Au**	30	<1	1	<1	2	<4	12	<2	<1	<1	1	<1	<1	9	10	4	2

*Analytical methods as in Table 2.
**in ppb

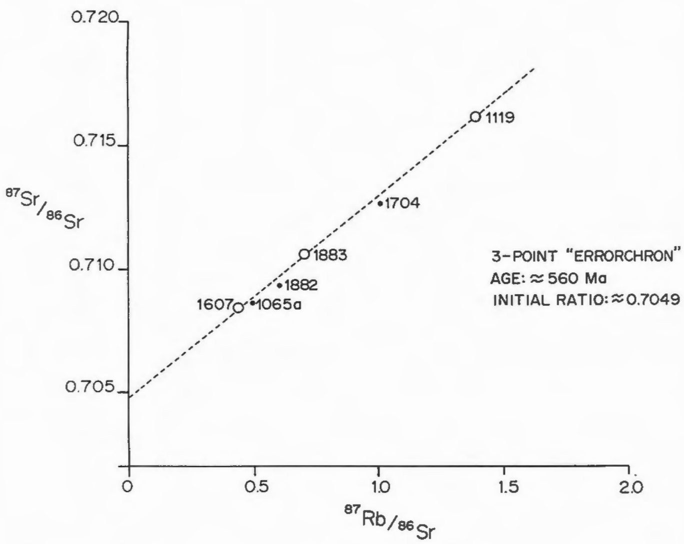

Figure 11. Rb-Sr isochron plot for samples from the Pollett River granodiorite. Age and initial ratio are based on the line through three samples.

Six samples from the Pollett River granodiorite, selected to show the least effects of the shearing and alteration which are pervasive in the pluton, do not give a reliable age (Table 5, Fig. 11). However, three of the samples define a line indicating an age of about 560 Ma, which can probably be considered to be a minimum age for the pluton. The apparent initial ratio (O.7049) is similar to that obtained for the Fortyfive River pluton (0.7029) and the Bonnell Brook pluton (0.7043).

Hence it seems likely that all of these plutons are of similar late Precambrian age and have low initial strontium isotope ratios, consistent with their inferred tectonic setting.

ECONOMIC ASPECTS

The characteristics of mineralization at the Teahan prospect and in other parts of the Caledonian Highlands have been summarized by Ruitenberg et al. (1979). The area continues to be a focus of exploration, most recently with emphasis on gold. The widespread occurrence of pyritiferous rocks, especially within map units 1 and 7, suggests that such exploration is warranted.

Analyses for Au in the samples of this study indicate background values of 1 to 10 ppb in the metavolcanic rocks (Tables 3, 4), with some higher values up to 30 ppb in metabasalts. The granitoid rocks have lower background values (generally less than 1 ppb) and a maximum value of 4 ppb in one of the samples from the Pollett River granodiorite (Tables 1, 2).

CONCLUSIONS

The abundance of pyroclastic rocks, subaerial volcanism, and arkosic sedimentary rocks in the map area are consistent with a volcanic arc environment. This is further substantiated by the geochemical data which indicate that the basaltic and andesitic rocks are volcanic arc tholeiites. This suggests that they formed in a relatively primitive subduction zone environment (Miyashiro, 1974), although the presence of rhyolite probably attests to the involvement of continental crust.

The expanded ''I-type'' nature of the granitoid rocks is consistent with origin at a convergent plate margin (Chappell and White, 1974; White and Chappell, 1983; Pitcher, 1982; Pearce et al., 1984). All of these granitoid units are of probable late Precambrian age, and show similar compositional trends which are typical of calc-alkalic suites. The granitoid rocks do not appear to be co-genetic with the tholeiitic volcanic rocks, which are probably somewhat older and represent an earlier phase of arc magmatism and probably different source rocks.

Table 5. Rb-Sr data* for samples from the Pollett River granodiorite unit of the Point Wolfe River pluton.

Sample Number	Rb	Sr	Rb/Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
1065A	64	372	0.174	0.503	0.70864
1119	166	345	0.481	1.393	0.71610
1607	68	450	0.152	0.439	0.70842
1704	111	315	0.353	1.021	0.71259
1882	80	323	0.211	0.609	0.70929
1883	81	338	0.246	0.712	0.71061

*Error limits are: Rb and Sr, ±5%; $^{87}Rb/^{86}Sr$, ±2%; $^{87}Sr/^{86}Sr$, ±0.006%. λ Rb = 1.42 x 10^{-11} y^{-1}. Analyses by J. Blenkinsop, Carleton University. Rb, Sr, and Rb-Sr ratios were determined by X-ray fluorescence. Sr was separated from the samples by conventional ion exchange techniques. Sr isotope ratios were measured in static mode on a Finnigan-MAT 261 multicollector mass spectrometer.

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Stratigraphy and volcanology of a portion of the Lower Devonian volcanic rocks of southwestern New Brunswick[1]

Nancy A. Van Wagoner[2] and Vincent K. Fay[3]

Van Wagoner, N.A. and Fay, V.K., Stratigraphy and volcanology of a portion of the Lower Devonian volcanic rocks of southwestern New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 69-78, 1988.

Abstract

A sequence of Devonian volcanic and sedimentary rocks along the coast of Passamaquoddy Bay was correlated with the Early Devonian Eastport Formation of Maine. The sequence comprises interbedded rhyolitic, andesitic and basaltic flows and pyroclastic rocks, and red and green shale and sandstone. Rhyolitic rocks are volumetrically most important, and intermediate rocks are rare. Mafic units form flows, scoria cones, phreatomagmatic tuff cones, and peperitic breccia deposits. Felsic units were emplaced as welded and nonwelded air-fall, ash cloud and ground surge, and pyroclastic flow deposits with rare lava flows. The lower sedimentary units are littoral deposits while the uppermost units are fluvial deposits, suggesting completely subaerial volcanism, with the elevation of the vents gradually increasing with time. Sulphide mineralization is mainly associated with mafic volcanic breccias deposited near sea level.

Résumé

On a corrélé une séquence de roches volcaniques et sédimentaires dévoniennes bordant la côte de la baie Passamaquoddy avec la formation d'Easport, d'âge dévonien inférieur, située dans le Maine. Cette séquence comprend des coulées rhyolitiques, andésitiques, basaltiques et des roches pyroclastiques, ainsi que des schistes argileux et des grès rouges et verts. Les roches rhyolitiques occupent le plus grand volume, les roches intermédiaires sont rares. Les unités mafiques forment des coulées, des cônes de scories, des cônes de tuf phréatomagmatique, et des dépôts de brèches pépéritiques. Les unités felsiques se sont déposées sous forme de dépôts pyroclastiques lithifiés ou non lithifiés, de nuages de cendres et de déferlantes basales, et aussi de dépôts pyroclastiques accompagnés de rares coulées de laves. Les unités sédimentaires inférieures sont des dépôts littoraux, les unités sommitales des dépôts fluviatiles, ce qui suggère que le volcanisme a été entièrement augmenté avec le temps. Les minéralisations sulfurées sont principalement associées aux brèches volcaniques mafiques qui se sont déposées à proximité du niveau de la mer.

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INTRODUCTION

A sequence of Devonian volcanic and sedimentary rocks is exposed along the coast of Passamaquoddy Bay, southwestern New Brunswick (Fig. 1). Pickerill and Pajari (1976) correlated these rocks with the Early Devonian (Gedinnian) Eastport Formation of Maine, and proposed that they also be termed the Eastport Formation.

Previous investigations of the area include Gesner (1839), Matthew (1865), Bailey and Matthew (1872, 1876), MacKenzie (1940), MacKenzie and Alcock (1960), Perry and Alcock (1960), Cumming (1967) and Ruitenberg (1968). The most detailed studies are by Hay (1967), Whaley (1981), and Fay (MSc thesis, University of New Brunswick, in prep.), who divided the Eastport into several distinct lithological types.

The purpose of this study is to determine the volcanic stratigraphy and facies relationships, physical volcanology, paleogeography, and volcanic evolution of the Eastport Formation of southwestern New Brunswick. This paper describes the stratigraphy of a portion of these rocks. The terminology used for the pyroclastic rocks is after Fisher (1966) and Schmid (1981). The usage of the term 'peperitic breccia' is from Cas and Wright (1987).

FIELDWORK

Fieldwork was completed during August, 1984 (Van Wagoner, 1984) and late July-August, 1986 (Van Wagoner, 1986). Mapping was done on 1974 and 1984 series air photographs at 1:10 000 scale. Stratigraphic sections were described on a bed by bed basis. Fieldwork concentrated on the area between Digdeguash Harbour and Bocabec Bay, south of the St. George Pluton. A total of 238 outcrop areas were described.

STRATIGRAPHY

The area comprises a sequence of interbedded rhyolitic, andesitic, and basaltic flows and pyroclastic rocks, and red and green-grey shale and sandstone (Fig. 2). The rhyolitic rocks form the thickest sequences. Intermediate rocks are rare. The extrusive sequence is cut by a series of primarily basaltic sills and dykes and rare rhyolitic intrusions.

Unit Dmfl

This unit comprises interbedded highly to sparsely amygdaloidal mafic flows and/or very shallow level intrusions, peperitic breccias, and mudstone and siltstone (Fig. 3). It has a maximum thickness of about 350 m.

The mafic flows and/or intrusions are 1 — 12 m thick. Groundmass is tachylitic. Microlites are primarily plagioclase with rare clinopyroxene and opaques. Amygdules are 5-15 %, mostly 0.1 — 0.5 cm in diameter, and round to irregular.

The peperitic breccias are fluidized mixtures of quench-fragmented volcanic debris and sediment (Pichler, 1966; Honnorez and Kirst, 1975). The fragmentation occurs either when magma intrudes through (Pichler, 1965; Kokelaar, 1982) or when lava flows over wet sediment (Waters, 1960; Schmincke, 1967). Because of the limited lateral control in this case, it is impossible to tell which happened. These breccias are 2 — 14 m thick and contain 30 to 90 % mafic volcanic clasts in a very fine 'cherty' mudstone matrix. Mafic clasts are the same as the flows described above, except that vesicles contain mudstone. Clasts are 2 mm — 0.5 m in size, and pillow-like to angular. Pillow-like fragments have delicate appendages that protrude into surrounding sediment, indicating that secondary transport was minimal. Most pillow-like clasts have partial chill margins (indicating breakage after chilling), but rare clasts have complete chill margins. In some cases, the margins of clasts or entire clasts are broken in a jig-saw puzzle fashion indicating quench shattering. Breakage of clasts by thermal quenching rather than expansion of volatiles is also indicated by the lack of vesicle controlled boundaries. The matrix is fluidized mudstone containing rare irregularly-shaped vesicles formed by vapourization of interstitial fluid when heated upon contact with magma or lava. Fluidization is indicated by soft-sediment deformation structures, and by penetration of the mudstone into vesicles of the lava. Euhedral pyrite crystals are concentrated in brecciated, peperitic and vesiculated zones.

Interbedded mudstone and siltstone layers are 1 — 5 m thick. Beds are 1 mm — 1 cm thick and massive, normally graded, internally finely laminated, or rarely cross-laminated. Soft sediment deformation structures are common and caused by a combination of density inversion and interaction of sediment with lava or magma.

This sequence formed by the interaction of lava or magma with cold, wet sediment. Quench fragmentation occurred around the margin of the flows or shallow level sills producing the peperitic breccias. Intrusion of magma or mixing of the lava with sediment resulted in the soft-sediment deformation. The emplacement of mafic lava above unconsolidat-

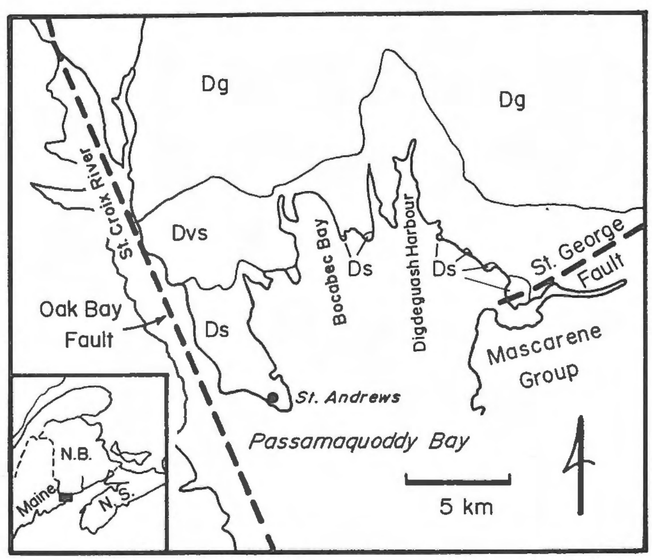

Figure 1. Index map of the study area showing the general geology after Pickerill and Pajari (1976) and Rutenberg and McCutcheon (1980). Dg = Devonian granitoid rocks of the St. George batholith, Dvs = Devonian volcanic and sedimentary rocks, Ds = undifferentiated Silurian and Devonian rocks.

ed, uncompacted, wet sediment, could create a density instability that would allow the mafic flow to bulldoze down into the underlying sediment. A subaerial environment is suggested by the absence of palagonite or relict sideromelane glass, pillowed flows, and pillow breccias. The sediments may be mudflat deposits similar to unit Drs1.

Unit Drs1

Unit Drs1 comprises interbedded red siltstone, grey-green mudstone, and rare green to buff sandstone. Maximum thickness is about 334 m. The sequence fines to the north.

The mineral fragments of the siltstone include quartz, plagioclase, and mica. Beds are commonly planar, 2 cm to 2 m thick, and are internally massive, finely horizonally laminated, cross laminated, herring bone cross laminated and normally graded. Other sedimentary structures are interference ripples, and lenticular siltstone ripples up to 5 cm in height in a mudstone matrix. Current ripple cross stratification modified by wave oscillation occurs in the rare sandstone lenses in the southern part of the area. Other sedimentary structures of the sandstone lenses are low angle cross stratification, and ripple cross stratification. Mudcracks and oscillation ripples are common. Local bioturbation occurs as horizontal and vertical burrows.

Unit Drs1 is interpreted to be a peritidal (supratidal, intertidal, and subtidal) deposit (Friedman and Sanders, 1979) based on the sedimentary structures and fine grain size. The sandstone deposits could be beach or bar deposits. This unit is similar to unit Drs2, described below, but lacks the volcanic lithic fragments and fossils observed in unit Drs2.

Dlt1

Unit Dlt1 is olive to brown accretionary and lithic lapilli tuff. It has a maximum thickness of about 50 m.

Lithic clasts are subround pumice, up to 1 cm in size, and up to 40 % of some layers. Accretionary lapilli are in layers 1-2 cm thick, are 3 mm to 1 cm in diameter, and are up to 60 % of the beds that contain them. Matrix is fine-ash tuff.

Beds are up to 0.5 m thick, and internally massive, cross laminated, and normally and reversely graded from fine to medium ash. Climbing ripples occur locally.

The occurrence of accretionary lapilli along with juvenile pumice fragments suggests that this unit originated from a phreatomagmatic eruption (Fisher and Schminke, 1984). Most of the bedding structures are consistant with an air fall deposit. However, the cross-stratification and climbing ripples require deposition by a current, and likely represent surge deposits (Cas and Wright, 1987).

Unit Dlt2

Unit Dlt2 is interbedded crystal tuff and crystal-lithic tuff with rare red and green siltstone and mudstone. Maximum thickness is about 100 m (Fig. 4). The stratigraphic position of this unit is uncertain because it is in fault contact with Dwt1. This is a minor fault, and it is suggested that Dlt2 is older than Dwt1. Drs1 is also older than Dwt1 and Dlt3, however, a contact relationship betweeen Drs1 and Dlt2 was not observed.

Individual tuff units are 5 to 15 m thick and represent a sequence of flow and surge deposits with minor air-fall and siltstone sedimentation (Fig. 4). The absense of welding in most cases along with the thinness of most of tuff units suggests that this sequence was deposited relatively distal to source.

Unit Dwt1 and Drf1

Unit Dwt1 comprises a sequence of banded and welded ash flow tuffs. Maximum thickness is about 270 m. Drf1 is a felsic lava flow exposed near the base of Dwt1.

The tuffs contain minor feldspar phenocrysts. The groundmass is primarily microcrystalline quartz and feldspar produced by granophyric recrystallization of the original glassy groundmass (Lofgren, 1971). Spherulites occur locally. Emplacement structures include flow banding which is contorted in places, minor flow brecciation, and columnar joints.

Unit Dlt3

Unit Dlt3 is interbedded red to buff crystal tuff, crystal-lithic tuff and lapilli tuff, with rare interbeds of red siltstone. The unit fines to the north where tuff units appear cherty. There is some evidence of welding to the south. The unit is interbedded with Dtb1. Maximum thickness is about 534 m.

Crystals are up to 15 %, and mostly feldspar. Lithic lapilli are 1 mm — 1 cm, subround to angular, up to 90 % of some layers, and include pumice, rhyolite, basalt, and rare mudstone. Welded beds contain fiamme. The pumice and fiamme are juvenile fragments, the other lithic fragments are accidental clasts, although the rhyolitic clasts may be cognate ejecta. The accretionary lapilli occur locally where they are up to 90 % of reversely graded layers. They are up to 1 cm in diameter but mostly 2-3 mm. The matrix material is fine to coarse ash. Devitrification textures are common and include spherulites, lithophysae, and granophyric recrystallization.

Beds are up to 2 m thick, internally massive, thinly laminated, normally graded, reversely graded, and multiply graded. Some of the welded tuffs exhibit convolute lamination, possibly due to rheomorphism (Schmincke and Swanson, 1967; Chapin and Lowell, 1979; Wolff and Wright, 1981).

This unit is interpreted to be mostly magmatic pyroclastic fall and flow deposits. The increase in the degree of welding and grain size to the south, suggests a southern source. The reversely graded accretionary lapilli layer suggests local hydroclastic airfall activity.

Unit Dtb1

Unit Dtb1 comprises heterolithic tuff breccias. It is interbedded with unit Dlt3 in the southern part of the area. Maximum thickness is about 115 m.

The tuff breccias contain 30 — 80 % clasts of pumice and rhyolite bombs, rhyolitic blocks, and siltstone and basaltic lapilli. The breccias locally contain fiamme. The pumice and

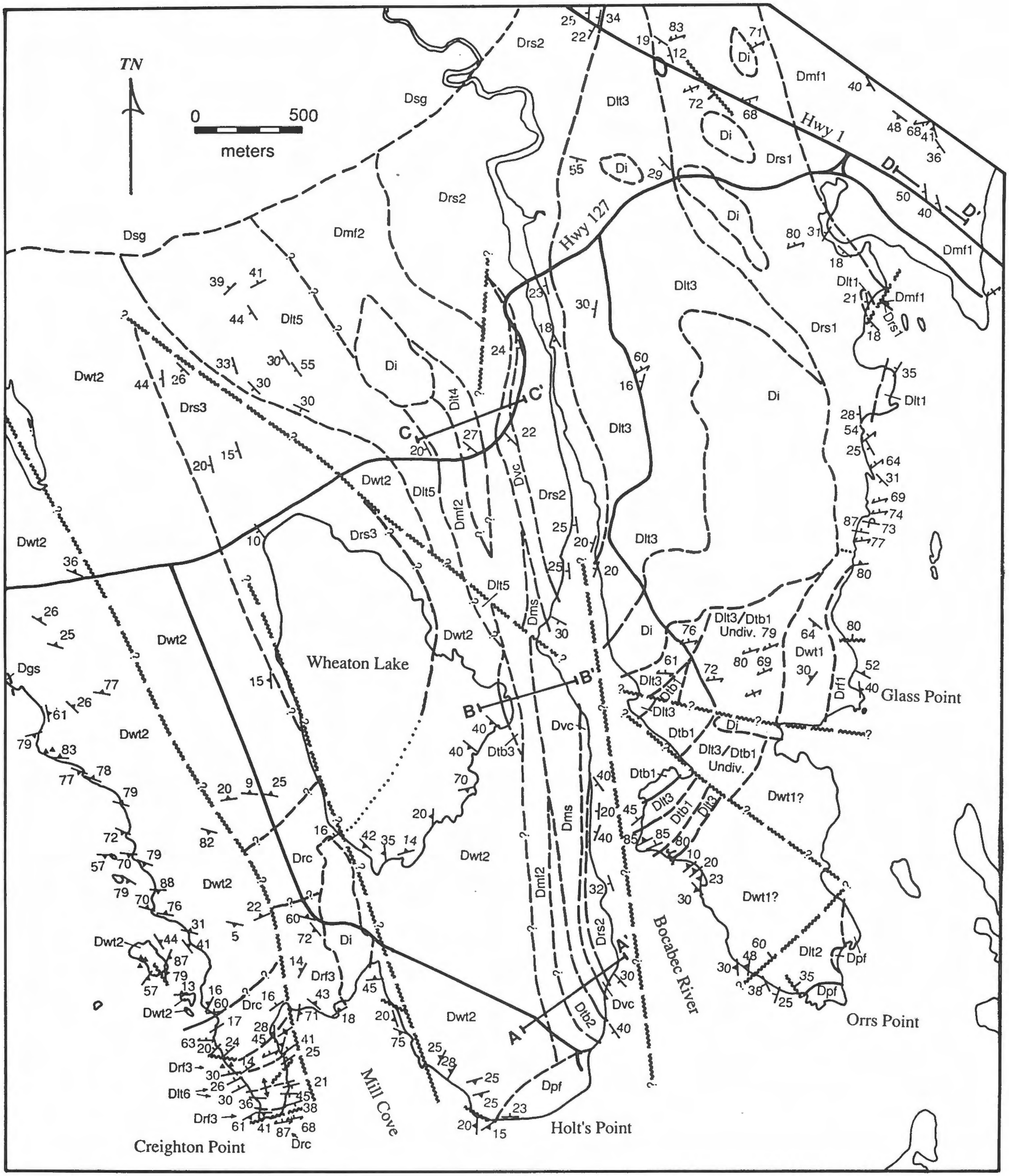

Figure 2. Geologic map of a portion of the Devonian volcanic and sedimentary rock sequence of the Passamaquoddy Bay area, southwest New Brunswick.

Dpf — Dpf--Perry Formation

--unconformity--

Upper Sedimentary Rocks

Dgs — Dgs--Green sandstone

Upper Lava Flows and Pyroclastic Rocks

Drc, Drs3, Dwt2, Drf2, Dlt6

Drc--Red conglomerate, siltstone and mudstone
Drs3--Red siltstone and gray mudstone
Drf2--Banded felsic lava flows and flow breccias
Dlt6--Felsic crystal tuff and heterolithic lapilli tuff
Dwt2--Welded felsic pyroclastic flows

Dlt5 ? Dtb3

Dlt5--Felsic crystal tuff and crystal-lithic lapilli tuff
Dtb3--Mafic heterolithic tuff breccia

Dmf2, Dlt4

Dlt4--Felsic lithic-lapilli tuff and crystal tuff
Dmf2--Highly amygdaloidal mafic lava flows

Dtb2 — Dtb2--Mafic heterolithic tuff breccia

Dms — Dms--Mafic scoria

Middle Sedimentary Rocks

Dvc — Dvc--Red siltstone and heterolithic volcaniclastic (lapillistone) lenses and layers

Drs2 — Drs2--Red mudstone, siltstone and fine sandstone

Lower Felsic Flows and Pyroclastic Rocks

Dlt3, Dtb1

Dtb1--Heterolithic felsic tuff breccia
Dlt3--Felsic crystal tuff,crystal lithic tuff and lapilli tuff

Dwt1, Drf1

Dwt1--Welded felsic pyroclastic flows
Drf1--Banded felsic lava flows

--fault--

Lower Sedimentary and Pyroclastic Rocks

Dlt2 -?- -?- Drs1 Dlt1

Dlt2--Felsic crystal tuff and crystal-lithic tuff with rare interbedded red and green siltstone and mudstone
Drs1--Red siltstone and mudstone with rare green sandstone
Dlt1--Accretionary and lithic lapilli tuff

Lower Mafic Volcanic Rocks

Dmf1 — Dmf1--Sparsely to highly amygdaloidal mafic flows or shallow level sills with brecciated margins, peperitic breccia, mudstone and siltstone

Intrusions

Di — Di--Other intrusions, primarily mafic, with minor intermediate and felsic intrusions

Dsg — Dsg--Devonian granitoid rocks of the Saint George batholith

Contact, dashed where approximately located, dotted where implied

Fault, dashed where approximately located

4 12 Strike and dip of beds, overturned beds

16 Strike and dip of flow banding

21 Strike and dip of cleavage, vertical cleavage

Brecciated zones

Map symbols queried where uncertain

rhyolite bombs, and fiamme are probably juvenile fragments. The other clast types are cognate or accidental clasts. The breccias are massive or vaguely medium to thickly bedded. Beds are very rarely finely laminated and cross laminated. Rare bedding sags occur beneath larger pumice clasts.

Unit Dtb1 is interpreted to be near-vent pyroclastic air-fall and rare ground surge deposits. The surge deposits are represented by the laminated and cross-laminated beds. Such beds are thin (< 30 cm) which is typical of ground surge deposits that occur within air-fall deposits (Roobol and Smith, 1976). The other bedding structures are more typical of air-fall deposits, probably from a magmatic plinian or ultra-plinian eruption (Cas and Wright, 1987). The large grain size and the occurrence of welding indicates that this is a near-vent deposit.

Unit Drs2

Unit Drs2 is brownish-red, thinly- to thickly-bedded mudstone to fine sandstone (Fig. 5) similar to Drs1. Maximum thickness is about 412 m.

The thinner beds are 1-8 cm thick, and are massive, horizontally-laminated or ripple cross-laminated mudstone to siltstone. Upper surfaces of coarser beds preserve interference current ripples, polygonal mud cracks, rain drop impressions, and scratch and drag marks. The thinner beds contain up to 50 % fossils of lingulid brachiopods, bilvalves, ostracods, rare gastropods, possible plant impressions, and vertical non-branching tubular burrows. The fauna are described by Pickerill and Pajari (1976).

The thicker beds are lense-shaped deposits of siltstone to fine sandstone up to 2 m thick. Lenses are vaguely horizontally laminated, low angle cross laminated, or ripple cross laminated. They rarely contain a basal lag gravel of siltstone rip-up clasts.

This unit is interpreted to be a higher mudflat tidal facies (Evans, 1965; Ruitenberg and McCutcheon, 1978; Pickerell and Pajari, 1976). This is suggested by the occurrence of shallow marine fossils, the fine sediment size, sedimentary structures, and the lack of indication of reworking by waves (Elliott, 1978). The coarser lense shaped beds are interpreted to be tidal channel deposits in the mud flat.

Unit Dvc

Unit Dvc is bedded brownish red siltstone as described above with lenses and beds of green volcanic fragments (Fig. 5). Maximum thickness is about 40 m. The abundance of volcaniclastic layers increases upward to 90 %, the unit coarsens upward, and bed thickness increases upward from 2 mm-3 cm to 10-30 cm. Most layers are lensoidal but irregular in shape due to soft sediment deformation.

Red siltstone layers contain about 1-25 % volcanic fragments, and are horizontally laminated, massive or very rarely ripple laminated. Coarse volcaniclastic layers are massive or rarely normally graded.

The lithic clasts of the siltstone and volcaniclastic layers are up to pebble sized, mostly subround, and heterolithic. The clast types are siltstone rip-up clasts, amygdaloidal and

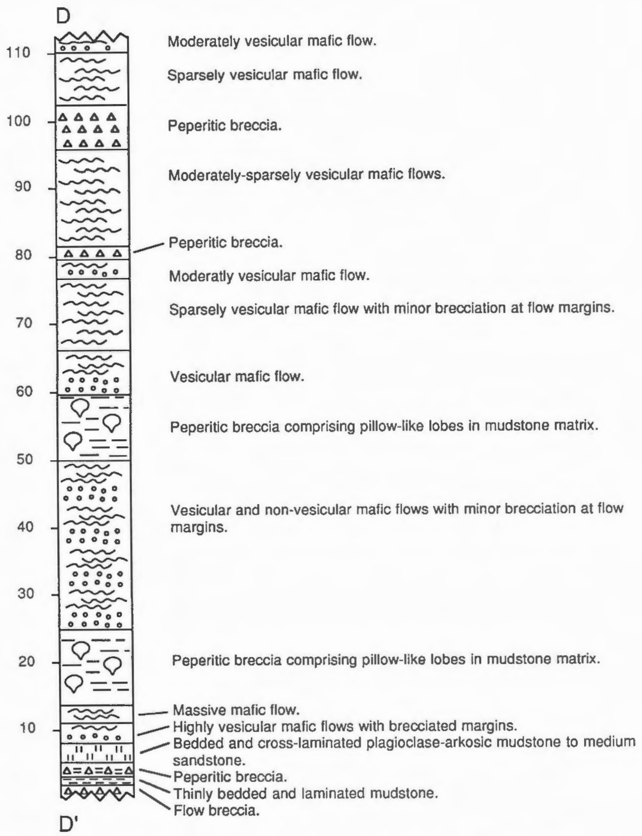

Figure 3. Representative section of unit Dms1. See figure 2 for location of section. Scale is in meters.

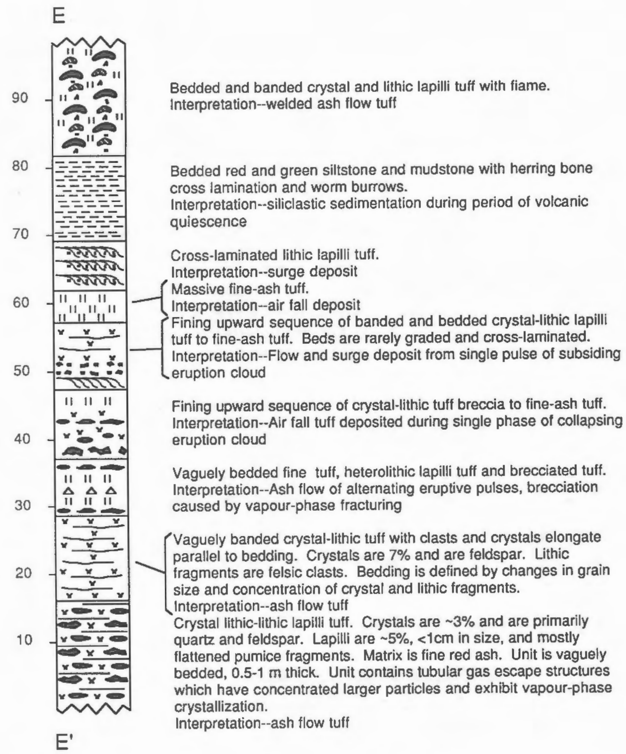

Figure 4. Representative section of unit D1t2. Location of section is shown on figure 2. Scale is in meters.

non-amygdaloidal mafic volcanic fragments, bubble wall and blocky mafic glass shards, rare armoured lapilli consisting of siltstone fragments mantled by volcanic glass, and rare, rhyolitic fragments.

This unit is interpreted to be a tidal mudflat faces, as unit Drs2, which was proximal to an intermittently active mafic volcano(es). The volcanic fragments are interpreted to be air-fall deposits because most (except the glass shards) are too large to have been transported by weak tidal currents. The rare volcanic fragments within the siltstone beds are scoriaceous fragments which were light enough to be transported by tidal currents. Little reworking and rapid burial due to high sedimentation rates in a low energy environment, is indicated by the excellent preservation of the glass shards and glass surrounding armoured fragments.

Unit Dms

Unit Dms is a mafic scoria. The unit is thickest in the vicinity of section B-B' and thins to the north and south (Fig. 5). Maximum thickness is about 82 m.

Basaltic fragments are ash to bomb sized, and commonly scoriaceous. The unit contains rare accidental fragments of siltstone. There is no matrix. It is cemented by agglutination of fluidal clasts and secondary calcite.

This unit is interpreted to be a mafic scoria cone, characteristic of a Strombolian style of volcanism. Close proximity to the vent is suggested by the large size of some of the bombs. The occurrence of accidental fragments of siltstone suggests that the volcano erupted through tidal flat deposits. Scoria cones are poorly indurated and highly susceptible to weathering and erosion (eg. Wood, 1980), so the thickness of the unit may reflect the erosional remnant of the unit rather than the original form. Because erosion of a scoria cone tends to follow the original profile of the cone it is suggested that the source of the cone was toward the centre of the unit. The unit is not symmetrical around its thickest part suggesting a wind to the south during the eruption.

Unit Dtb2

Unit Dtb2 is a vaguely bedded heterolithic tuff breccia. It is about 35 m thick in the southern part of the map area and pinches out to the north (Fig. 5).

Beds are 30 cm to 6 m thick. Clasts are up to 1 m, averaging 1-2 cm, and are angular to subangular. The clasts are siltstone, amygdaloidal and nonamygdaloidal basalt, gabbro, and rare cored bombs of lithic clasts rimmed by basalt. The matrix is nonwelded vitric tuff comprising angular and bubble wall mafic glass shards.

The tuff breccia is interpreted to be a phreatomagmatic explosion breccia. Fragmented country rock is represented by the clasts of siltstone, gabbro, and angular accidental volcanic fragments. Glass shards in the matrix and some of the mafic volcanic clasts comprise the magmatic material. The combination of blocky nonvesicular shards and vesicular shards suggests they formed by a combination of vesiculat-

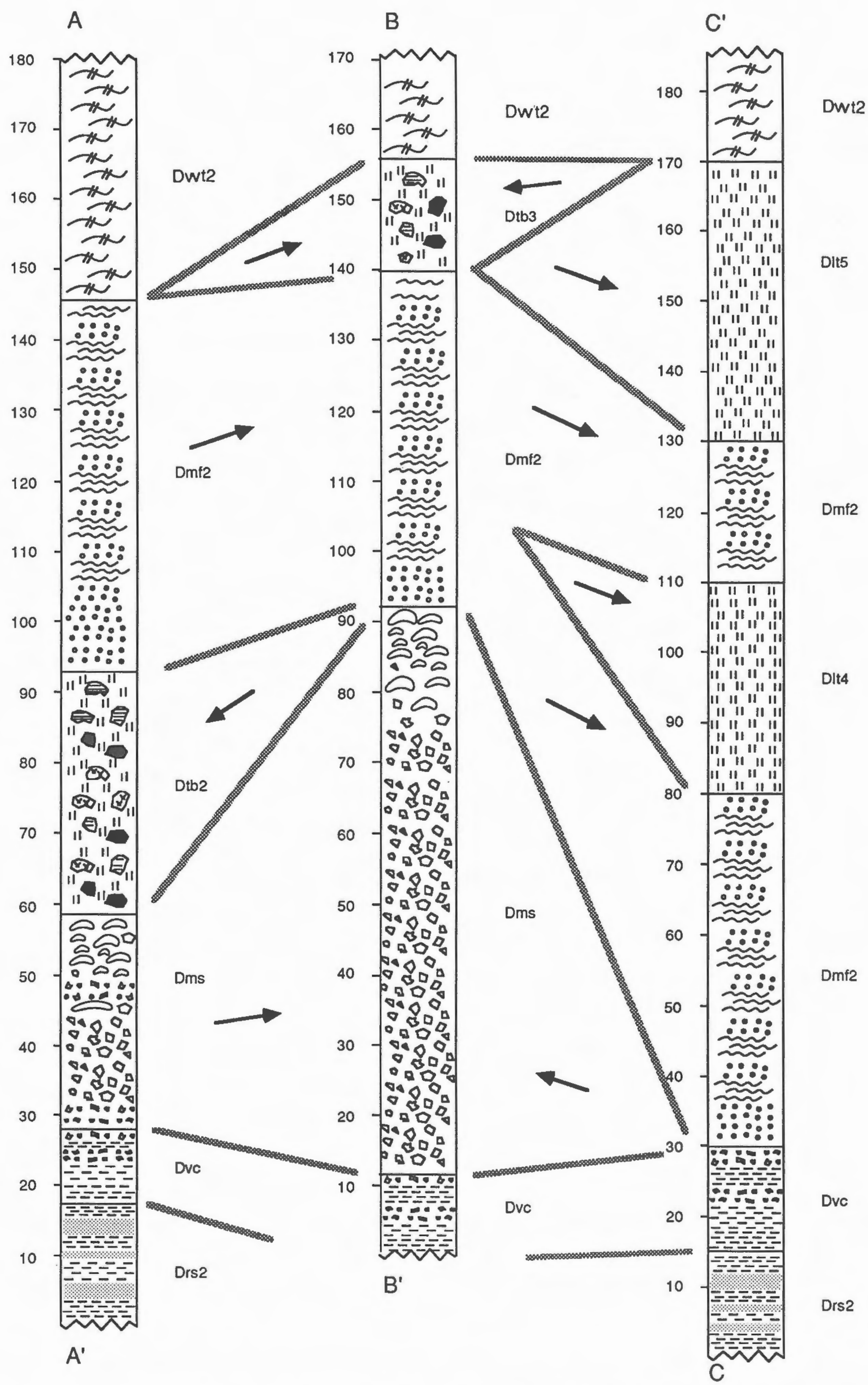

Figure 5. A correlation of units from the middle part of the section. Arrows indicate direction to the source of some of the units. Location of the sections is shown on figure 2. Scale is in meters.

ing magma and quenching by steam. Lack of welding indicates that temperatures were relatively low in the eruption system. The large clast size indicates a high energy eruption. Lack of any internal structures suggests that this is an air-fall deposit, probably around a tuff cone. Because this unit would be poorly indurated when deposited, the shape of the deposit may reflect an erosional remnant.

Unit Dmf2

Unit Dmf2 comprises highly amygdaloidal pahoehoe flows. Maximum thickness is about 280 m in the northern part of the map area. The unit thins to the south (Fig. 5).

Flows are 1-2 m thick. Amygdules are 3 — 85 % and up to 1.5 cm in size. Rare gas cavities are up to 40 x 15 cm. Flow contacts are iron stained and undulatory chill margins 2-3 cm thick.

These flows are interpreted to be subaerial pahoehoe flows. Although pahoehoe flows form in the deep water (eg. Ballard et al., 1979) the high vesicularity of these flows, lack of evidence of interaction between lava and water, and the other rock associations, indicates that these are subaerial flows.

Unit Dlt4

This unit is thinly to thickly bedded, intermediate to felsic lapilli tuff and crystal tuff. Maximum thickness is about 30 m. It is interbedded with Dmf2 and truncated by an intrusion in the northern part of the area (Fig. 5) and pinches out to the south.

The lapilli tuff layers contain up to 90 % angular to rounded cognate clasts of mafic volcanics, pyrogenic feldspar crystals, and accidental siltstone fragments. The matrix is vitric and largly recrystallized. The crystal tuff is thinly bedded but beds are internally massive. The origin of this unit is uncertain. It may be an air-fall deposit.

Unit Dtb3

Unit Dtb3 is a massive heterolithic tuff breccia of limited extent. Maximum thickness is about 15 m and the unit thins rapidly to both the north and the south (Fig. 5).

The unit contains 40 — 60 % clasts of siltstone, gabbro, and amygdaloidal basalt. Clasts are up to 1 m in size. The siltstone and gabbro clasts are angular. Some of the basaltic clasts are bomb-shaped. The matrix is vitric tuff. This unit probably is the remnant of a tuff cone similar to unit Dtb2.

Unit Dlt5

Unit D1t5 in interbedded crystal-tuff and crystal-lithic lapilli tuff. It has a maximum thickness of about 40 m and pinches out to the south. The entire sequence coarsens upward and bed thickness increases upward.

Crystals are alkali feldspar laths which are up to 10 % of some beds. Lithic fragments are mafic and felsic volcanic fragments which are angular to subround, up to 3 cm in size, up to 40 % of the beds, and locally flattened parallel to bedding suggesting welding. The matrix is pumice, rare bubble wall glass shards, and recrystallized vitric material. In places pumice is draped over lithic fragments suggesting welding. The juvenile fragments are volcanic ash, pumice and crystals of alkali feldspar. The cognate fragments are the mafic and felsic lithic clasts.

The unit is thinly to medium bedded. Beds are internally massive, finely laminated, normally and reversely graded, except near the top of the sequence where one bed is cross laminated. Beds are commonly discontinuous. Bedding sags are rare.

This unit is interpreted to be a magmatic air-fall deposit. A magmatic eruption is indicated by the vesicularity of the glass shards and pumice. The bedding structures are mostly typical of air-fall deposits. The crossbedded layer near the top of the unit may be a ground surge deposit. Evidence for welding suggests this is a near vent deposit.

Unit Dwt2

Unit Dwt2 comprises banded rhyolitic welded tuff with minor densely welded heterolitic tuff breccia. It is about 1130 m.

The welded tuff contains vitric groundmass, feldspar, minor mafic minerals, and stretched pumice which defines flow banding. Flow banding is undulating to complexly folded. Flow tops are often brecciated. Devitrification structures include spherulites and lithophysae. The flows contain rare accidental fragments of basalt.

The tuff breccias contain clasts of basalt, rhyolite, and rare gabbro. Clasts are up to 40 cm in size, and mostly subangular. Some of the rhyolitic clasts are slightly stretched due to flowage. Matrix is fine grained banded rhyolite similar to the groundmass of the pyroclastic flows.

Unit Dwt2 is interpreted to be densely welded rhyolitic pyroclastic flows. The dense welding of the tuffs and coarseness of the tuff breccias suggests that these are near-vent deposits.

Unit Dlt6

Unit Dlt6 is interbedded crystal tuff and heterolithic lapilli tuff. Maximum thickness is about 65 m.

Beds are 1-3 cm, or 10-50 cm thick. Thicker layers are welded tuff with 3-5 % fiamme, and up to 70 % angular to subangular felsic and mafic lithic lapilli fragments. Beds are symmetrically graded, normally graded and locally flow banded. The upper 2-3 cm of thicker beds are locally vaguely laminated. The thinner beds are interbedded with thicker beds and are massive fine tuff with crystal and pumice fragments.

The thicker beds are interpreted to be welded ash-flow tuff deposits. The laminated tops of these beds are probably ash cloud surge deposits. The interbedded thin massive tuff beds are probably air-fall deposits. The welding suggests that this is a near-vent deposit.

Unit Drf3

Unit Drf3 is a banded rhyolitic flow that is locally brecciated. It has a maximum thickness of about 75 m and is similar to Drf1.

Unit Drs3

Unit Drs3 is interbedded red siltstone, grey mudstone, and rare green sandstone. The unit has a maximum thickness of about 210 m. It is similar to Drs1 and Drs2, and is also interpreted to be a tidal flat deposit.

Unit Drc

Unit Drc is interbedded red conglomerate, siltstone and mudstone. It has a maximum thickness of 165 m.

Clasts which are pebble size and larger are of amygdaloidal basalt, diabase, rhyolite, mudstone and siltstone. Maximum clast size is 5 cm and most clasts are subangular. The sand-sized particles are quartz, plagioclase, mica, and felsic volcanic rock fragments. The matrix is clay-sized material.

Beds are 0.5 cm to 1 m thick. The thickest beds tend to be the coarsest beds, but finer grained beds form the largest percentage of the unit. The entire sequence coarsens upward and the abundance of volcanic clasts increases upward. Beds are massive, normally graded, current-ripple and ripple-drift cross laminated and parallel laminated. There is rare pebble imbrication in the conglomerate beds. Other sedimentary structures are current ripple marks, oscillation ripple marks, cut and fill forming conglomerate channels, mudcracks, shrinkage polygons, concretions up to 3 cm in diameter, and raindrop impressions.

The rock type and structures of this unit are consistent with an alluvial fan deposit (Reineck and Singh, 1975) on the slope of a volcanic edifice. The scour and fill, current ripple crossbedding and lamination, and gravel lag deposits are characteristic of braided channels on an alluvial fan. The massive conglomerates and those with imbricated clasts are probably mudflow deposits. Upward coarsening of the sequence represents either progradation of the fan or migration of channels across the fan.

Unit Dgs

Unit Dgs is a green sandstone. The basal part of this unit is exposed in the western portion of the map area. In the map area the unit is thickly bedded. Beds are massive or vaguely parallel laminated. This unit was interpreted to be a fluvial deposit (V.K. Fay, pers. comm., 1987).

GOLD ASSAYS

Gold values were determined for 87 of the samples. The samples analyzed were those used for whole-rock major and trace element geochemistry, and hence are the freshest and most homogeneous rocks. Most samples have gold values < 1 ppb, with rare values up to 19 ppb, suggesting that the area is either unimportant for gold mineralization or that the wrong samples were analyzed.

SUMMARY

The map area is characterized by alternating mafic and felsic volcanism, although the felsic rocks are volumetrically more important. Several types of eruptive systems are represented. Hawaiian and Strombolian volcanism is indicated by the scoria cone (Dms) and mafic flow deposits (Dmf1 and Dmf2). More explosive Plinian or Sub-Plinian eruptions are indicated by some of the felsic pyroclastic deposits such as units Dlt2, Dlt3, Dlb1, Dlt5, Dlt6, Dwt1, and Dwt2 and associated rhyolitic flows (Drf1 and Drf2). Phreatomagmatic (Vulcanian) volcanism is suggested for some of the felsic lapilli-tuff units which contain accretionary lapilli (Dlt1 and parts of Dlt3) amd the mafic tuff breccias (Dtb2 and Dtb3).

There is evidence for most types of pyroclastic deposits including air-fall, pyroclastic flows and ash cloud and ground surge deposits. Both welded and nonwelded varieties occur in the area and there is some evidence for rheomorphism. In the lower part of the section, mafic flows or shallow-level sills show evidence for interaction with wet sediment producing a sequence of peperitic breccias.

The shale, sandstone and mudstone units, Drs1, Drs2, and Dvc all are tidal flat to beach facies rocks. Unit Dvc has a large component of volcanic clasts indicating nearby mafic Strombolian-type volcanism. The uppermost sedimentary rock units, Drs3 and Dgs are fluvial deposits. Sedimentation is quickly established during brief intervals of volcanic quiescence.

The volcanism was subaerial, but near sea level (littoral) near the base of the section, with elevation increasing upward. This is suggested by the upward change is the nature of sedimentation from littoral to fluvial. There is evidence for several small volcanic centres. A southern source area is suggested for the felsic pyroclastic units Dwt1, Dlt3 and Dtb3, whereas, the source for the mafic flows of unit Dmf2 is apparently to the north. Other units, such as Dms, Dtb2 and Dtb3, have a limited aerial extent and are apparently near their source areas.

Upper levels of a volcanic belt are indicated by the paucity of reworked volcanic deposits and courseness of the tuff breccias. Although not described here, the intrusions commonly have a fine grain size and are relatively vesicular which is consistent with occurrence in the upper levels of a volcanic belt.

Sulphide mineralization is primarily related to mafic volcanic breccias deposited near sea level. Gold values are low.

ACKNOWLEDGMENTS

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Paleomagnetic investigations on the Iles de la Madeleine, Gulf of St. Lawrence

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Abstract

A paleomagnetic study of the Lower Permian Cap aux Meules Formation on the Iles de la Madeleine reveals a history of tectonic motion, recorded by hematite during deposition and diagenesis. An average of results yields a mean pole (129E, 45N) in agreement with the Permo-Carboniferous (Kiaman) magnetic signature of North America. However, pole positions from individual sites vary in longitude. The pattern may have resulted from a rotation of a block during remanence acquisition. The sense of rotation is suggested by mineralogy. A westward shift in the pole correlates with increasing alteration of the redbeds. The oldest poles are to the east, suggesting a counterclockwise rotation of the Iles de la Madeleine during the Permian.

Résumé

Une étude paléomagnétique de la formation de Cap aux Meules, d'âge permien inférieur et située dans les îles de la Madeleine, révèle une série d'épisodes de mouvements tectoniques, enregistrés par l'hématite durant sa sédimentation et sa diagénèse. En faisant la moyenne des résultats, on obtient un pôle moyen (129°E, 45°N) qui concorde avec la signature magnétique permo-carbonifère (Kiaman) de l'Amérique du Nord. Cependant, après avoir déterminé la position des pôles, on note que ceux-ci ont une longitude différente d'un site à l'autre. Il est possible que cette différence résulte de la rotation d'un bloc au moment où était acquise la rémanence. Le sens de la rotation est indiqué par la minéralogie. Le déplacement vers l'ouest des pôles présente une corrélation avec le degré croissant d'altération des couches rouges. Les pôles les plus anciens se trouvent à l'est, ce qui indique une rotation des îles de la Madeleine en sens inverse des aiguilles d'une montre durant le Permien.

GEOLOGICAL BACKGROUND

The youngest sedimentary rocks exposed on the Îles de la Madeleine are eolian redbeds of the Cap aux Meules Formation. D. Brisebois (1981) placed this formation in the Lower Permian, stratigraphically above the Pictou Group, or correlative with its uppermost part. The islands were formed by localized uplift on normal faults, which was associated with rising salt domes. Gravity and seismic data indicate an abundance of low density evaporite structures in the Magdalen Basin (Watts, 1972). An exploratory well drilled in 1975 by INRS-PETROLE on Île Brion encountered a thick salt member of Windsor age. The movement of this salt formed the horsts responsible for the development of the Îles de la Madeleine. The main episode of faulting closely followed the deposition of the redbeds.

The Cap aux Meules Formation is composed of two members: Etang du Nord and Etang des Caps. Both are eolian sandstones. The lower Etang du Nord member contains abundant lacustrine intercalations. The upper Etang des Caps member is characterized by a uniform grain size, and giant crossbedding. The paleoenvironment of the Cap aux Meules Formation (particularly the upper member), is a continental desert.

METHOD

A stratigraphically representative collection of samples for paleomagnetic work was drilled from both members of the Cap aux Meules Formation (Fig. 1, Table 1). The collection was obtained at 21 sites with an average of 7 cores per site.

Approximately half of the collection was subjected to thermal treatment in an electric furnace (Roy et al., 1972). The other half was treated in an alternating field (AF) demagnetizer (Roy et al., 1973). Chemical treatment in 5N HCl was attempted on several samples. The rocks are poorly consolidated, and samples disintegrated in acid before any

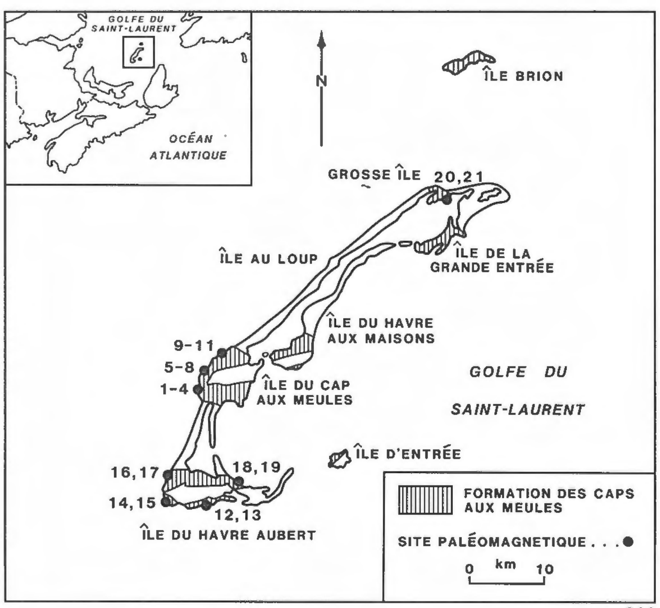


Figure 1. Paleomagnetic sampling localities of the Cap aux Meules Formation. Map adapted from Brisebois (1981).

changes in the direction or intensity of the remanence vector could occur. Some samples also disintegrated under high temperature treatment (above 640C). Nevertheless, for most of the collection, an unblocking temperature (T_{ub}) of 670C or higher was reached. In isolated cases T_{ub} was above 715C (instrument limit)*. Such high T_{ub} accompanied by low intensities of NRM (in the order of 0.01 Amps/metre) indicates that the magnetic signal in these rocks is due exclusively to fine grained hematite. The remanence was also very stable under AF treatment, and in many cases the alternating field stability (afs) exceeded the instrument limit (280 millitesla).

PALEOMAGNETIC RESULTS

A reversed polarity period of considerable duration, the Kiaman Magnetic Interval (Roy and Morris, 1983), has been recognized worldwide for rocks of Middle Pennsylvanian to Permian age. The Cap aux Meules Formation is predominatly reversely polarized. Within-site averages were calculated (Fig. 2), and most of the sites (excluding 9, 10, 20, and 21) were combined into an overall mean, expressed as pole 129E, 45N (Fig. 2). This pole agrees statistically with Upper Carboniferous or Lower Permian reference poles for North America (Roy et al., 1983).

However, several factors indicate that the mean Cap aux Meules pole does not represent a single remanence acquisition event in the Lower Permian. These factors are:

1. The occurrence of normal polarities in some sites. This suggests that a post-Kiaman overprint affected some locatities. A pre-Kiaman original signature is ruled out because the Cap aux Meules Formation is stratigraphically far above the paleomagnetic horizon marker, which identifies the beginning of the Kiaman Interval (Roy and Morris, 1983).

Table 1. Description of sampling in Cap aux Meules Formation

Stratigraphic sequence of paleomagnetic sites		Colour of hand specimen	Type of hematite in thin sections
Member	Site No.		
	16, 17	Brick red	Detritus and overgrowths
Etang des Caps	9-11	Pink to yellow	Overgrowths and recrystallized hematite
	5-8	Yellow to red	Limited detritus, pervasive overgrowths
	14-15	Brick red	Abundant detritus, some overgrowths
	12,13	Light pink	Detritus and overgrowths
Etang du Nord	20,21	Bright pink to yellow	Recrystallized hematite, minor overgrowths
	1-4	Light pink (Site 2-Brick red)	Minor detritus, overgrowths (Site 2 - more detritus)
	18,19	Pink to brick red	Detritus, minor overgrowths

* The cause of such high observed T_{ub} of the Cap au Meules hematite is still under investigation.

2. Two-component remanences isolated from single samples. Many samples from sites 4, 12 and 13 exibit two distinct remanences, or a trend from one to the other, under AF or thermal treatment. Because both remanences reside in hematite they may also be present, but impossible to separate, in other sites due to similarities in T_{ub}, a.f.s., and intensity.

3. Longitudinal distribution of pole averages from individual sites. A noticeable spread in the longitude of the Cap aux Meules poles about a relatively consistent latitude (ca. 43N), is evident in Figure 2. A systematic relationship was found between variations in the mineralogy of hematite and different pole positions (Fig. 3). It appears that a migration of the pole from east to west may record continuous diagenetic changes.

4. Recent overprint. The scattered results from sites 9, 10, 20 and 21 can be explained as a result of the superposition of an overprint by the recent field on the existing Upper Paleozoic remanence. These sites are notably rich in recrystallized secondary hematite.

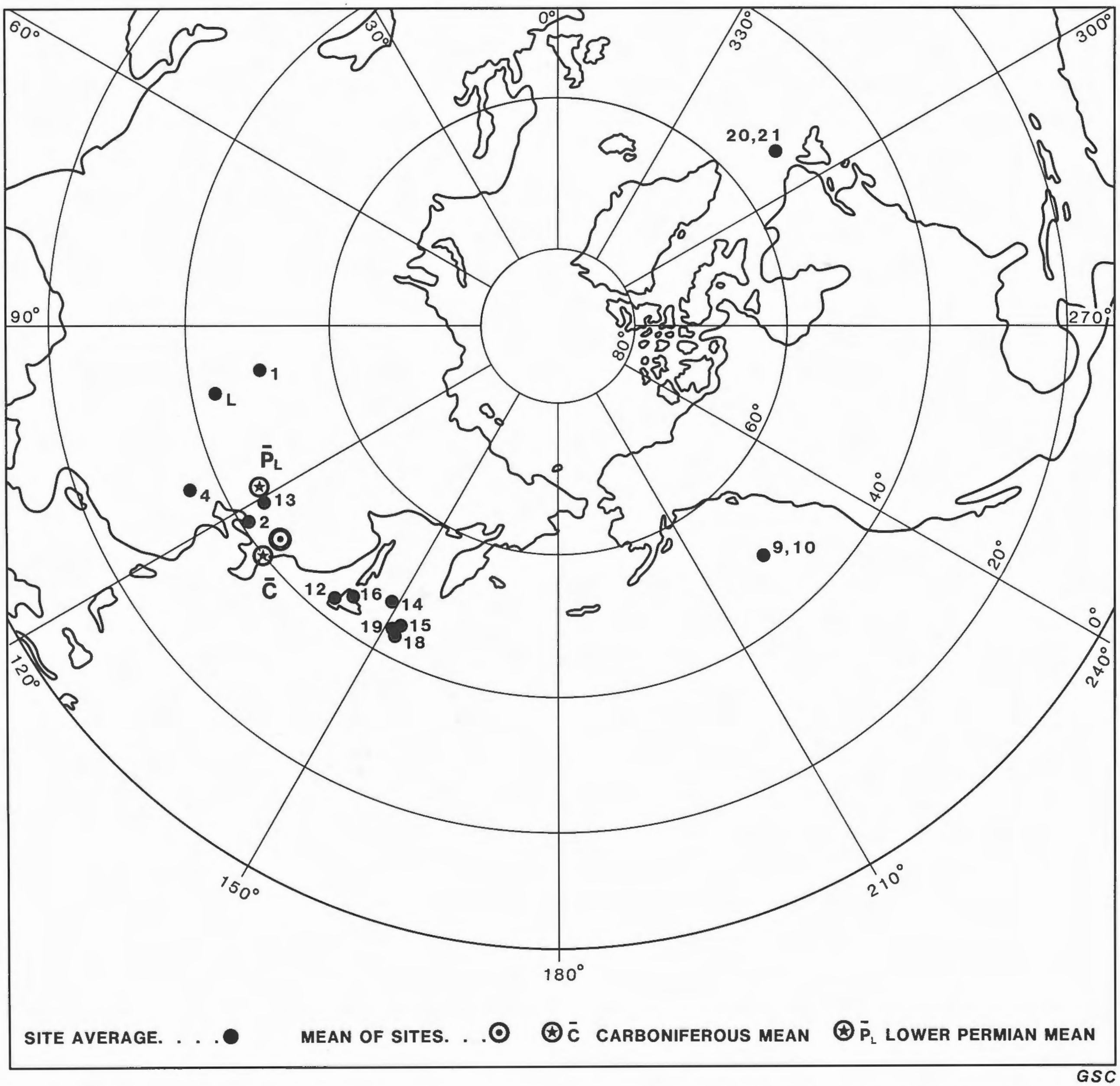


Figure 2. Paleomagnetic site poles. L = average of leached sites 3, 5-8 and 11, which had few specimens and could not be averaged individually. Carboniferous and Lower Permian means from Roy et al., (1983).

INTERPRETATION

The comparison of a given site's pole position to the type of hematite present suggests that the westward migration of the pole correlates with increasing alteration of the redbeds. Figure 3 is a schematic representation of this correlation.

Primary remanent magnetization

Detrital to early diagenetic features can be seen in the hematite responsible for poles that fall on the eastern end of the observed longitudinal distribution. In outcrop these rocks are brick red, and in thin section are seen to be rich in opaques. Figure 4a shows rounded hematite grains in crossbedded bands. These grains may have been deposited as magnetite, and subsequently oxidized. Therefore, the term "primary remanence" applied to the magnetic signature of this type of hematite includes early diagenesis.

Figure 5 shows a typical primary remanence from site 14. Unblocking of a stable reversed remanence occurs in the 670-680C range. A straight line, terminating at the origin, can be approximated through both vector diagrams. This feature suggests that a single component magnetization is present.

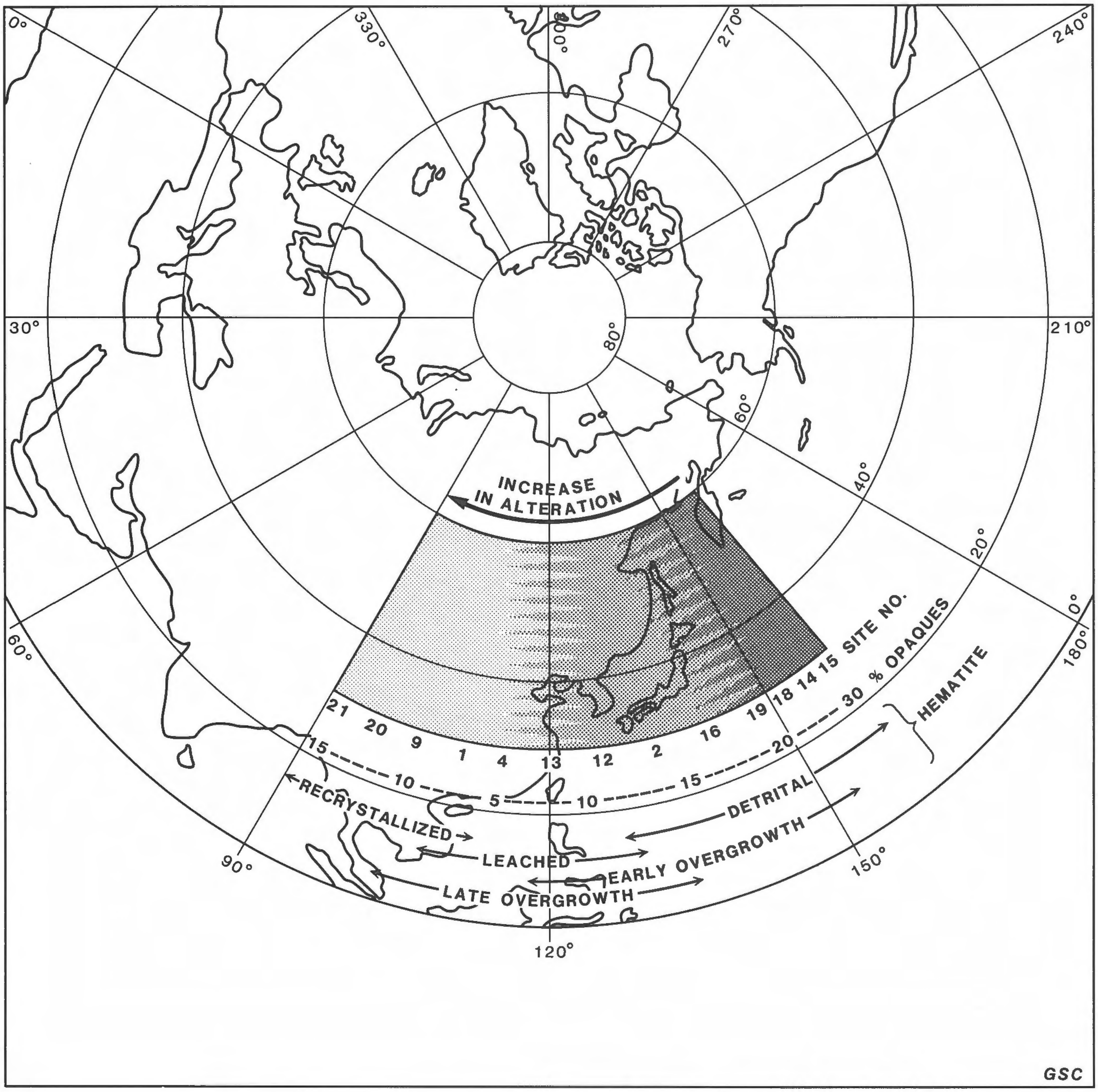
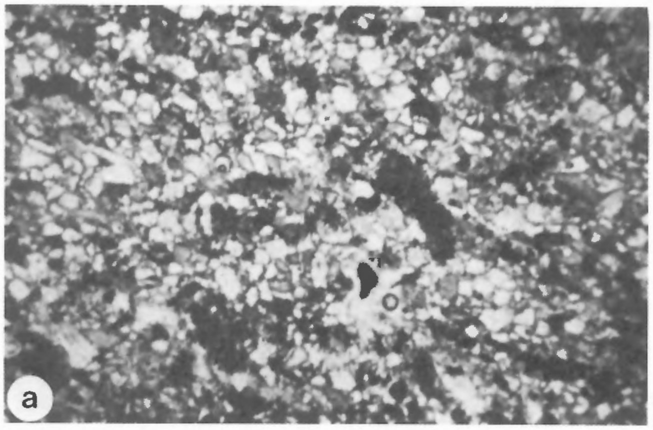


Figure 3. Distribution of poles as a function of hematite mineralogy.



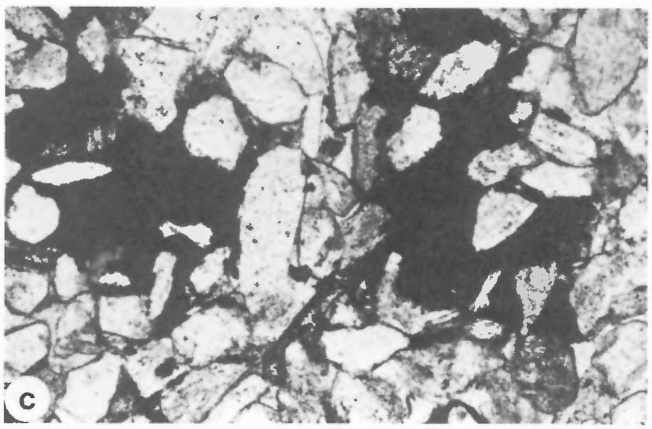


Figure 4a. Detrital hematite in cross-bedding. Site 14, 2.5X. b) Remobilized hematite in the form of late stage overgrowths. Site 6, 10X. c) Recrystallized hematite of recent origin. Site 21, 10X.

Intermediate and secondary remanent magnetization

The westward migration of the site poles correlates with an increasing amount of alteration observed in thin sections (Fig. 4b). Recognizably detrital features decrease at the expense of hematitic overgrowths, and an overall decrease in opaques is evident, indicating that hematite has been removed from the system. The rocks are lightest in colour near the middle of the spectrum (sites 12 and 13), and composite directions encountered here can be related to incomplete remobilization of hematite. This remobilization was caused by leaching solutions moving along faults that cut the Cap aux Meules Formation soon after deposition. In the field, many fault zones are visible as green areas, devoid of hematite.

The secondary remanent magnetization, exemplified by site 1 in Figure 6, produces a pole in the western end of the longitudinal distribution, where the greatest amount of leaching was observed. The unblocking temperature is slightly higher (705C) than that of the primary remanence, possibly due to the finer grain size of remobilized hematite. Most of the unblocking spectrum appears as a straight line in vector diagrams; however, minor directional changes at low temperatures suggest the presence of more than one component.

Recent overprints

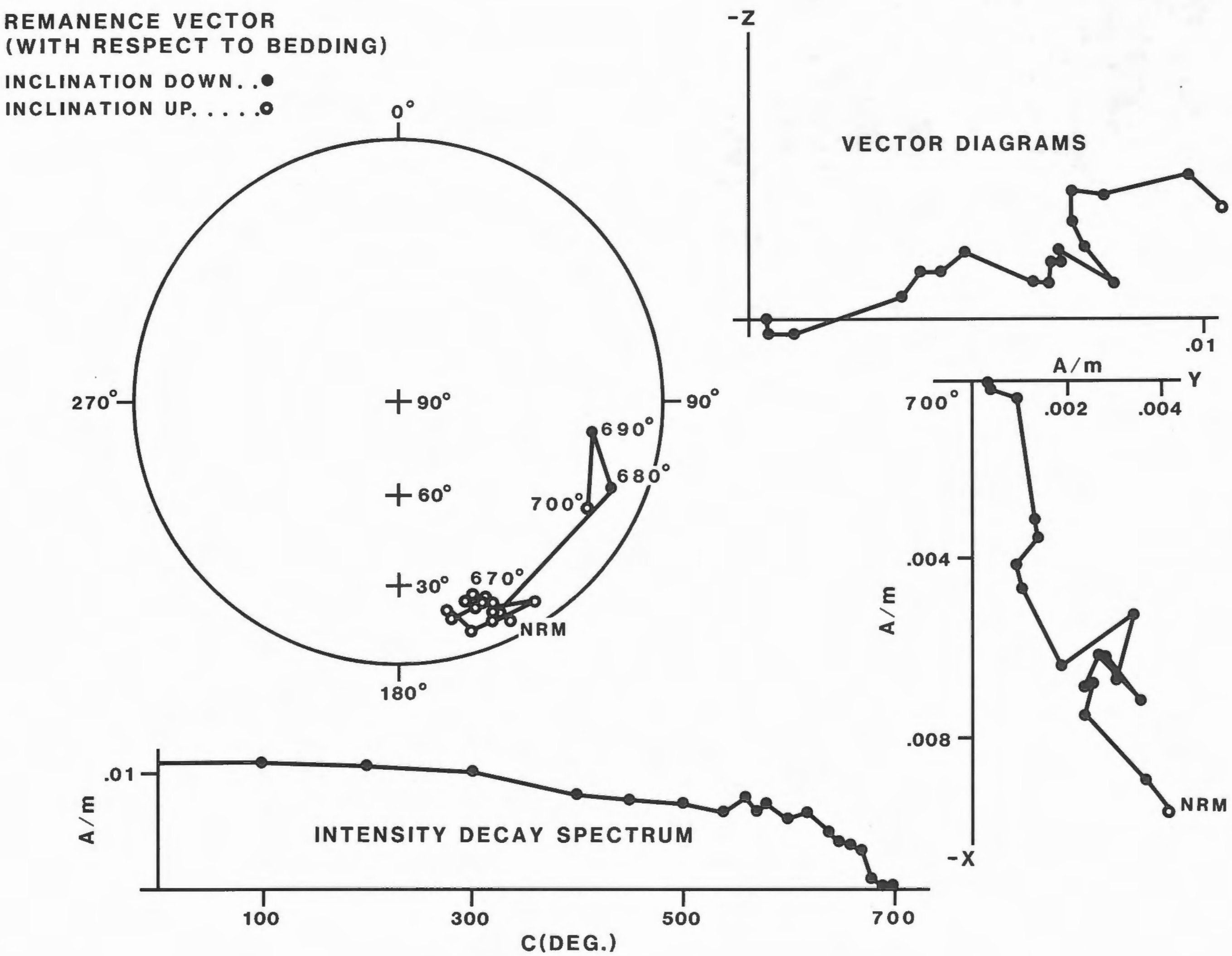
The predominant reversed polarity of the primary and secondary remanences (Fig. 5, 6), and intermediate pole positions given by other sites, suggest that the process responsible for the longitudinal distribution of poles occurred during the Kiaman Interval. Results from sites 9, 10, 20 and 21 do not fall on the longitudinal distribution (Fig. 2), but were included on Figure 3 for completeness, to demonstrate the trend in hematite mineralogy. A rise in opaques is evident in these sites. Recrystallized hematite (Fig. 4c) that gives rise to bright pink and yellow hand specimens, is responsible for scattered remanence directions of normal polarity. Figure 7 shows a steep positive remanence which is not randomized in alternating fields of 280 millitesla. The demagnetization trend as seen in vector diagrams does not approach the origin, indicating the presence of more than one component, that could not be separated. The scatter of these sites can be explained by introduction of hematite into the Cap aux Meules Formation in recent times, resulting in a pervasive overprint, that has obscured older remanences.

Age control

The supposition, that reversed polarities place the main remanence acquisition event in the Lower Permian, is substantiated by a test using bedding corrections. All remanence vectors presented in this study are corrected for the attitude of the bedding at the sampling locality. In most cases the deviation from the horizontal was not sufficiently great for the bedding correction to alter the results significantly. At sites 18 and 19 the attitude of the bedding was substantially disturbed by faulting. It was possible to reconstruct the faulted block to its original orientation. The bedding-corrected results from this block are in good agreement with the primary remanence (Fig. 8) indicating that the remanence here was acquired prior to faulting.

Table 2. Average remanence vectors and paleomagnetic poles from the Cap aux Meules Formation (α_{95} = radius of circle of confidence; δp, δm = semi-axes of the oval of confidence at 0.05 probability level; average "L" as defined in Fig. 2).

Remanence	Sites	Vector corrected for bedding declination	inclination	α_{95}	Paleomagnetic pole Longitude	Latitude	δp	δm
Primary	14,15,18,19 (N=4)	157°	-14°	3.5°	151°E	45°N	1.8°	3.6°
Transitional	2,12,13,16 (N=4)	170°	-4°	11.4°	132°E	44°N	5.7°	11.4°
Secondary	1,4, "L" (N=3)	190°	0°	25.0°	105°E	42°N	12.5°	25.0°
Cap aux Meules Average	All above (N=11)	172°	-6°	10.6°	129°E	45°N	5.3°	10.6°



PRIMARY REMANENT MAGNETIZATION

SITE 14, NO. 144A

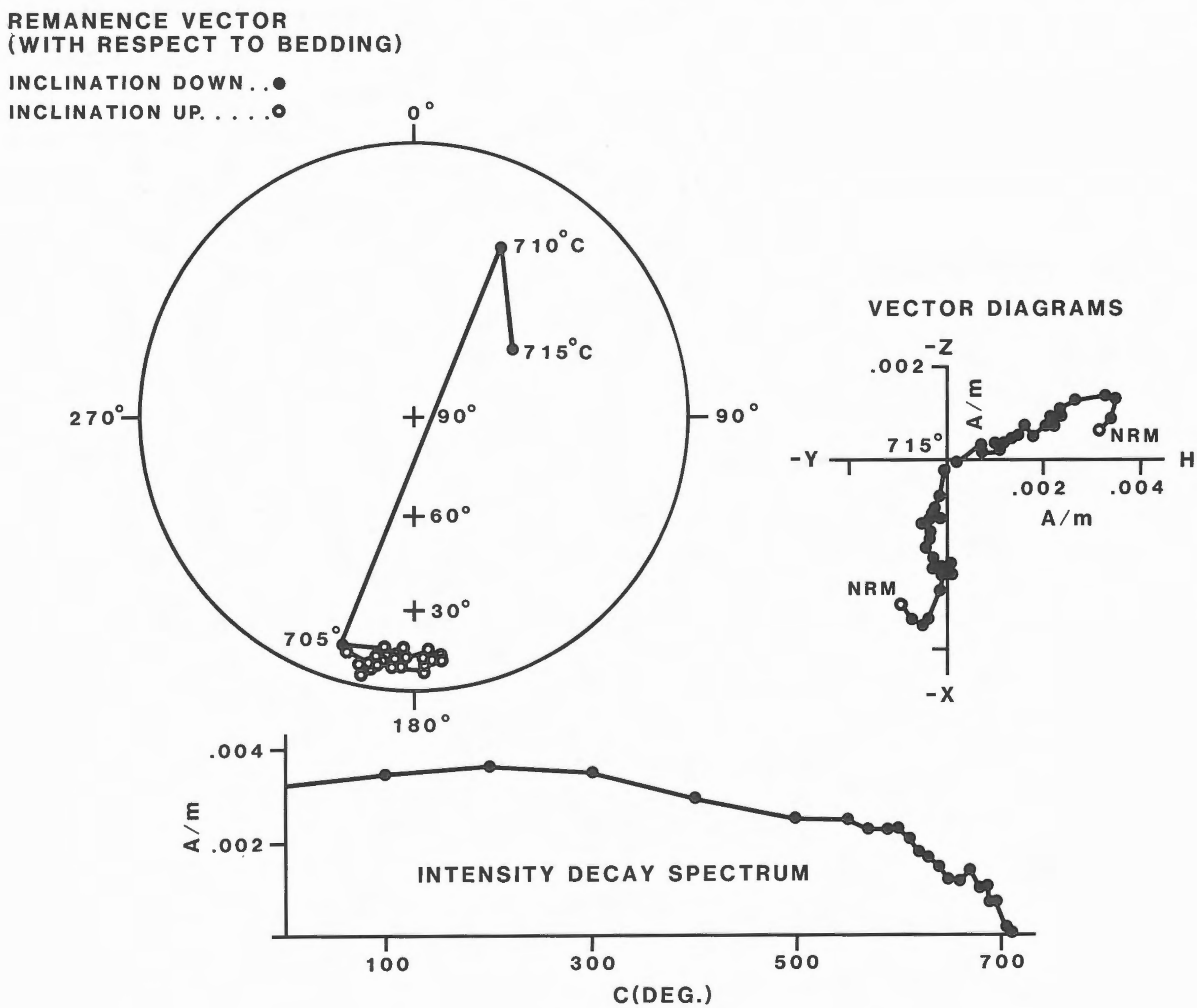
Figure 5. Paleomagnetic signature typical of detrital to early diagenetic hematite.

CONCLUSIONS

The Cap aux Meules poles form a longitudinal trend younging, on the basis of hematite mineralogy, from east to west. Three data sets were distinguished as follows:

1. Sites containing predominantly detrital to early diagenetic hematite.
2. Partially leached sites, where detrital hematite and secondary overgrowths are mixed.
3. Extensively leached sites, where secondary overgrowths predominate.

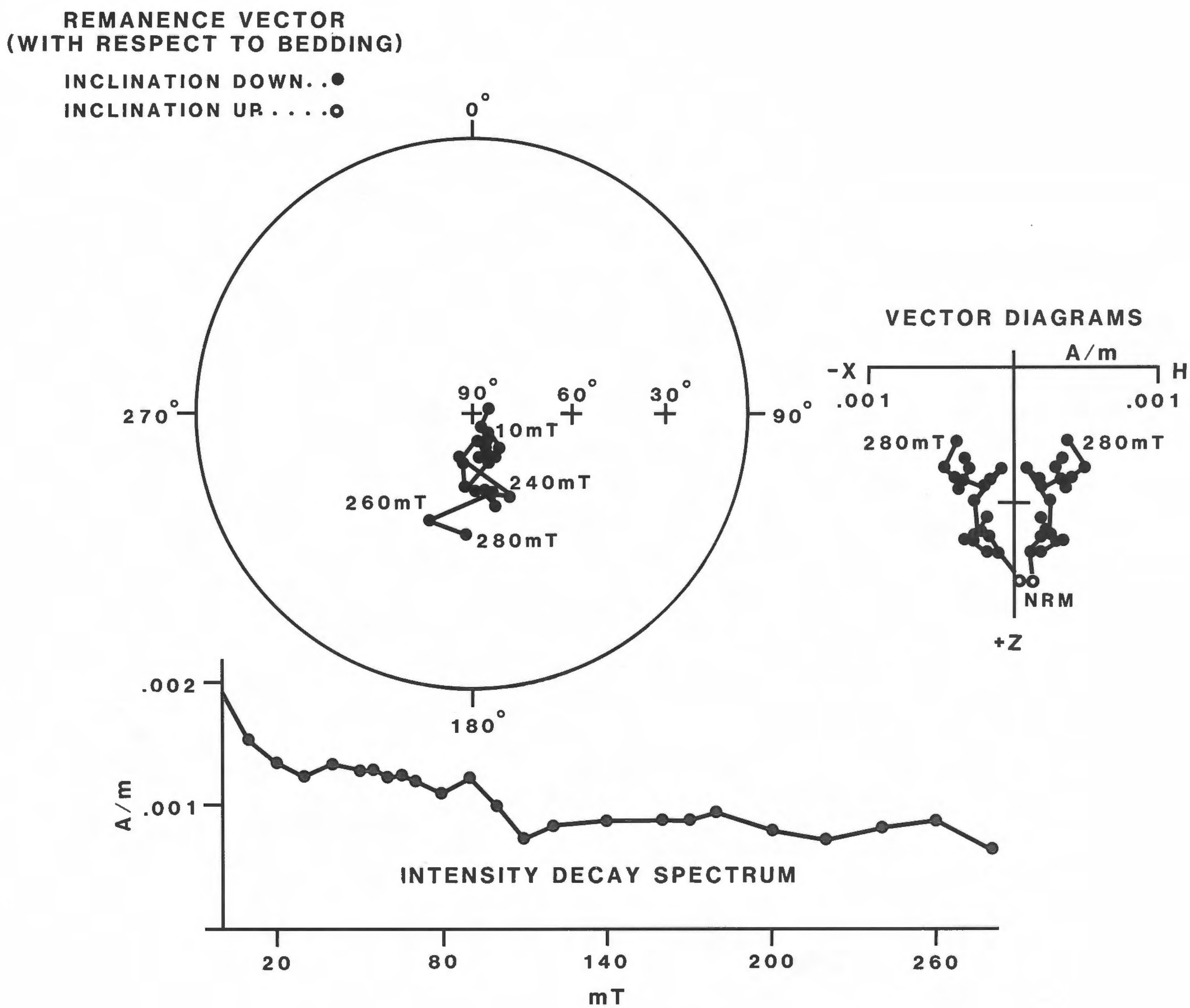
The resulting pole positions are shown in Table 2 and Figure 9. A general east to west trend from the Carboniferous to the Triassic is visible in the North American paleomagnetic results summarized by Irving and Irving (1982), and shown for reference in Figure 9. The westward migration of the Cap aux Meules pole can be produced by a counterclockwise rotation (increase in declination of the remanence vector through time, with no change in inclination). Figure 9 shows that the primary Cap aux Meules pole is significantly east of the North American polar wander curve of Irving and Irving (1982). The transitional and secondary pole positions are statistically the same as those from elsewhere in North America. It follows that crustal motion in the Magdalen Basin was more pronounced than that of the North American craton. Counterclockwise rotation with respect to the craton appears to have occurred during the Kiaman Interval. Salt horizons present at depth in the basin may have acted as zones of weakness, and locally enhanced tectonic movements.



SECONDARY REMANENT MAGNETIZATION

SITE 1, NO. 015B

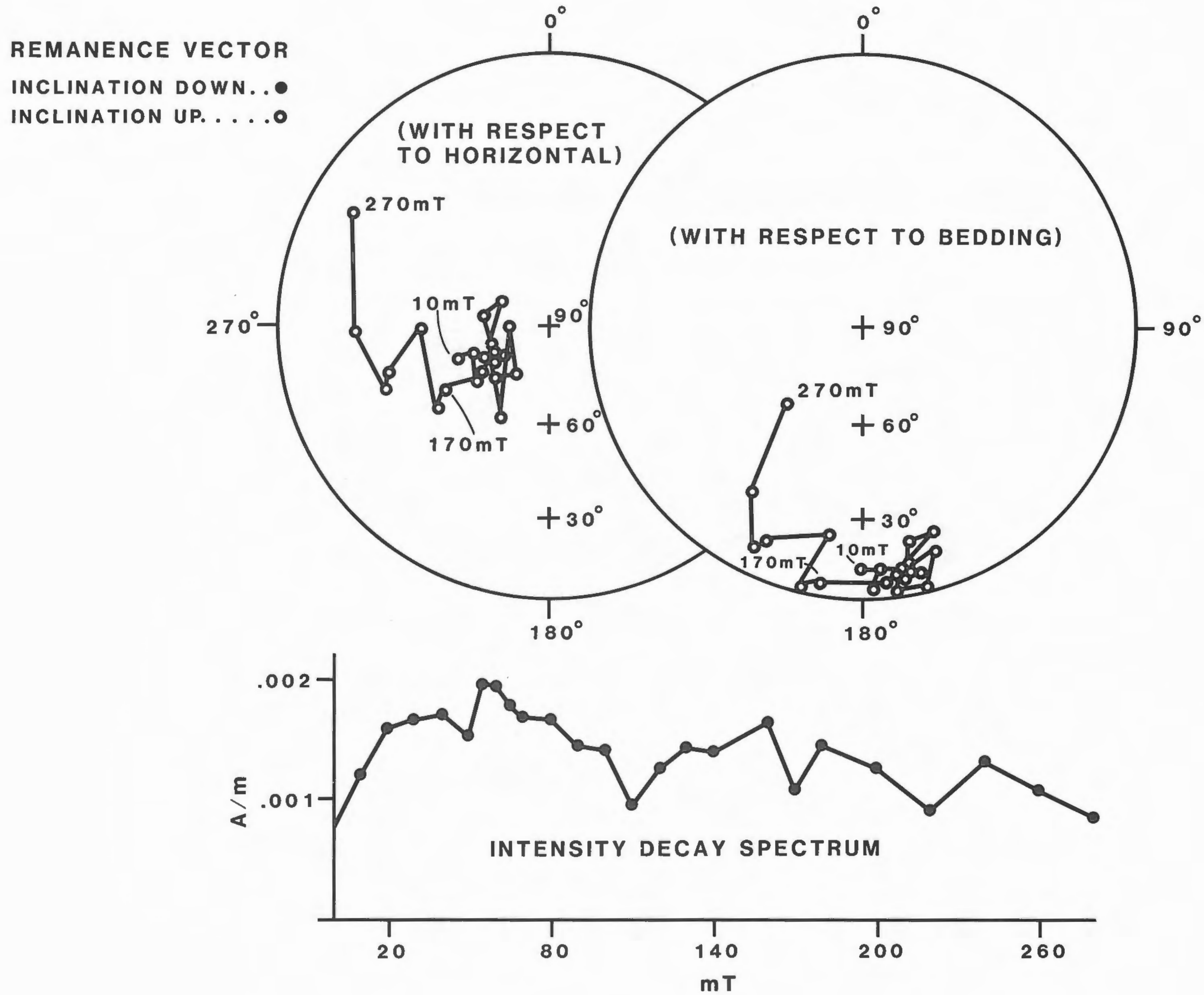
Figure 6. Paleomagnetic signature typical of late diagenetic (leached) hematite.



RECENT OVERPRINT

SITE, 20, NO. 208B

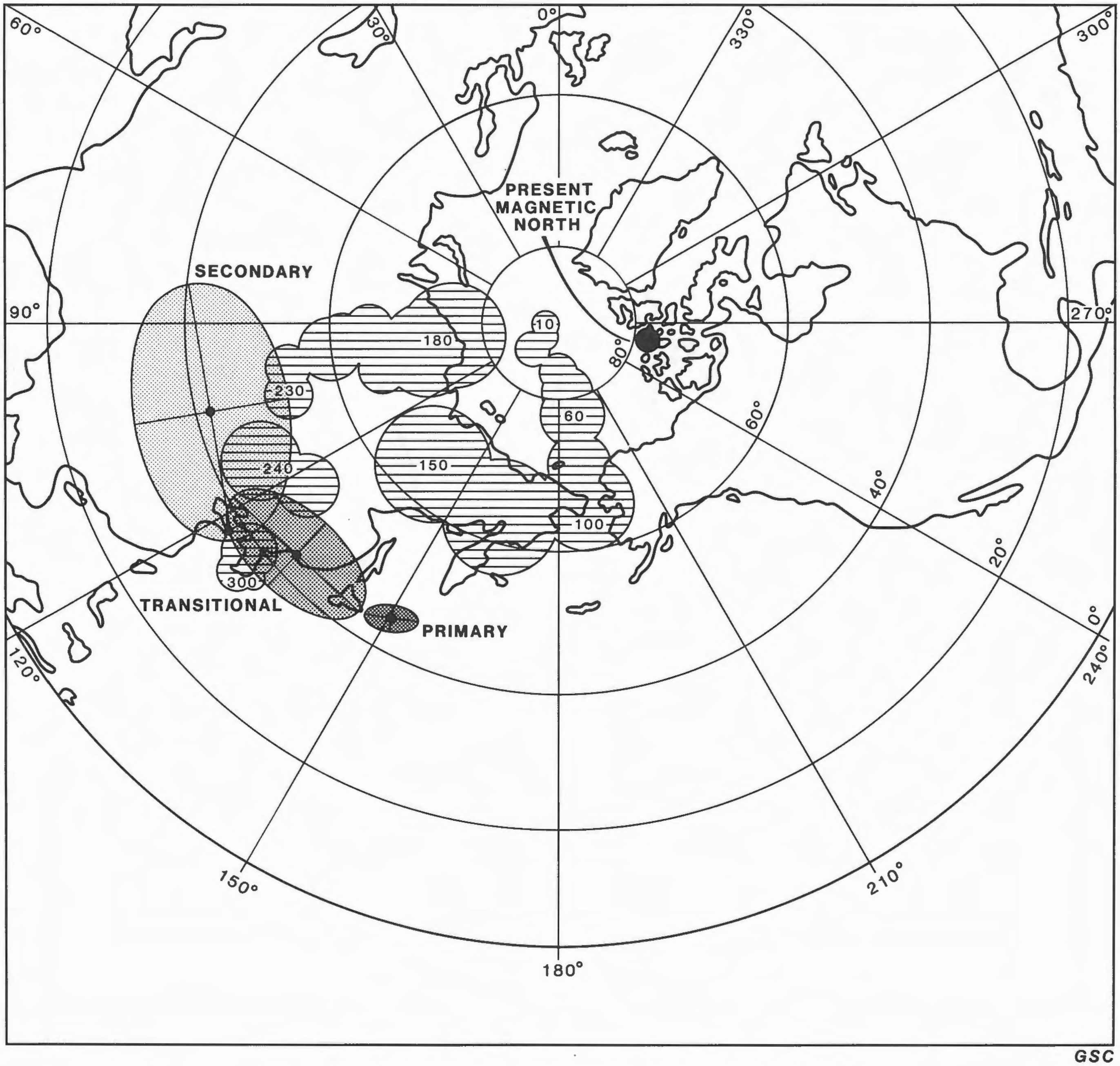
Figure 7. Paleomagnetic signature typical of sites containing recrystallized hematite.



BEDDING CORRECTION

SITE 18, NO. 185B

Figure 8. Age constraint of primary remanence by bedding correction.



NORTH AMERICAN MEAN PALEOPOLES
20Ma WINDOW (AFTER IRVING & IRVING, 1982)

Figure 9. Counterclockwise rotation of the Cap aux Meules pole with respect to the North American apparent polar wander path of Irving and Irving (1982). Ages in Ma are shown on the path.

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Tectonic-stratigraphic subdivisions of central Newfoundland[1]

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Abstract

A major two-fold division of the Dunnage Zone and redefinition of the Gander Zone are proposed for central Newfoundland.

The Dunnage Zone is bisected into a northwestern Notre Dame Subzone and southeastern Exploits Subzone by the Red Indian Line. It is a faulted boundary separating Ordovician volcanic-sedimentary rocks that exhibit a variety of geological contrasts.

Three isolated areas of metaclastic rocks with little or no volcanic material are assigned to the Gander Zone. These define the Gander Lake, Mount Cormack and Meelpaeg subzones. The metaclastic rocks are in places surrounded by rocks of the Exploits Subzone and they coincide with metamorphic and plutonic culminations. Boundaries are demonstrable faults in most places.

The Red Indian Line is interpreted as a major terrane boundary within the Newfoundland Dunnage Zone. Rocks of the Mount Cormack and Meelpaeg subzones may be Gander Lake inliers, occurring as structural windows through the Exploits Subzone.

Résumé

On a proposé, pour le centre de Terre-Neuve, la subdivision de la zone de Dunnage en deux grandes parties, et une nouvelle définition de la zone de Gander.

La zone de Dunnage est partagée également en la sous-zone nord-ouest de Notre-Dame et la sous-zone sud-est d'Exploits, par la ligne de Red Indian. Il s'agit d'une limite créée par une faille séparant des roches qui montrent une multitude de contrastes géologiques.

On a attribué à la zone de Gander trois régions isolées de roches métaclastiques contenant peu ou pas de produits volcaniques. Ces régions définissent les sous-zones de Gander Lake, de Mount Cormack et de Meelpaeg. Les roches métaclastiques sont par endroits au voisinage de roches appartenant à la sous-zone d'Exploits, et coïncident avec des culminaisons métamorphiques et plutoniques. Les limites sont des failles dont la présence est démontrable dans la plupart des endroits.

On a interprété la limite de Red Indian comme étant la limite d'un grand terrane, à l'intérieur de la zone de Dunnage à Terre-Neuve. Les roches des sous-zones de Mount Cormack et de Meelpaeg sont peut-être des enclaves de la formation de Gander Lake, qui se présentent sous forme de fenêtres structurales dans toute la sous-zone d'Exploits.

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INTRODUCTION

Newfoundland has a long history of tectonic-stratigraphic zonal subdivisions because its early Paleozoic rocks are so variable and distinctive across the well-exposed northeast coastal section. The spatial divisions are useful, both for an understanding of the orogen and a practical framework for synthesis and presentation of information. The subdivision of widest usage is that of the Humber, Dunnage, Gander and Avalon zones from west to east across the orogen (Williams, 1979). The Humber Zone is the Appalachian miogeocline. Outboard zones are suspect terranes (Williams and Hatcher, 1983) or composite suspect terranes (Keppie, 1985). The interpretation of some tectonic-stratigraphic zones as suspect terranes enhances their importance. The more descriptive terms, zone and subzone, are recommended where the nature of boundaries is controversial.

Central Newfoundland has the widest area of well-preserved Ordovician volcanic and sedimentary rocks in the internal domain of the Appalachian Orogen. It comprises the Dunnage Zone, noted for its Ordovician volcanic rocks and ophiolite suites, and the Gander Zone, defined on its Ordovician and earlier(?) metaclastic rocks. The Dunnage Zone is composite and amenable to internal tectonic-stratigraphic subdivision. Only a major two-fold division is of concern here. Rocks of the Gander Zone are monotonous and the zone is redefined and expanded to include discrete areas of metaclastic rocks outside the original type area at Gander Lake.

Much new mapping in central Newfoundland has been conducted over the last 10 years by the Newfoundland Department of Mines. This is augmented by geochemical, metallogenic, geochronological, paleontological and geophysical studies. In light of a greatly expanded data base, it is timely to reassess subdivisions of the central Newfoundland Appalachians. Meaningful subdivisions always provide an impetus to research and lead to better conceptual models.

The first author commenced a project in 1987 to investigate major zone boundaries in central Newfoundland in co-operation with his co-authors and other geologists of the Newfoundland Department of Mines. The project is ongoing and conducted in anticipation of onland Lithoprobe deep seismic experiments across central Newfoundland.

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CENTRAL NEWFOUNDLAND SUBDIVISIONS AND BOUNDARIES

The first order tectonic-stratigraphic divisions of central Newfoundland are the Dunnage and Gander zones. The boundary of the Dunnage Zone with the Humber Zone or Appalachian miogeocline to the west is the Baie Verte-Brompton Line (Williams and St. Julien, 1982). It is traceable from Baie Verte to the southern end of Grand Lake (Fig. 1). Farther south, relationships are confused with the likelihood of structural co-mingling of Humber and Dunnage rocks, all cut by numerous intrusions and affected by high grade regional metamorphism (Herd and Dunning, 1979; Chorlton, 1984; Dunning and Chorlton, 1985; van Berkel et al. 1986). In this region, the Humber-Dunnage boundary is placed for convenience at the Long Range Fault. The boundary between the Gander Zone and the Avalon Zone is the Dover Fault in the north (Blackwood and Kennedy, 1975), and the Hermitage Bay Fault in the south (Blackwood and O'Driscoll, 1976). Both zones are cut indiscriminately by a large granite batholith in central Newfoundland.

The Dunnage Zone is subdivided into a northwestern Notre Dame Subzone and southeastern Exploits Subzone based on a variety of geological contrasts between rock groups. The Notre Dame-Exploits boundary is the Red Indian Line, named after Red Indian Lake that lies on its inland course. The names Exploits and Notre Dame were used in a previous subdivision (Williams et al. 1974), but the divisions proposed here are new. Rocks and faunas of Exploits affinity also occur in a small area at Indian Bay Big Pond in the northern Gander Zone (Wonderley and Neuman, 1984).

The Gander Zone is redefined and expanded to include rocks of three discrete areas; Gander Lake, Mount Cormack and Lake Meelpaeg. Each is regarded as a subzone of the Gander Zone. All exhibit contrasts with bordering rocks of the Dunnage Zone.

Recent studies along the south coast of Newfoundland between La Poile Bay and Grey River (Blackwood, 1985; O'Brien et al. 1986; O'Brien, 1987) indicate rocks and structures that are atypical of both the Gander and Dunnage zones. Another zone or subzone is anticipated in this area pending further work. Metamorphic rocks on the southeast side of the Cape Ray Fault may belong to the Meelpaeg Subzone or else represent high grade equivalents of rocks of the Exploits Subzone.

Dunnage Zone subdivisions

The Notre Dame and Exploits subzones are outlined in Figure 1. The Notre Dame Subzone is bordered by the Humber Zone to the west and the Exploits Subzone to the east and southeast. The Exploits Subzone is bordered by the Notre Dame Subzone and the new Gander Lake Subzone of the Gander Zone. Much of its southern boundary is cut by granites between Burgeo and Bay d'Espoir, and this boundary is placed tentatively at the Bay d'Est Fault farther west. The Mount Cormack and Meelpaeg divisions of the Gander Zone occur within the limits of the Exploits Subzone.

The Red Indian Line, separating the Notre Dame and Exploits subzones, is a late rectilinear fault in most places. In a few places, it is marked by mylonites. Locally, it is interrupted by intrusions.

Across Notre Dame Bay, the Red Indian Line has a clear morphological expression, marked by steep valleys and long arms of the sea (e.gs, Lukes Arm, Sops Head, Tommys Arm faults). Where exposed, it is a steep brittle fault downthrown on its southern side. Inland to the southwest, the Red Indian Line follows the valley of Tommys Arm Brook and it reappears as a tectonic boundary between metavolcanic rocks (Notre Dame) and biotite grade metagreywackes (Exploits) on the eastern side of Great Gull Pond (Swinden and Sacks, 1986). Irregularities in this area are attributed to the nearby Hodges Hill batholith and the boundary is truncated by a circular pluton south of Great Gull Pond. Farther southwest toward Red Indian Lake (Kean and Jayasinghe, 1980), the boundary is unexposed. On the southeast side of Red Indian Lake it is marked by a narrow band of Silurian red sandstones.

The Red Indian Line follows the Lloyds River valley southwestward, and it separates the Annieopsquotch Complex from the Victoria Lake Group between Wigwam Brook and Burgeo Road (Dunning, 1984). A mylonite zone, developed in rocks of the Victoria Lake Group, occurs at or near the boundary on Burgeo Road. Farther southwest, the Red Indian Line merges with the Exploits-Meelpaeg boundary and the Exploits Subzone is eliminated. The line marks the faulted southeast border of Devonian sedimentary and volcanic rocks farther southwest where it continues as the Cape Ray Fault (Brown, 1977).

Rocks of the Notre Dame and Exploits subzones exhibit a variety of contrasts including the following: stratigraphy, lithology, structure, faunas, plutonism, lead isotopic signatures in mineral deposits, and geophysics.

Stratigraphy

Stratigraphy is the most obvious difference between rocks of the Notre Dame and Exploits subzones in northern localities. This is expressed by the presence of Upper Ordovician and Silurian marine greywackes and conglomerates that overlie Caradocian black shales of the Exploits Subzone, and their absence in the Notre Dame Subzone. The greywackes (Sansom Formation and equivalents) and conglomerates (Goldson Formation and equivalents) reach combined thicknesses of 3000 m and constitute a continuous conformable sequence that bridges the Ordovician-Silurian boundary. Younger Silurian rocks are redbeds and terrestrial volcanics that occur in both subzones.

Silurian olistostromes and mélanges are special rock types unique to the Exploits Subzone of Notre Dame Bay.

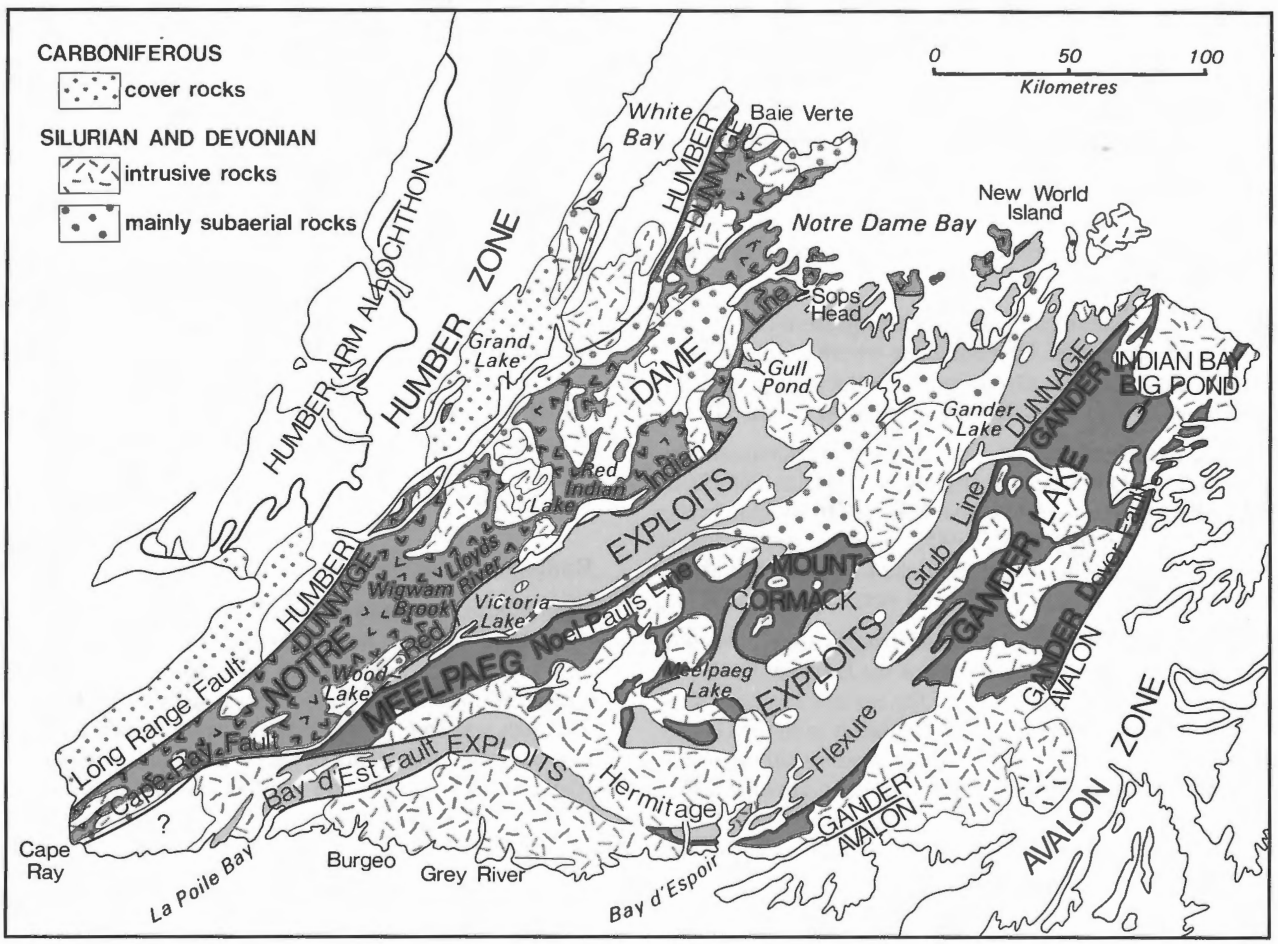

Figure 1. Tectonic-stratigraphic subdivision of central Newfoundland.

Matrix shales are fossiliferous at New World Island (McKerrow and Cocks, 1978) and the chaotic rocks contain blocks of Ordovician volcanic rocks, limestones, shales and sandstones. At Sops Head, the mélange contains volcanic fragments and associated Sansom greywackes contain ophiolite detritus, derived presumably from the Notre Dame Subzone (Nelson, 1981; Nelson and Casey, 1979). The thick and extensive Dunnage Mélange (Hibbard and Williams, 1979) of eastern Notre Dame Bay is another lithological unit peculiar to the Exploits Subzone. As a broad generality, Ordovician sections of the Exploits Subzone contain more shales and other sedimentary rocks compared to Ordovician sections of the Notre Dame Subzone.

Structure

Structural contrasts between the Notre Dame and Exploits subzones compliment the stratigraphic contrasts. A sub-Silurian or sub-Devonian unconformity throughout the Notre Dame Subzone contrasts with the continuous Ordovician-Silurian sections of the northern Exploits Subzone. Sub-Silurian unconformities occur at Baie Verte Peninsula; between the Lower Ordovician Snooks Arm Group and overlying Cape St. John Group, between the Ordovician Flatwater Group and overlying Mic Mac Lake Group, and between the Burlington Granodiorite and the Mic Mac Lake Group (Hibbard, 1983). Silurian red sandstones of the Springdale Group overlie the Ordovician Lushs Bight and Roberts Arm groups and Springdale volcanic rocks and porphyries contain inclusions of underlying mafic volcanic rocks and serpentinite (Dean, 1978). Tonalite that cuts ophiolitic rocks at Glover Island of Grand Lake is nonconformably overlain by Silurian sandstones (Knapp, 1982), Silurian red sandstones and volcanics overlie the Ordovician Annieopsquotch Complex at Victoria Lake and Burgeo Road (Chandler, 1982; Chandler and Dunning, 1983), and Devonian sandstones and volcanics of the Windsor Point Group overlie Ordovician trondhjemites and gneisses along the Cape Ray Fault (Chorlton, 1984).

A sub-Silurian unconformity is also present on the west side of White Bay in the Humber Zone where Wenlock sandstones of the Sops Arm Group overlie allochthonous Ordovician rocks of the Coney Head Complex (Williams, 1977).

In the Exploits Subzone, Ordovician-Silurian unconformities are rare. One example at New World Island has Silurian Goldson conglomerate above Middle Ordovician mafic volcanic rocks (Arnott et al. 1985); another just northeast of Victoria Lake has the Rogerson Lake Conglomerate above granite porphyry (Kean, 1977; and pers. comm., 1987).

Two Middle Ordovician unconformities are recorded in central Newfoundland. One separates the Davidsville Group and underlying ophiolitic rocks of the eastern Exploits Subzone close to the northern part of its boundary with the Gander Zone (Blackwood, 1982). A similar relationship occurs at the western boundary of the Notre Dame Subzone where sedimentological evidence implies an unconformity between the Ordovician Flatwater Group and adjacent ophiolites of the Baie Verte Peninsula (Hibbard, 1983).

Along the Red Indian Line, Silurian or Devonian rocks are much less deformed on its Notre Dame side compared to Silurian rocks on its Exploits side. This is most striking at Victoria Lake and Burgeo Road where Silurian rocks above the Annieopsquotch Complex are mildly folded and lack penetrative cleavage, whereas the nearby correlative Rogerson Lake Conglomerate on the Exploits side is a phyllite with extremely flattened or attenuated clasts.

Taconic Orogeny affected all of the Notre Dame Subzone and its effects are recorded across the Humber Zone or Appalachian miogeocline. Large parts of the Exploits Subzone are unaffected by Taconic Orogeny. Acadian (Devonian) orogeny affected both subzones, and in the vicinity of the Red Indian Line it is more intense in the Exploits Subzone.

Fauna

Ordovician brachiopods and other shelly faunas from volcanic-dominated sections of interior parts of the Appalachian Orogen contrast with coeval faunas of the miogeocline and interior North America. The brachiopod faunas are assigned to the Celtic realm and thought to have evolved in isolation on ocean islands (Neuman, 1984). All of the Celtic Lower to Middle Ordovician brachiopod faunas of the Newfoundland Dunnage zone occur east of the Red Indian Line. They are commonest in the Exploits Subzone of Notre Dame Bay, but also occur in the Mount Cormack Subzone of the Gander Zone and in a possible Exploits satellite in the northern Gander Zone (Wonderley and Neuman, 1984). In contrast, a coeval Ordovician conodont fauna from the Buchans Group of the Notre Dame Subzone has North American affinities (Nowlan and Thurlow, 1984). This suggests geographical separation of the Notre Dame and Exploits subzones during the Early to Middle Ordovician.

Plutonic rocks

Plutonic rocks are distinctive in the Notre Dame and Exploits subzones. Ordovician tonalites are common throughout the Notre Dame Subzone along its full length, but rare elsewhere. Large Devonian composite batholiths with granitic phases that cut mafic phases, e.g. Mount Peyton, Hodges Hill, Fogo batholiths, are confined to the Exploits Subzone. In the Notre Dame Subzone, middle Paleozoic intrusions are large granite batholiths that are commonly alkali, e.g. the Topsails plutons (Taylor et al. 1980).

Radiogenic lead in mineral deposits

Lead isotopic signatures in volcanogenic sulphide deposits of the Exploits Subzone contrast with those of the Notre Dame Subzone. Lead in deposits of the Exploits Subzone is relatively radiogenic and plots along a steep trend between the Zartman and Doe (1981) orogene and upper crustal growth curves. Lead in deposits of the Notre Dame Subzone is relatively non-radiogenic and plots below the orogene growth curve (Swinden and Thorpe, 1984). This indicates that lead in the different subzones represents different sources. Possibly the host volcanic rocks represent entirely different tectonic-stratigraphic environments (Stephens et al., 1984; Swinden and Thorpe, 1984).

Magnetic anomalies

The Notre Dame Subzone has higher magnetic anomalies across Notre Dame Bay compared to those of the Exploits Subzone (Zietz et al., 1980). These reflect the abundance of mafic volcanic rocks in the Notre Dame Subzone. A similar pattern is evident from Victoria Lake to Cape Ray, with high magnitude, short wavelength positive anomalies in the Notre Dame Subzone (marking ophiolite complexes and mafic plutons) and lower anomalies in the Exploits Subzone (expressing mainly sedimentary rocks, granites and metamorphic rocks).

Gravity anomalies

Bouguer gravity anomalies are higher, up to 50 mGal, in the Notre Dame Subzone and lower, commonly negative, in the Exploits Subzone (Haworth et al., 1980).

GANDER ZONE SUBDIVISIONS

The Gander Zone is defined on its monotonous clastic sedimentary rocks and paucity of volcanic rocks. The type area is Gander Lake, where its rocks are faulted against adjacent rocks of the Dunnage Zone. The Gander Zone is expanded to include metaclastic rocks in the Mount Cormack and Meelpaeg Lake areas in central Newfoundland. These define the Mount Cormack and Meelpaeg subzones. The type area is named the Gander Lake Subzone. Extending the Gander Zone and defining a Gander-Dunnage boundary in central and southern Newfoundland has always been problematic.

The Gander Lake Subzone lies east of the Dunnage Zone and its rocks are traceable from northeast Newfoundland to well south of Gander Lake. The Exploits Subzone-Gander Lake Subzone boundary is the Gander River Ultrabasic Belt (Jenness, 1958) or GRUB Line (Blackwood, 1978; 1982). It is defined by ophiolite occurrences along most of its length, or by faults. Relationships are debatable along the irregular inland boundary, which may be a fault or stratigraphic contact (Blackwood and Green, 1982). The Gander Lake Subzone may continue as a narrow peripheral band around the Hermitage Flexure, but here metamorphism and plutonism have obscured the definitive characteristics of its sedimentary rocks.

The Mount Cormack Subzone is completely surrounded by rocks of the Exploits Subzone. Boundaries are faulted and locally interrupted by intrusions. Incomplete ophiolite complexes of the Exploits Subzone occur around much of the periphery of the Mount Cormack Subzone. They preserve enough stratigraphy to indicate their facing directions, which in each case is upwards and outwards from the Mount Cormack Subzone. Accordingly, the regional structure is viewed as an eroded dome with the Mount Cormack Subzone forming a structural window through the Exploits Subzone (Colman-Sadd and Swinden, 1984). This relationship is critical to interpretations of the Gander-Dunnage boundaries elsewhere.

The Meelpaeg Subzone is surrounded by rocks of the Exploits Subzone, except southwest of Victoria Lake where the Exploits Subzone is missing and the Notre Dame and Meelpaeg subzones are juxtaposed at the Red Indian Line.

The boundary between the sedimentary and volcanic rocks of the Exploits Subzone and metaclastic rocks of the Meelpaeg Subzone, Noel Pauls Line (Brown and Colman-Sadd, 1976; Colman-Sadd, 1987), is interpreted as a fault in the vicinity of Noel Pauls Brook. A small ultramafic occurrence, possibly ophiolitic, at its extreme northeast termination implies a tectonic, rather than stratigraphic boundary between rocks of the adjacent subzones. The Meelpaeg Subzone is faulted against Exploits volcanic and ophiolitic rocks south and east of Meelpaeg Lake (Colman-Sadd, 1984; 1985; Dickson and Delaney, 1984). Farther south the boundary is truncated by the North Bay granite batholith.

The boundary is well-exposed at Victoria Lake where it was investigated during 1987. There, dark rusty sedimentary rocks of the Victoria Lake Group containing numerous mafic sills and dykes (Exploits) abut light grey foliated to mylonitic granite and leucocratic garnetiferous psammitic schists (Meelpaeg). An abrupt increase in regional metamorphic grade occurs at the contact, from laminated sediments with abundant mafic intrusions of greenschist metamorphic facies to rusty garnetiferous schists and amphibolites. Cleavage and/or schistosity in the Victoria Lake Group is parallel to foliation or a mylonitic fabric in adjacent Meelpaeg granites. The contact dips steeply to moderately southeast and a steep down-dip lineation is present in rocks on both sides of the boundary. Whether the boundary between Meelpaeg granite and Exploits sedimentary rocks is a tectonized intrusive contact, or a primary tectonic boundary, depends on the interpretation of the garnetiferous psammitic schists. If they are an integral part of the Victoria Lake Group, the Exploits-Meelpaeg contact is intrusive. If they are unrelated to the Victoria Lake Group, the boundary is tectonic.

An absence of granites in the Victoria Lake Group, and intense deformation all along this straight and continuous Exploits-Meelpaeg contact, supports a primary tectonic relationship between rocks of the two subzones. From northeast to southwest along Victoria Lake, the outcrop width of the Victoria Lake Group is progressively more narrow. The Rogerson Lake Conglomerate is a conspicuous marker that is 2 km from the Exploits-Meelpaeg boundary in the northeast, only a few hundred metres from the contact at the southwest end of Victoria Lake, and apparently truncated by the faulted contact at Wood Lake. The pattern favours a tectonic Exploits-Meelpaeg boundary that acutely truncates the Victoria Lake Group southwestward along its course. Farther south, where the Victoria Lake Group is missing, the Meelpaeg-Notre Dame boundary is the Red Indian Line.

On the Burgeo Road, on the southeast boundary of the Meelpaeg Subzone, foliated to mylonitic pink granite is faulted against mildly deformed dark grey shales of the Bay du Nord Group. As at Victoria Lake, intense ductile deformation occurs on the Meelpaeg side of the boundary.

Monotonous clastic and metaclastic rocks of the Gander Lake, Mount Cormack and Meelpaeg subzones are all viewed as roughly correlative. A lack of volcanic components and apparent continental derivation for Gander Zone rocks contrast with mixed sedimentary-volcanic assemblages and volcanic derivation of Exploits Subzone rocks. However, limited paleontological data suggest similar Early to Middle Ordovician ages. Fossils from both limestone clasts and matrix in a conglomerate of the Mount Cormack Subzone are of Llandvirn-Llandeilo age and correlate with fossils from the Exploits Subzone. These and Arenigian fossils at Indian Bay Big Pond are all characteristic of the Celtic biogeographic province (Neuman, 1976; 1984; Colman-Sadd and Swinden, 1984; Boyce, 1987; Wonderley and Neuman, 1984).

High grade regional metamorphism and plutonism are as characteristic of the Gander Zone and its divisions as the rocks themselves. Regional metamorphism increases in intensity across the Gander Lake Subzone inward and away from the Exploits Zone boundary. Numerous biotite granites and garnetiferous muscovite granites that vary from massive to intensely foliated are characteristic of its higher grade inward parts. Much lower grade rocks occur on the opposite side of the Gander-Avalon boundary.

In the Mount Cormack Subzone, prograde metamorphism affected the metaclastic rocks from greenschist to amphibolite facies, and within an area of migmatization there is a further progression from the production of a granitic melt to conversion of the rocks to homogeneous granodiorite. Small garnet-muscovite granites, located on or close to the sillimanite isograd, are interpreted as the products of anatexis. In general, lower grade metamorphic rocks occur around the edge of the Mount Cormack Subzone, adjacent to the boundary faults and surrounding Exploits ophiolite complexes. High grade rocks are in two centrally located culminations (Colman-Sadd and Swinden, 1984).

Metasedimentary rocks in the northern portion of the Meelpaeg Subzone are mostly metamorphosed to sillimanite grade, although within a few kilometres of the contact with the Exploits Subzone to the northwest, metamorphism is greenschist facies. Towards the south, the sillimanite bearing rocks pass into rocks showing in situ melting, resulting in a complete range from small isolated granitic segregations to migmatitic gneisses and granites containing metasedimentary xenoliths. A large area of biotite granite in the northern part of the subzone varies from mildly foliated to intensely foliated and mylonitic, and exhibits the same structures as adjacent metasedimentary rocks.

Near Victoria Lake, the Meelpaeg Subzone consists of foliated to intensely foliated and locally mylonitic granite. Sparse occurrences of metasedimentary rocks are medium- to coarse-grained garnetiferous schist in concordant bands with foliated granite. Intensely foliated granite and coarse-grained quartz-feldspar-biotite migmatite extend from the southeast margin of Victoria Lake more than 50 km to the southwest. The southeastern portion of the subzone is mainly foliated to massive biotite granite, with local inclusions of meta-sedimentary rocks, e.g. at Buck Lake rock quarry off the Burgeo Road.

DISCUSSION AND INTERPRETATION

Tectonic-stratigraphic subdivision of central Newfoundland highlights some existing problems in regional correlation and rock relationships. A zonal subdivision has all the advantages of a suspect terrane analysis, and is applicable even where the nature of boundaries is debatable.

Rocks of the Notre Dame and Exploits subzones record an extensive history of oceanic volcanism and sedimentation. They are interpreted as oceanic suspect terranes juxtaposed at the Red Indian Line. Rocks of the Gander Lake, Mount Cormack and Meelpaeg subzones are all equated as integral parts of the Gander Zone. Their sedimentary provenance and plutonic styles favour a continental or continental-margin environment that did not experience contemporary volcanism. As well, Dunnage Zone rocks overlie an ophiolite substrate wherever original basement relationships are known. An ophiolite substrate has nowhere been demonstrated for rocks of the Gander Zone.

The ophiolite-based sequences of the Dunnage Zone are uprooted from the ocean in which they originated. It follows that they either overlie continental crust or disturbed overthickened oceanic rocks that extend to the present mantle. The northeast Newfoundland seismic line (Keen et al., 1986) and marine reflection profiles elsewhere in the Canadian Appalachians (Glen Stockmal and Francois Marillier, pers. comm., 1987) indicate that surface boundaries are displaced with respect to deep crustal boundaries. A western block (Humber) meets, or almost meets, an eastern block (Gander?) beneath the central Dunnage Zone. This implies that the western Dunnage or Notre Dame Subzone is thrust above the Humber Zone, and the eastern Dunnage or Exploits Subzone is thrust above the Gander Zone, although other models are possible.

Middle Ordovician emplacement of the Humber Arm Allochthon across the Humber Zone in western Newfoundland coincides with the absence of Caradocian and younger Ordovician rocks of the Notre Dame Subzone and a coextensive sub-Silurian unconformity. Deposition continued through this same interval in north-central parts of the Exploits Subzone, although eastern parts of the Exploits Subzone were already disturbed as indicated by a Middle Ordovician unconformity above ophiolite occurrences at the GRUB Line and development of Ordovician mélange (Pajari et al., 1979).

The simplest and most attractive model to explain Dunnage Zone-Gander Zone relationships is one involving tectonic transport of the Exploits Subzone across the Gander Lake Subzone, with the Mount Cormack and Meelpaeg subzones forming structural windows. We favour faulted boundaries between the Dunnage and Gander zones where relationships are uncertain, although the apparent conformity of a possible Exploits outlier at Indian Bay Big Pond and rocks of the Gander Lake Subzone suggests an original stratigraphic relationship between the Dunnage and Gander zones (Wonderley and Neuman, 1984; O'Neill, 1987). Structural overprinting, metamorphism and plutonism all serve to confuse original relationships.

The time of emplacement of the Exploits Subzone above the Mount Cormack and Meelpaeg subzones is difficult to ascertain. Colman-Sadd and Swinden (1984) suggested a Silurian time of emplacement in the case of the Mount Cormack Subzone because Ordovician and Silurian rocks of the Exploits Subzone are conformable and the first deformation of the Ordovician rocks appears to be the same as the first deformation of the Silurian Botwood Group (Karlstrom et al., 1982). Alternatively, an occurrence of shaly mélange below the Coy Pond Complex at the eastern edge of the Mount Cormack Subzone and a monomict ultramafic pebble conglomerate above the complex invite comparisons with Ordovician relationships along the GRUB Line.

Sedimentological data indicate a Late Ordovician to Early Silurian linkage between the Notre Dame and Exploits Subzones (Nelson, 1981; Nelson and Casey, 1979). The presence of Silurian mélanges in the Exploits Subzone implies a head-on mechanism of convergence.

At Cape Breton Island on the opposite side of the Gulf of St. Lawrence, Grenvillian basement of the Humber Zone almost abuts the Avalon Zone (Barr et al., 1987), and the Dunnage Zone is absent. Northeastward from the Cape Breton constriction in the Appalachian Orogen, the Dunnage Zone is narrow in southwest Newfoundland and wider to the northeast. Where it is breached by structural windows of the Mount Cormack and Meelpaeg subzones, the Dunnage crust is thin. Positive Bouguer gravity anomalies indicate a thicker Dunnage crust farther northeast in Newfoundland and it may extend to mantle depths at a narrow collisional zone offshore (Keen et al., 1986).

More studies are essential and necessary to decipher the full geological history of this critical area of the Appalachian Orogen.

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A preliminary report on geology of the eastern Cobequid Highlands, Nova Scotia[1]

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Abstract

The eastern Cobequid Highlands are underlain by Precambrian to early Carboniferous rocks. To the south of the Rockland Brook Fault (RBF), the oldest rocks consist of the Great Village River Gneiss overlain by quartzites and schists (Gamble Brook Formation, GBF). The present contact between the gneisses and GBF is a ductile shear zone. GBF is unconformably overlain by metabasalts and metasediments (Folly River Formation) which may be contemporaneous with the late Precambrian Jeffers Group and Warwick Mountain Formation to the north of RBF. Late Precambrian-early Cambrian intrusions occur only to the south of RBF and provide a minimum age for deformation and metamorphism of the host rocks. Late Ordovician-early Devonian rocks consist of fossiliferous sediments. Devono-Carboniferous rocks consist of siliceous sediments to the south and bimodal volcanic rocks to the north of RBF. The area was intruded by Devono-Carboniferous plutons and deformed by Namurian folds and thrusts associated with dextral motion on major east-west faults.

Résumé

Des roches datant du Précambrien au Carbonifère inférieur occupent le sous-sol de l'est des hautes-terres de Cobequid. Au sud de la faille de Rockland Brook (RBF), les roches les plus anciennes se composent du gneiss de Great Village River recouvert par des quartzites et des schistes (formation de Gamble Brook, GBF). Le contact actuel entre les gneiss et la GBF est une zone de cisaillement ductile. La GBF repose en discordance sous des métabasaltes et des métasédiments (formation de Folly River), qui sont peut-être contemporains du groupe de Jeffers et de la formation de Warwick Mountain, d'âge précambrien supérieur, situés au nord de la RBF. Les intrusions datant du Précambrien supérieur au Cambrien inférieur ne se manifestent qu'au sud de la RBF, et établissent l'âge minimum de la déformation et du métamorphisme des roches favorables. Les roches datant de l'Ordovicien supérieur au Dévonien inférieur se composent de sédiments fossilifères. Les roches dévoniennes et carbonifères se composent de sédiments siliceux au sud et de roches volcaniques bimodales au nord de la RBF. Cette région a été traversée par des plutons d'âge dévonien et carbonifère et déformée par des plis et charriages d'âge namurien associés à un mouvement dextre le long d'importantes failles à orientation est-ouest.

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INTRODUCTION AND GEOLOGICAL SETTING

Mapping in the eastern Cobequid Highlands was begun in summer 1987 as part of the Canada — Nova Scotia Mineral Development Agreement 1984-1989. The location of the area mapped is shown in Figure 1. The purpose of this project is to map the distribution of the pre-Carboniferous rocks and to gain an understanding of the geological history of the area.

The Cobequid Highlands are unconformably overlain to the north and east by middle to Late Carboniferous rocks (Donohoe and Wallace, 1982) and are bounded to the south by the Cobequid fault (Fig. 2). They are underlain by Proterozoic to Lower Carboniferous rocks. The Cobequid Highlands lie within the Avalon "Composite" Terrane of the Appalachians (Williams and Hatcher, 1983; Keppie, 1985). This terrane contains rocks with diverse Precambrian histories, but similar Early Paleozoic histories, suggesting that it amalgamated in the latest Precambrian. This event has been attributed to the Avalonian (e.g. O'Brien et al., 1983) or Cadomian (e.g. Keppie, 1985) orogeny.

PREVIOUS WORK

The entire Cobequid Highlands was mapped by H.V. Donohoe Jr. and P.I. Wallace in the middle to late 1970s and maps were published at a 1:50 000 scale (Donohoe and Wallace, 1982). A series of reports and field guides have been published by these authors (Donohoe and Wallace, 1985, and references therein). In these reports, many of the stratigraphic names used in this report were formally or informally defined. Donohoe and Wallace (1982) considered the oldest rocks in the Cobequid Highlands to the Mount Thom and Bass River complexes (which they considered to be Helikian or Hadrynian) on the basis of their high metamorphic grade, absence of sedimentary structures and more complicated structural history than the late Precambrian Jeffers and Warwick Mountain formations. They interpreted the Bass River Complex to consist of a gneissic basement unconformably overlain by a cover of quartzites, biotite schists and mafic metavolcanic rocks. Paleozoic rocks mapped by Donohoe and Wallace (1982) consist of upper Ordovician to upper Silurian siltstones, wackes and minor volcanics of the Wilson Brook Formation, and rhyolitic flows and volcanic wackes of the middle to late Silurian Earltown Formation overlain by middle to upper Devonian redbeds, followed unconformably by a thick sequence of bimodal volcanic rocks (Fountain Lake Group) and redbeds.

Cullen (1984) completed a detailed structural study on the Bass River Complex. He interpreted the complex as a multiply deformed basement-supracrustal sequence, the basement represented by the Great Village River Gneiss unconformably overlain by the Gamble Brook Schist and Folly River Schist.

Pe-Piper and Piper (1987) studied the late Precambrian volcanic rocks of the western highlands. They upgraded the Jeffers Formation to group status and divided it into a number of formations. They interpreted the group to consist of a thick sequence of bimodal volcanic rocks overlain by turbidites.

PRESENT WORK

Most of the definitions of formations used by Donohoe and Wallace (1982) are followed in this report. Formational status is assigned to the Gamble Brook and Folly River successions because they can be mapped by their lithotypes and stratigraphic position. Formal definitions will be made elsewhere.

The age and distribution of lithologies in the eastern highlands is shown in Figure 2. In some areas, the distribution shown in Figure 2 is similar to that of Donohoe and Wallace (1982), in other areas it is significantly different. Some of the important differences include the subdivision of the Gamble Brook formation into two mappable units (2a and 2b, Table 1); the recognition of a ductile shear zone between the Gamble Brook formation and the Great Village River Gneiss (i.e. the contact between units 1 and 2, Fig. 2); the documentation of an unconformity between the Gamble Brook and Folly River formations (the contact between units 2 and 3, Fig. 2); the assignment of most of the middle Silurian Earltown volcanics of Donohoe and Wallace (1982) to the late Precambrian (informally described here as the Dalhousie Mountain Volcanics, unit 4c); the abandonment of the Earltown Formation; the placing of much of the early Tournaisian Nuttby Formation into an unnamed late Ordovician Silurian Formation (unit 6); the recognition of volcanic rocks of possible Silurian age (unit 6a) in the northwestern part of the area; the recognition of a major ENE synclinal Carboniferous fold in the southern highlands that effects rocks of the Bass River Complex; the probable ages of some of the plutons based on recent radiometric data; the distribution of Devono — Carboniferous plutons (unit 11), and the abundance of Devono-Carboniferous thrusting and reverse faulting in the northern highlands.

STRATIGRAPHY

The stratigraphy of the eastern Cobequid Highlands is shown in Table 1. Most of the units have been described in detail by Donohoe and Wallace (1985) and Cullen (1984) and are only briefly discussed here. The distribution of lithologies varies considerably across the Rockland Brook Fault (RBF) which here defines the boundary between the northern and

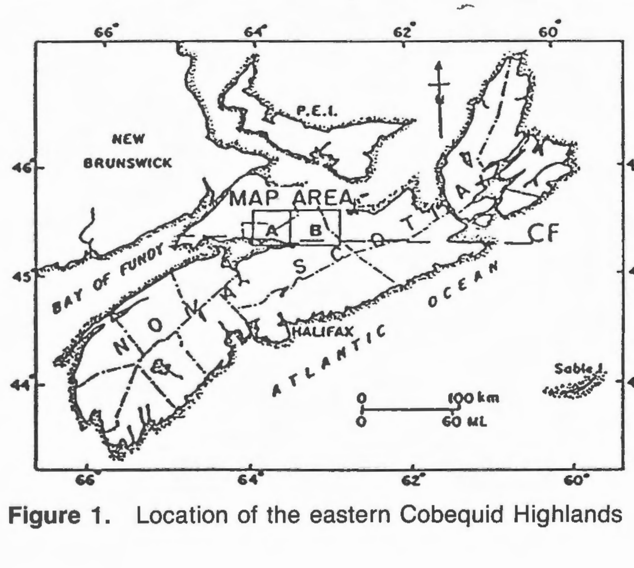


Figure 1. Location of the eastern Cobequid Highlands

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

LEGEND TO FIGURE 2. STRATIGRAPHY MODIFIED AFTER DONOHOE AND WALLACE 1982

DEV-CARB	11 Devono - Carboniferous Plutons------------(a) gabbro (b) diorite (c) granite		
DEV-CARB	9 Fountain Lake Group (a) mafic (b) felsic		10 Nuttby Formation
	---------Unconformity---------		
DEV	8 Portapique River and Murphy Brook Formations		
ORD-SIL	7 Wilson Brook Formation	ROCKLAND BROOK FAULT	
ORD-SIL	6a Silurian Volcanic Rocks ?? ---unconformity---	ROCKLAND BROOK FAULT	6 Undivided ---unconformity---
	?	ROCKLAND BROOK FAULT	5 Late Precambrian - Cambrian Plutons (a) granite gneiss (b) granite (c) appinitic gabbro
HADRYNIAN	4 (a) Jeffers Group (b) Warwick Mountain Formation (c) Dalhousie Mountain Volcanics		3 Folly River Formation Unconformity-------
HADRYNIAN			2 Gamble Brook Formation (a) lower (b) upper Unconformity ? -----
			1 (a) Mount Thom Complex (b) Great Village River Gneiss (age unknown)

southern highlands. The Bass River Complex, the Mount Thom Complex and Hadrynian plutons occur exclusively to the south of the fault. The Late Precambrian Warwick Mountain Formation, and Jeffers Group, and the Devono Carboniferous volcanic rocks occur only to the north of the fault (Table 1). Silurian rocks predominantly occur to the north of RBF. The oldest rocks that unequivocally occur on both sides of the RBF are the Devono-Carboniferous plutons.

Precambrian rocks

The oldest rocks in the area are the gneisses and paragneisses of the Mount Thom Complex (unit 1a) and the Great Village River Gneiss (unit 1b, Bass River Complex). The Mount Thom Complex consists of ortho- and para-gneisses and schists. The orthogneisses are predominantly quartzofeldspathic and contain a foliation defined by biotite, garnet, and muscovite. The Great Village River Gneiss of the Bass River Complex strongly resembles the Mount Thom Complex. Donohoe and Wallace (1985) proposed that these rocks are correlative. The Great Village River Gneiss consists of quartzofeldspathic orthogneisses (Great Village River orthogneiss of Cullen, 1984), biotite-garnet psammitic paragneisses (Donohoe and Wallace, 1985) and hornblendeplagioclase amphibolites (the Bass River amphibolite of Cullen, 1984). All lithologies display several generations of smallscale folds, all of which post-date the development of the dominant metamorphic foliation. The present contact between the Great Village River Gneiss and the overlying Gamble Brook formation (unit 2) is interpreted as a ductile shear zone defined by intense local mylonitization, local development of C-S fabrics, and small-scale isoclinal folds. It is exposed on Portapique River, Great Village River and Rockland Brook (locations c, d and e, Fig. 2) The contact obscures the original relationships. It is therefore not certain whether the original contact was unconformable or not. The age of the Mount Thom Complex and Great Village River Gneiss is unknown. Radiometric data are, at present, poorly constrained (Gaudette et al., 1984).

The Gamble Brook formation (unit 2) consists of two distinct units. The lower part (unit 2a) contains abundant orthoquartzites and arkosic quartzites with interlayered biotitemuscovite-garnet psammitic schists and minor carbonate beds or lenses. A mylonitic fabric and a well developed mineral lineation occur adjacent to the contact with the

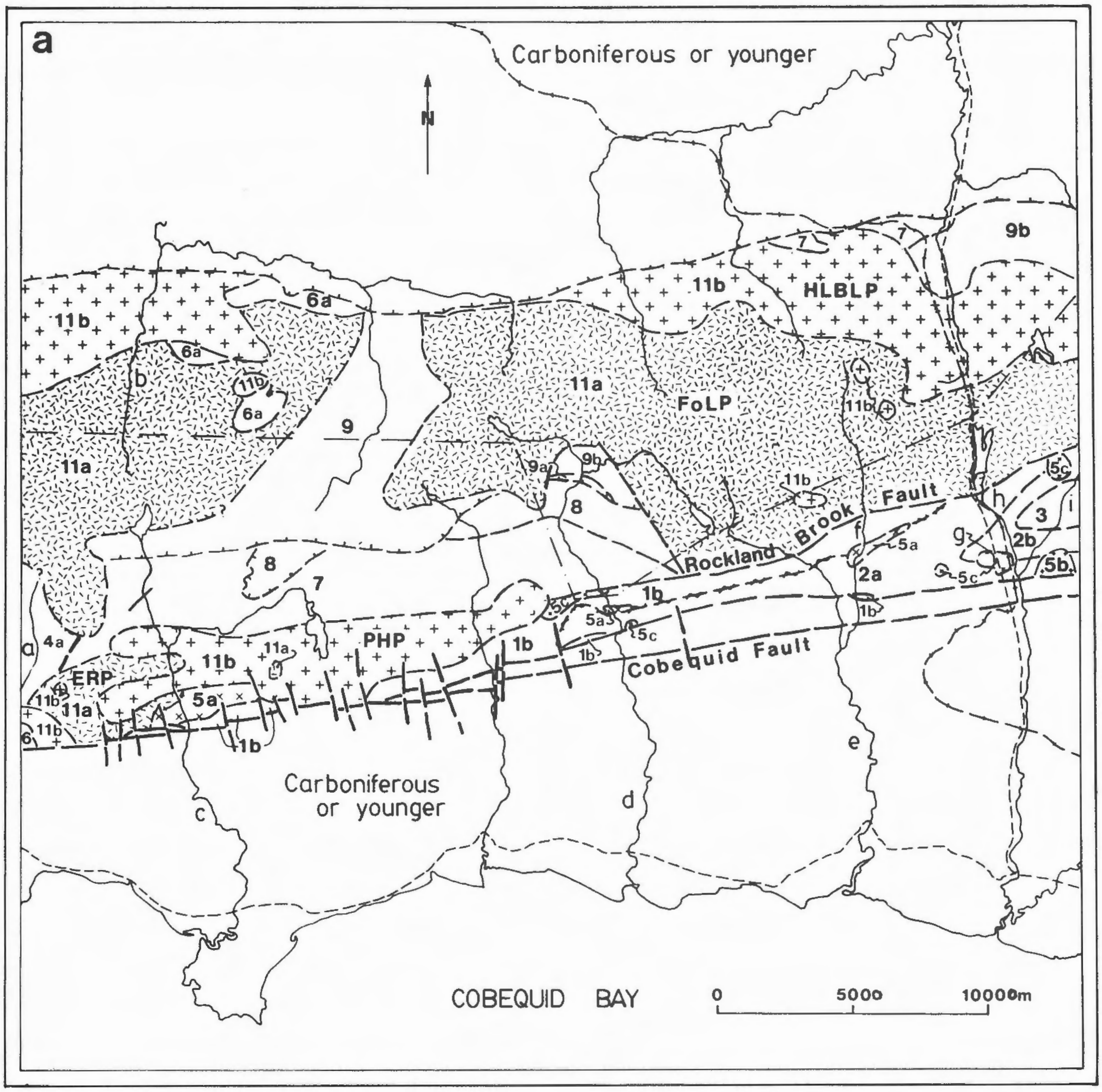
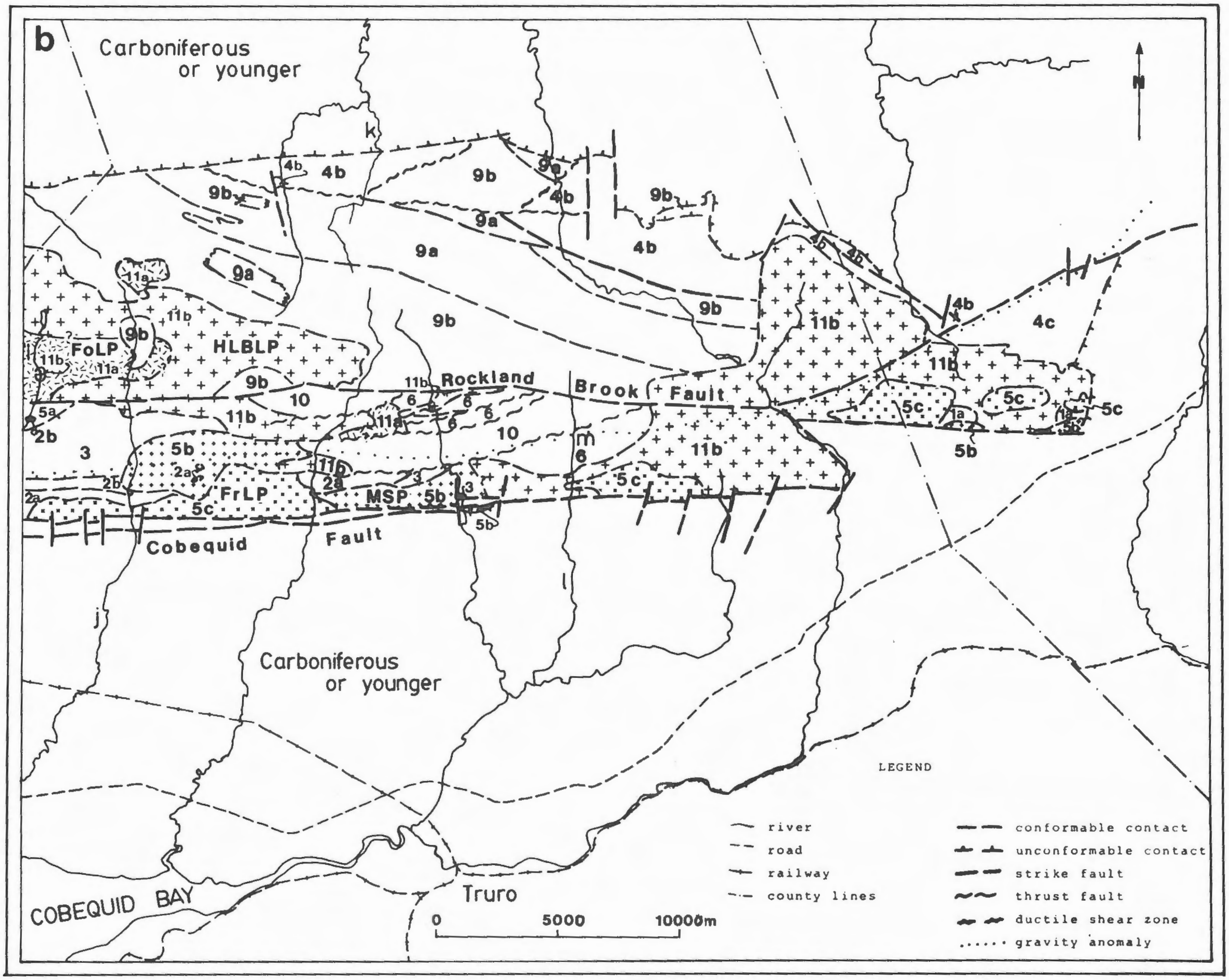


Figure 2. Preliminary geological map of the eastern Cobequid Highlands. See Table 1 for lithologies. Abbreviations used on the map are as follows; ERP, Economy River Pluton; PHP, Pleasant Hills Pluton; FoLP, Folly Lake Pluton; HLBLP, Hart Lake-Byers Lake Pluton; FrLP, Frog Lake Pluton; MSP, McCallum Settlement Pluton. Place-names used in text: a, East River; b, Wyvern; c, Economy River; d, Portapique River; e, Great Village River; f, Rockland Brook; g, Folly Mountain Quarry; h, Folly Lake Quarry; i, East Folly River; j, Debert River; k, Miller Brook; l, North River; m, Nuttby.

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 in many instances (e.g. the quarry south of Folly Lake, location h, Fig. 2). The upper part (unit 2b) of the Gamble Brook formation is distinctly more pelitic and is dominated by biotite-muscovite and biotite-garnet schists in addition to quartzite and psammitic schists. At present, the best estimate for the depositional age of the Gamble Brook formation is derived from a comparison with a similar sequence in southern New Brunswick (Donohoe and Wallace, 1985). There, the Brookville Gneiss is overlain by orthoquartzites and marbles (Greenhead Group). The marbles contain stromatolites of mid-Riphean age (Hofmann, 1974). If the correlation between New Brunswick and Cobequid successions is valid, it provides a depositional age for the Gamble Brook formation and a minimum mid-Riphean age for the Great Village Gneiss (Donohoe and Wallace, 1985). The contact between the Gamble Brook formation and the overlying Folly River formation (unit 3) is interpreted as an unconformity. The contact is exposed on East Folly River (locality i, Fig. 2) and on a tributary of East Folly River. In the former locality the contact is concordant, but, in the latter it is discordant, and the intense mylonitic fabric in the Gamble Brook formation is truncated by a mafic dyke that is only locally cleaved and is petrographically and geochemically indistinguishable from the abundant dykes and flows that characterise the Folly River formation. Both these contacts mark a major lithological break between the quartzitedominated metasediments to the north (i.e. Gamble Brook formation) and a mafic volcanic-dyke complex to the south



(i.e. Folly River formation). An unconformity is also suggested by the contrasting structural style in the Gamble Brook and Folly River formations.

The Folly River formation (unit 3) contains mafic flows, hyaloclastites and tuffs, volcanogenic sediments and minor jasperitic ironstones (the Folly River Schist of Donohoe and Wallace, 1985 and Cullen, 1984). However, the formation also contains abundant mafic dykes and chloritic metasediments that are interlayered with the volcanics and are interpreted as distal turbidites. The cleavage is heterogeneously developed particularily in the mafic volcanic rocks and dykes which may vary locally from massive to schistose. Lavas show a variety of igneous textures, including ophitic to subophitic, pilotaxitic, porphyritic and vesicular. They consist of plagioclase ± augite , actinolite ± hornblende, Fe-Ti oxides, chlorite, epidote , rare quartz and biotite. Dykes within the Folly River Formation have a similar mineralogy and display ophitic to subophitic texture. The upper contact of the Folly River formation is not exposed. The age of the Folly River formation is unknown, but it could be late Precambrian. It may therefore be a correlative of the Jeffers Group and Warwick Mountain Formation. A minimum age for the formation is given by the Debert River Pluton (596±70 Ma, Donohoe et al., 1986, Rb-Sr whole-rock) which post-tectonically intrudes the Folly River formation. The Folly River formation was therefore deformed prior to the latest Precambrian or early Cambrian.

The Jeffers Group (unit 4a) and Warwick Mountain Formation (unit 4b) occur exclusively to the north of RBF. Jeffers Group occurs only in the extreme west near East River (location a, Fig. 2). The Warwick Mountain Formation occurs only in the northern part of the area (e.g. location k, Fig. 2). The description of the Jeffers Group (unit 4a) is given in Pe-Piper and Piper (1987) and is not repeated here. Rocks of the Warwick Mountain Formation (unit 4b) strongly resemble those of the Jeffers Group (Donohoe and Wallace, 1982). They consist of interlayered felsic and mafic volcanics and turbidites and characteristically display a penetrative flat-lying cleavage. The Dalhousie Mountain Volcanics (unit 4c) are similar lithologically, but lack the penetrative fabric. They strongly resemble the late Precambrian rocks of the Keppoch Formation in the Antigonish Highlands. The relationship of these rocks with units 1-3 to the south of RBF is unknown. The contacts with other units in the map area are generally faults; thus age and field relationships are deduced from the western Cobequids (Pe-Piper and Piper, 1987) where the Jeffers Group is cut by late Precambrian plutons. Correlation of the Jeffers Group with the Warwick Mountain Formation is based on lithological comparison. No definitive field relationships are exposed. The age of the Dalhousie Mountain Volcanics is poorly constrained by field relationships. These rocks are tentatively assigned to the late Precambrian on the basis of lithological comparison with the Keppoch Formation.

Paleozoic rocks

Cambrian to early Ordovician strata do not occur in the area. Late Ordovician to early Devonian rocks are sparsely represented in thrust slices to the south of RBF (unit 6) but are more widespread the north of RBF (Wilson Brook Formation, unit 7). Ordovician- early Devonian (undivided) rocks south of the RBF consist of fossiliferous grey and green siltstones and shales.Fossil datasuggest an Ordovician or Silurian age (Donohoe and Wallace, 1982). Comparison with rocks of a similar age to the north of RBF is hindered by poor exposure. The Wilson Brook Formation is dominated by locally fossiliferous green, grey and black micaceous siltstones, wackes and minor tuffs. These rocks resemble the late Ordovician to early Devonian Arisaig Group and a correlation between these rocks has been proposed by Donohoe and Wallace (1985). The age of the Wilson Brook Formation is indicated by fossils which range from Late Llandovery to Pridolian (Donohoe and Wallace, 1982).

Silurian volcanism (Earltown Volcanics of Donohoe and Wallace, 1982) is not as extensive was previously believed. Most of these sequences are assigned herein to the late Precambrian Dalhousie Mountain Volcanics. It is thus suggested that the Earltown Formation be abandoned and that all Silurian sediments to the north of RBF be assigned to the Wilson Brook Formation. A newly recognized suite of steeply dipping bimodal volcanic rocks, tentatively assigned to the Silurian, occurs in the Wyvern area (unit 6a, location b, Fig. 2) apparently underlying sediments of the Wilson Brook Formation. It is also possible, however, that they could belong to the Jeffers Group.

The Wilson Brook Formation is overlain by the Murphy Brook and Portapique River formations (unit 8). The nature of the contacts between these formations is not understood. The Portapique River Formation lithologically resembles the Murphy Brook Formation and these formations may be correlative. They consist of dark grey, green and red clastic rocks. The age of the Murphy Brook Formation is indicated by the presence of Emsian to Eifelian plant fossils. However, no fossils have been found in the Portapique River Formation to date. Based on field relationships, the age of the Portapique River Formation is bracketed between that of the underlying Wilson Brook Formation and overlying Devono-Carboniferous volcanic rocks.

In the Portapique River area (location d, Fig. 2), the Devono-Carboniferous Fountain Lake Group (unit 9a) unconformably overlies the Murphy Brook Formation. The Fountain Lake Group is extensive (Fig. 2) but occurs only to the north of RBF. This sequence consists of a thick succession of interlayered felsic (ignimbrites and flows) and mafic volcanic rocks and minor interbedded clastic sediments. Abundant dykes and sills are associated with these rocks (Chatterjee, 1984). In the eastern highlands, Donohoe and Wallace (1982) divided these rocks into felsic and mafic dominated formations (the Byers Brook and Diamond Brook formations, units 9a and 9b respectively). The Fountain Lake Group is unconformably overlain by the Westphalian Boss Point Formation (Carboniferous or younger, Fig. 2). Contact relationships with other Devono-Carboniferous strata are unclear. In the Debert River area, Cormier (1982) obtained a 341 +4 Ma age (Rb-Sr whole-rock isochron) indicating that at least some of the volcanism is approximately Visean. Thus, from these relationships, the Fountain Lake Group may range from middle Devonian to Namurian in age. However, the stratigraphy may be considerably more complicated than displayed in Figure 2. From drill hole data, Chatterjee (1984) documented the existence of an unconformity within the Byers Brook Formation of Donohoe and Wallace (1982). Therefore, although the age of the sequence above the unconformity is probably Visean, the age of the sequence beneath the unconformity is not known and its surface expression has not as yet been identified. Furthermore, in many areas assigned to the Fountain Lake Group, the age of these rocks is poorly constrained from the field relationships and it is conceivable that some rocks may be as old as late Precambrian.

Devono-Carboniferous strata south of RBF are represented by the Nuttby Formation (unit 10) which contains early Tournaisean spores (Donohoe and Wallace, 1982). The formation contains grey quartzite, minor polymictic conglomerate, dark grey siltstones and shales and is exposed only in the Nuttby area (location m, Fig.2). Relationship with older or younger units are not exposed.

PLUTONIC ROCKS

Plutonic rocks in the eastern Cobequid Highlands appear to be either late Precambrian — early Cambrian or Devono-Carboniferous in age. Precambrian-early Cambrian plutonic rocks (unit 5) consist of granite gneiss, granite -granodiorite and appinitic diorite and occur exclusively to the south of RBF. In contrast, Devono-Carboniferous plutons are ubiquitous and consist of granite, diotite and gabbro. The Rockland Brook Fault forms the southern contact for many of the Devono-Carboniferous plutons in the northern highlands. However, locally, intrusive contacts with basement rocks of the southern highlands are preserved.

Precambrian plutons

Granite gneiss (unit 5a) within the Bass River Complex consists of sericitised porphyroclasts of orthoclase, plagioclase and quartz with rare biotite and muscovite. It contains a mylonitic fabric and a mineral lineation. Its occurrence is restricted to the ductile shear contact zone between the Gamble Brook formation and the Great Village River Gneiss. Contacts with country rocks are welded. They are concordant on a regional scale but are generally locally discordant. The age of the gneiss is poorly constrained. Rb-Sr whole-rock isochrons yield ages of 642+-15 and 626+-22 (Gaudette et al., 1984).

Late Precambrian granite and diorite plutons (units 5b and 5c) between Debert River and North River have been mapped in detail. The Frog Lake pluton (FrLP, Fig. 2) is an appinitic diorite complex of small stocks and sills that intrude the Gamble Brook and Folly River formations. The principal lithologies are diorite and hornblendite. These two rock types contain similar minerals and textures: they differ principally in the relative abundance of plagioclase and amphibole. Most of the rocks are equigranular; some are porphyritic and a few show ophitic texture. They consist principally of plagioclase, amphibole, interstitial biotite, quartz and alkali feldspar. The pluton resembles the late Precambrian Jeffers Brook diorite in the western Cobequid Highlands. Pegmatitic granite veins within the pluton are spatially associated with digestion of xenoliths.

The Frog Lake Pluton is cut by granites of probable late Precambrian age (unit 5b), namely the Debert River Pluton to the northwest and the McCallum Settlement Pluton (MSP, Fig. 2) to the east. These relationships provide a minimum age for the Frog Lake Pluton. The Debert River Pluton consists of granite and granodiorite, whereas the McCallum Settlement Pluton is granite, granodiorite and tonalite. These granitoids appear to be relatively high level, with common roof pendants. The Debert River Pluton is cut by stocks of Devono- Carboniferous granites.

The Late Precambrian granites are medium to coarse equigranular with subhedral granitic texture. The major minerals are quartz, plagioclase (An_{30-45}), orthoclase $\pm$ microcline, with minor biotite (commonly replaced by chlorite), muscovite, epidote, opaque oxides, apatite, and accessory allanite and zircon. The granodiorites are medium grained equigranular rocks and consist of plagioclase , biotite, hornblende, quartz, and minor perthite, apatite sphene, opaque oxides, and actinolite.

Devono-Carboniferous plutons

The northwestern part of the area includes several distinct plutons (compare Donohoe and Wallace, 1982) including granodiorite, granite and gabbro/diorite. The granodiorite is medium grained with biotite and hornblende as major phases. A distinctive white granite which intrudes the possible Silurian volcanic rocks, differs from granites of the Hart Lake — Byers Lake (HLBLP, Fig. 2) and Pleasant Hills plutons (PHP, Fig. 2) (described below) in its colour and lack of graphic texture. Gabbro/diorite contains labradorite, clinopyroxene, hornblende (some rimming clinopyroxene) and biotite.

The Folly Lake and Economy River plutons (FoLP, ERP, Fig. 2) consist of coarse grained gabbro intruded firstly by porphyritic diorite followed by equigranular fine- to medium-grained diorite and granodiorite, late fine grained granitic dykes and sills, (which are locally pegmatitic), and rare porphyritic mafic dykes. The Folly Lake pluton intrudes the Devono-Carboniferous Fountain Lake Group and is intruded by the Devono-Carboniferous (Donahoe et al., 1986) Hart Lake — Byers Lake pluton. Hybridization occurs near the contact, suggesting injection of granite into partially molten diorite. The gabbro is dark grey with prismatic green-grey plagioclase. It has a subophitic texture and consists of equant augite, elongate plagioclase crystals, olive-green hornblende, greenish and bluish actinolite, dark brown biotite, opaque oxide minerals, apatite, zircon, penninite and sphene. Porphyritic diorite contains phenocrysts of subhedral plagioclase showing substantial alteration to muscovite in a matrix of plagioclase, hornblende, actinolite, biotite and opaque oxides, with minor interstitial quartz and alkali feldspar, and accessory apatite, epidote, zircon and sphene. Granodiorite is similar, but has a higher proportion of interstitial quartz and alkali feldspar in graphic intergrowths. Leucocratic granite dykes and sills cutting the pluton appear similar to Devono-Carboniferous granites. The porphyritic mafic dykes cutting the diorite have phenocrysts of plagioclase and clinopyroxene and a groundmass of plagioclase, actinolite and opaque oxides. Unusual megacrystic plagioclase crystals are rounded and appear to be partly digested.

Carboniferous granitic plutons, comprising the Hart Lake — Byers Lake pluton (HLBLP, Fig. 2), Pleasant Hills Pluton (PHP, Fig. 2) and granitic pods within the Folly Lake pluton, are distinctive red to pink granites. The large plutons typically have a chilled margin hundreds of metres wide consisting of porphyritic or equigranular granophyric granites. The dominant phase is coarse grained perthite- and quartz-rich granite. This pluton is cut by later pegmatitic veins, many of which show mineralization, (magnetite, pyrite, chalcopyrite and ?siderite). The porphyritic margins of the plutons have phenocrysts of plagioclase, perthite and quartz in a micrographic, micro- myrmekitic or fine grained granitic groundmass containing zircon, sphene, biotite, opaque oxides, rutile, apatite, chlorite, hornblende and occasional fluorite. There are also rare miarolitic cavities. Coarse granite shows graphic intergrowths of perthite and quartz, with rare plagioclase, riebeckite, brown tourmaline, hornblende, sphene, zircon, fluorite and opaque oxides. Granite pods have a fine grained granitic texture and comprise plagioclase, orthoclase, quartz, hornblende, green biotite, dark brown biotite, apatite, opaque oxides and epidote. In places, feldspars show rapakivi texture. Some prismatic hornblende crystals contain feldspar inclusions. The Hart Lake — Byers Lake pluton is intruded by a number of large diabase dykes with a fine grained subophitic texture consisting of plagioclase, augite, actinolite, antigorite pseudomorphs after olivine, biotite, chlorite, opaque oxides and apatite.

In several areas, late gabbros appear to cut all earlier plutonic rocks. These gabbros have been mapped in the Wyvern plutonic complex, and as stocks along the Cobequid fault zone. The dykes that cut the Hart Lake-Byers Lake pluton are probably equivalent to these gabbros.

STRUCTURE

The detailed structure of the Bass River Complex is summarized here. The Great Village River Gneiss contains moderately inclined, generally shallowly plunging isoclinal folds of several generations. Fold axes vary locally indicating that sheath folds may exist. Many generations fold a metamorphic fabric and are thought to be related to the second phase of a deformation and are broadly cogenetic. They are coaxial and co-planar. No D_1 folds (i.e. folds to which the compositional layering is axial planar) were found (compare Cullen, 1984). The earliest of the D_2 folds contain an axial planar fabric and the later D_2 folds deform this fabric. The contact between the Gamble Brook Formation and the Great Village River Gneiss is interpreted as a ductile shear zone. This zone is characterised by local tectonic interleaving of the gneiss and quartzite, by syntectonic intrusion of granite gneiss and by locally intense mylonitization imparting a mylonitic fabric within the quartzitic units of the Gamble Brook Formation. Mineral lineations, interpreted as stretching lineations, generally plunge moderately to the east. Isoclinal folds within the Gamble Brook Formation fold the mylonitic fabric but are thought to be cogenetic with mylonitization. These folds are isoclinal, moderately inclined, with a wide range of fold axis orientations. To the south of the contact zone, mylonitization and mineral lineations are only locally developed or absent (e.g. the Folly Lake Quarry, location g, Fig. 2). There the folds are relatively steeply plunging. Although several generations of co-axial — co-planar folds may be recognized (the earliest of which folds bedding), all the folds are thought to be cogenetic. The style of deformation in the Gamble Brook Formation probably represents sinistral transtension.

The Folly River Formation is deformed by isoclinal folds and thrusts. The first generation of folds deforms bedding and imparts penetrative axial planar cleavage in the metasediments . In the mafic rocks, the cleavage is only locally penetrative. Second generation folds deform the cleavage and therefore are most visible in the metasediments. They are generally co-planar and co-axial with first generation folds. The style of folding is consistent with dextral transpression.

The age of the metamorphic fabric in the Great Village River Gneiss is unknown. The age of mylonitization is constrained by the age of intrusion of the late Precambrian syntectonic granite gneiss in the vicinity of shear zone. Xenoliths of mylonitic quartzite of the Gamble Brook Formation occur in the Debert River Pluton which has been dated at 596 ± 70 Ma (Donohoe et al., 1986). The Debert Pluton is interpreted to be post-tectonic with respect to both the Gamble Brook Formation (on the basis of the above relationships) and the Folly River Formation, and therefore provides a minimum age for deformation of both formations.

The entire eastern Highlands was deformed locally by thrusts and isoclinal folds subsequent to the deposition of the Nuttby Formation and Fountain Lake Group, i.e. synchronous or post- Namurian. The deformation is most intense in the southern Highlands where the entire map pattern is deformed into a regional shallowly plunging ENE -trending isoclinal syncline that involves the Bass River Complex as well as younger rocks (Fig. 2). The overall ENE trend of these folds is consistent with dextral Carboniferous motion on major east-west faults (e.g. Keppie, 1985; Donohoe and Wallace, 1985). To the north of RBF, the style of deformation is different. Evidence of deformation is given by the angular unconformable relations between the Fountain Lake Group and Westphalian rocks (Donohoe and Wallace, 1982), and by the occurrence of NNW-dipping thrusts and reverse faults. The best example of thrusting occurs on Miller Brook (location k, Fig. 2) where the Warwick Mountain Formation is thrust upon the Fountain Lake Group.

DISCUSSION

Although the relative ages of events seem well established from field relationships, systematic geochronological work is required to document accurately the age of these events such that the regional significance of the area in the evolution of the Avalon Terrane can be evaluated. The interpretation that a ductile shear zone separates the Great Village River Gneiss from the Gamble Brook Formation and that an unconformity exists between the Folly River and Gamble Brook Formations represents a major advance in the understanding of the Bass River Complex. The geochemistry of the Folly River Formation is being studied to aid in the interpretation of the tectonic environment.

The stratigraphy of the Silurian rocks in the northern highlands is well documented (Donohoe and Wallace, 1982) and it may be possible to extrapolate this stratigraphy in order to subdivide the Wilson Brook Formation. The documentation by Chatterjee (1984) of an angular unconformity within the Fountain Lake Group indicates that this succession requires detailed study and is one of the least understood sequences in the Cobequid Highlands. The petrology of Devono-Carboniferous plutons is complicated by evidence of several phases of granitoid plutonism and of assimilation and hybridization of gabbro by diorite.

The Paleozoic structural history remains a problem. It is not clear whether the deformation in the northern and southern Highlands is synchronous or whether they represent separate events. The development of Carboniferous folds in the southern Highlands indicates that there may be difficulty in distinguishing the effects of Precambrian and Carboniferous deformation. The age and style of faulting is also a major problem. Some of the E-W faults display evidence of Carboniferous movement (e.g. RBF) while others do not. It is clear that the RBF represents a major fault within the Cobequid Highlands because no rock units older than Devono-Carboniferous plutons can be unequivocally demonstrated to occur on either side of this fault.

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Geology and mineralization of the Jumping Brook metamorphic suite, Faribault Brook area, western Cape Breton Island, Nova Scotia.[1]

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Abstract

The Jumping Brook metamorphic suite in the Faribault Brook area hosts gold-bearing, polymetallic sulphide mineralization of uncertain origin. Detailed geological mapping (1:2500) and logging of 2900 m of drill core have been conducted to document the lithologies, lithological succession, and structure associated with the mineralization. A sequence of siliceous volcanogenic metasedimentary rocks lies beneath the Faribault Brook metavolcanic rocks, indicating two phases of felsic volcanism. Cross-sections indicate the volcanic — sedimentary sequence was anticlinally folded during F2 deformation. Polymetallic Zn-Cu-Pb mineralization of pre-metamorphic origin is associated with the felsic volcanogenic metasedimentary rocks. Disseminated- and vein-style arsenopyrite, possibly of several generations, is also present. The assemblage pyrrhotite + pyrite + sphalerite + chalcopyrite + arsenopyrite is associated with D3 shearing and may have been, in part, mobilized from highly deformed and altered metabasite during metamorphism.

Résumé

La suite métamorphique de Jumping Brook située dans la région de Faribault Brook contient une minéralisation sulfurée à caractère polymétallique et aurifère, et d'origine incertaine. On a effectué des travaux détaillés de cartographie géologique (1/2500) et étudié les diagraphies de 2900 m de carottes de sondage de façon à obtenir des données pétrographiques, et des données sur la succession lithologique et sur les structures associées à la minéralisation. Une séquence de roches métasédimentaires siliceuses d'origine volcanique repose sous des roches métavolcaniques de Faribault Brook, ce qui indique qu'il y a eu deux phases de volcanisme felsique. Les coupes stratigraphiques révèlent que le plissement de la séquence volcanique et sédimentaire, survenu au cours de la déformation F^2, a donné naissance à un anticlinal. La minéralisation polymétallique en Zn, Cu et Pb, d'origine prémétamorphique, est associée aux roches métasédimentaires issues de roches felsiques d'origine volcanique. On rencontre aussi de l'arsénopyrite, peut-être constituée en plusieurs étapes, sous forme de disséminations et de filons. L'assemblage pyrrhotine + pyrite + sphalérite + chalcopyrite + arsénopyrite est associé à un cisaillement D^3 et a peut-être été partiellement mobilisé à partir d'une métabasite fortement déformée et altérée, au cours du métamorphisme.

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INTRODUCTION

The Jumping Brook metamorphic suite (Jamieson et al., 1987) in the Faribault Brook area occurs in a north-south trending belt and consists of lower amphibolite grade volcanic, plutonic, and sedimentary rocks (Fig. 1) which host numerous base and precious metal showings (Milligan, 1970; Covey, 1979; Chatterjee, 1980; Ponsford and Lyttle, 1984; Woods, 1986). Despite extensive exploration, the origin, age, and structural -stratigraphic setting of these showings have remained obscure. This study concentrates on the Faribault Brook — Grandin Brook transect (Fig. 2) across the deformed and erratically mineralized transition from metavolcanic to metasedimentary rocks. Field work during the summer of 1987 consisted of detailed logging of 2900 m of drill core, and mapping for 6 km along Faribault and Grandin brooks at 1 :2500 scale. Drill core and surface sulphide showings were sampled for future geochemical and petrological work. The preliminary field results presented below document the lithologies, lithological succession, and some features of the associated mineralization.

PREVIOUS WORK

The regional geology of the west — central Cape Breton Highlands has been described by McLaren (1955), Milligan (1970), Currie (1975, 1982, 1987), Barr et al. (1985), and Jamieson et al. (1987). A regional stratigraphic framework has been determined by Woods (1986) and Jamieson et al. (1987). Craw (1984) documented the structural history of the suite along the Cheticamp River. Plint (1987) and Plint and Jamieson (in press) describe the metamorphic and tectonic history of the region between Pleasant Bay and Faribault Brook.

Local investigations within the Faribault Brook area have produced diverse conclusions regarding mineralization and deformation. Covey (1979) emphasized the association of quartz-sericite schists with mineralization and explained a lack of correlation between drill holes by block faulting. Chatterjee (1980) studied mineralized quartz-sericite schists and proposed a hydrothermal replacement origin for the sulphides. Connors (1986) and Plint et al. (1986) found garnet overgrowing sphalerite in quartz — sericite schist indicating a pre-metamorphic origin for some sulphides.

REGIONAL GEOLOGY

The Jumping Brook metamorphic terrane in the Faribault Brook area is bounded to the north and northwest by the Cheticamp Pluton, to the west by the Fisset Brook Formation and to the east by the Salmon Pool Pluton (Fig. 1). Known contacts with the Jumping Brook metamorphic suite in this area are faulted, although Woods (1986) considered the Cheticamp Pluton to be intrusive based on spatially related soil geochemical anomalies and the presence of concordant felsite layers in sedimentary rocks along lower Faribault Brook.

Late Proterozoic — Cambrian

Foliated plutonic rock (Unit 1)

A small, foliated, dioritic body is exposed along the northern branch of Fisset Brook. Barr et al. (1986) and Jamieson et al. (1987) considered this lithology to be distinct from the Cheticamp Pluton, although Woods (1986) considered it an outlier of the Cheticamp Pluton.

Cheticamp Pluton (Unit 2)

The Cheticamp Pluton is locally foliated along its margins and nonfoliated internally, and ranges from granodiorite to monzogranite in composition (Barr et al., 1986). The radiometric age of the pluton is Cambrian(Cormier, 1972; Barr et al., 1986; and Jamieson et al., 1986) with a U-Pb (zircon) age of 550 +/- 8 Ma (Jamieson et al., 1986). Muscovite from the southern part of the pluton has a late Cambrian 40Ar/39Ar cooling age (Reynolds, pers. comm., 1987).

Lower Paleozoic

Jumping Brook metamorphic suite (Unit 3)

The Jumping Brook metamorphic suite has been divided into several lithological units by Jamieson et al. (1987), three of which constitute the low grade portion of the suite present in the vicinity of Faribault Brook. Regional mapping by Jamieson et al. (1987) and property mapping by Woods (1986)

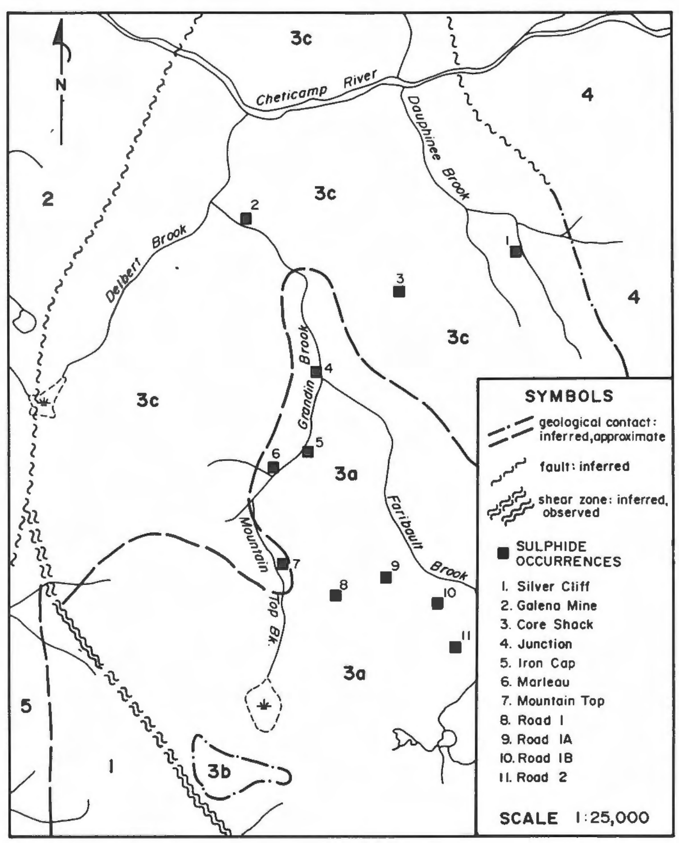

Figure 1. Geology of the Jumping Brook metamorphic suite in the vicinity of Faribault Brook, showing the locations of the most important sulphide occurrences. Unit numbers correspond to text.

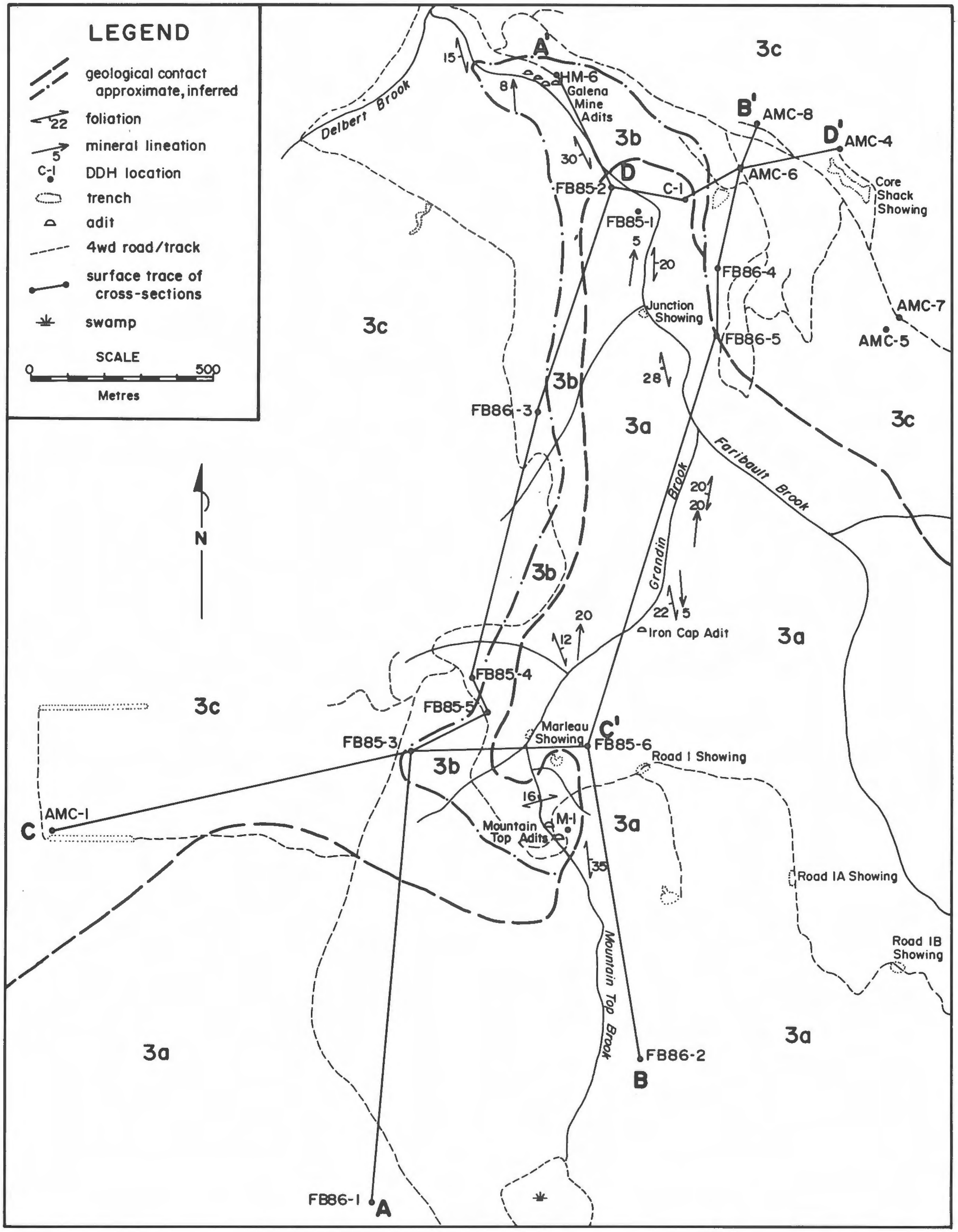

Figure 2. Location of cross-section surface traces and diamond drill holes in Figures 3 to 6. Local geology inferred from drill log data. Unit numbers correspond to text.

have determined that the Faribault Brook metavolcanic rocks (unit 3a) form the structurally lowest unit of the suite, succeeded in turn by the Barren Brook schist (unit 3b) and the Dauphinee Brook schist (unit 3c). A 1:10 000 scale map shows the distribution of these units in the Faribault Brook area (Fig. 2). The trace of unit 3b(Fig. 2) is drawn from drill core data in correlation with surface exposures.

The Faribault Brook metavolcanic rocks (unit 3a) include massive to strongly schistose tholeiitic rocks, probably derived from flows, and are cut by metadiorite intrusions (Connors, 1986). The Barren Brook schist (unit 3b) consists of fine- to coarse-grained, quartz-eye-bearing metasedimentary rocks and quartz — sericite schist. Connors (1986) identified a felsic volcanic protolith for the quartz — sericite schist based on prismatic terminations and resorption textures within some quartz grains, although most of the quartz is detrital. The Dauphinee Brook schist (unit 3c) comprises fine- to medium-grained pelites and semi-pelites which are compositionally layered and isoclinally folded.

Conclusive evidence concerning age relations between the Faribault Brook metavolcanic unit and the structurally overlying metasedimentary rocks has not been documented. The metavolcanic and metasedimentary rocks may be coeval, conformable and interlayered, or may be separated by a structural discontinuity(Jamieson et al., 1987; Woods, 1986). The absolute age of the Jumping Brook metamorphic suite has not been established and it may be late Proterozoic (Woods, 1986) or Lower Paleozoic (Currie, 1982; Jamieson et al., 1986; Jamieson et al., 1987; Currie, 1987; Plint and Jamieson, in press).

Devonian — Carboniferous

Salmon Pool Pluton (Unit 4)

The Salmon Pool Pluton forms the western margin of the Jumping Brook metamorphic terrane in the Cheticamp River area. The pluton is a homogeneous, undeformed syenogranite with a distinctive intrusion breccia along its eastern boundary (Jamieson et al., 1987). Jamieson et al. (1986) determined the age of the the pluton to be 365 +5/-10 Ma, and considered it subvolcanic to the Fisset Brook Formation.

Fisset Brook Formation (Unit 5)

The Fisset Brook Formation, of Devonian age, is a suite of bimodal volcanic rocks and associated redbeds (Kelley and Mackasey, 1964; Blanchard et al., 1984) which forms two belts near the western margin of the Cape Breton Highlands. Undeformed, generally plagioclase-phyric diabase dykes, probably related to the Fisset Brook Formation, have locally intruded rocks of the Jumping Brook metamorphic suite.

STRUCTURAL SETTING

A complex structural history has been reported by various authors for the region (Craw, 1984; Connors, 1986; Plint, 1987; Currie, 1987: Jamieson et al., 1987). Only those structural features which relate to the association of lithology and mineralization are reported here.

Primary structures such as bedding, graded beds, and granular textures are commonly preserved in the competent metasedimentary rocks. However, primary structures within metabasites such as contacts between individual flows, or between flows and intrusions, are rarely preserved owing to the heterogenous nature of the deformation. Primary structures are tectonically modified and serve only as indications of the original lithology.

The dominant structural element is a pervasive foliation (S^1) produced by tight to isoclinal folding. In the Faribault Brook area, primary structures are oriented parallel, or nearly parallel, to this foliation. Bedding is transposed along foliation planes, and rootless, intrafolial folds are common in the metasedimentary rocks although rare in the metabasites.

A second phase of north-south trending anticlinal structures (F^2) has been mapped and is associated with a uniform, north-northeast trending mineral and boudin lineation (L^2). The plunge of the lineation is 5° north in Mountain Top Brook, steepens to 20° north along Grandin Brook, and flattens to 5° north in lower Faribault Brook. The plunge of this lineation defines the plunge of the units shown in cross-section.

Ductile shear zones (D^3) in metabasite trend parallel to the pervasive S^1 foliation. Shear zones range from 5 cm to 1 m wide and have been enriched in biotite and chlorite. Sericite and garnet are developed along the margins of some shear zones. Randomly oriented, porphyroblastic hornblende has overgrown some shear zones, while in others mimetic overgrowths of hornblende are parallel to the shear direction. An envelope of 5 % to 40 % biotite surrounding these shears commonly hosts stringer and disseminated mineralization. Shear zones have formed preferentially in metabasites adjacent to metasedimentary rocks, particularly in the mixed sub-unit, owing to the competency contrast between these lithologies.

The final stage of deformation involves high angle brittle faulting (D4) along northeast-trending fault zones. These zones, 10 cm to 2 m wide, are characterized by chlorite, carbonate, and hematite fault gouge and are commonly associated with fold hinges of F^2 folds.

LITHOLOGIC SUB-UNITS

Metabasite

Metabasite is used as a composite term for several distinct mafic lithologies in the area, although it is dominated volumetrically by massive to moderately foliated, fine-grained chlorite-albite-hornblende schist. Silica content is correlated with colour; silica-rich rocks are blue, silica-poor are green. An increase in competence corresponds with higher silica. Zones with high proportions of epidote 'balls' and associated carbonate veins are common. Contacts between texturally distinct lithologies of this type vary locally from sharp to gradational. The metabasite is interpreted as flow rock with some tuff.

Contiguous zones of moderately foliated, carbonate-rich metabasites occur in some drill holes, and are composed mainly of fine-grained chlorite, carbonate, biotite and hornblende. Moderately foliated chlorite-albite-hornblende schist with 20 % to 40 % disrupted carbonate veins and segregations is less common. Carbonate-rich zones occur at random intervals in the metabasite and are normally less than 5 m thick; however, drill holes FB85-1 and FB85-6 each contain zones up to 30 m.

Medium- to coarse-grained, equigranular, massive to weakly foliated plagioclase-hornblende-chlorite schists are common in the lower sections. These have interlocking grain textures and fine-grained margins and are probably metadiorites.

Mixed-zone

This sub-unit is distinguishable by characteristic interlayering of metabasite and metasedimentary rocks at 10 cm to 2 m intervals. Metadiorites are not known in these zones although all other types of metabasite are represented. The metasedimentary rocks include quartz-sericite schist, siliceous metagreywacke, and the chloritic metagreywacke described below.

Quartz-sericite schist

The sub-unit has an overall waxy appearence and a white to yellow colour, and consists of quartz-eyes in a quartz-sericite matrix with local biotite, chlorite and garnet is the common assemblage. Quartz-eyes are white and blue rounded to prismatic grains, ranging in modal proportion from 10 to 40 per cent. Garnet is locally important and is commonly associated with massive and vein type mineralization. The presence of fuchsite, noted during field mapping (Woods, pers. comm., 1987), was confirmed in drill core.

Felsic intrusion

A distinctive pink quartz-sericite schist, which lacks biotite and chlorite, is consistently associated with arsenopyrite — chalcopyrite mineralization. Mineralized quartz- sericite schist of this type cuts the foliation and apparent layering in highly schistose metabasite in outcrop at the Marleau Showing in upper Grandin Brook. Siliceous, competent margins about 10 cm wide suggest chilling, and thus an intrusive origin. Similar possible felsic intrusions are distinguished in drill core by their colour, distinctive sulphide mineralization, and in drill hole FB86-1 by an apparent intrusion within a metadiorite. This lithology forms zones less than 2 m wide, and is not known above the level of the uppermost metabasite.

Metagreywacke

This unit comprises a variety of flysch lithologies. Siliceous metagreywacke contains quartz and minor biotite, chlorite, chloritoid and sericite. It also contains 5 % to 10 % rounded, 5-mm quartz-eyes and minor, rounded, 2-mm feldspar grains. Included within this sub-unit are biotite quartzite layers, 50 cm to 3 m thick, that contain rare 2-mm to 5-mm quartz-eyes. Cross-sections show a unit of siliceous metagreywacke (Fig. 3) derived from felsic volcanic products at the deepest known level of the suite, which may represent a fold limb of the Barren Brook schist known on surface, or a separate unit related to an earlier phase of felsic volcanic activity. East-west axial planes of folds, required if the upper and lower siliceous metasedimentary rocks are stratigraphic equivalents, have not been documented, therefore two periods of felsic volcanic-derived turbidite sedimentation are inferred.

Metagreywacke contains proportionally more finer grained quartz and feldspar grains, increased chlorite, biotite, sericite and chloritoid contents and much less matrix silica, compared with siliceous metagreywacke.

Chloritic metagreywacke exhibits a further decrease in grain size, with minor 1-mm to 2-mm quartz grains, no visible feldspar, and increased chlorite content. Porphyroblasts of 1-2 mm biotite or chloritoid are common, and locally, 1-mm garnet porphyroblasts are present.

LITHOSTRATIGRAPHIC CROSS-SECTIONS

North-south litholostratigraphic cross-sections (Fig. 3, 4) are spatially separated by 400 m and are parallel to both the strike of the units defined by Jamieson et al. (1987) and the regional schistosity. East-west cross-sections (Fig. 5, 6) are drawn 1.5 km apart, normal to regional strike and schistosity.

The cross-sections show vertical variations in lithology which are more complex in detail than the regional relationships defined by Jamieson et al. (1987). The deepest vertical sections were drilled in holes C-1 and FB85-3, which penetrated similar structural levels. Both holes encountered approximately 50m of interlayered metagreywacke and siliceous metagreywacke of the Barren Brook schist, and only minor thicknesses of metabasite in their deepest sections. At the structural top of this section, the dominantly metasedimentary lithologies are gradational upwards, across a mixed zone characterized by thin layers of alternating metasediment and metabasite, into a 75-m-thick sequence of metabasite of the Faribault Brook metavolcanic unit. The metabasite sequence is overlain by a 75-m section of metabasite containing thick intervals of metagreywacke and siliceous metagreywacke. This interval is strongly deformed, particularly along discrete shear planes within the metabasite. The uppermost metabasite layer marks the top of the Faribault Brook metavolcanic unit and the lower contact of the Barren Brook schist unit. The Barren Brook schist is represented by a 25-m-thick sequence of quartz-sericite schist, quartzite, and siliceous metagreywacke. In drill hole FB86-3, these lithologies include a silicified metagreywacke with relict quartz eyes. The Barren Brook schist grades upwards into chloritic metagreywacke of the Dauphinee Brook schist. Metasedimentary lithologies are interlayered on all scales so that the boundary between the two metasedimentary units is approximate. Only the lower 125 m of the Dauphinee Brook schist unit has been penetrated by drill holes or is exposed in Faribault Brook, although regionally the unit is estimated to be greater than 800 m thick.

The broad unit designations of Jamieson et al. (1987) are useful for regional correlations. However, it is necessary to subdivide these units because more numerous, distinct lithological changes are represented in the local succession. The lithologic sub-units are plotted on the cross-sections (Fig. 3 to 6), along with structural features and major types of mineralization, to show lateral and vertical variation in

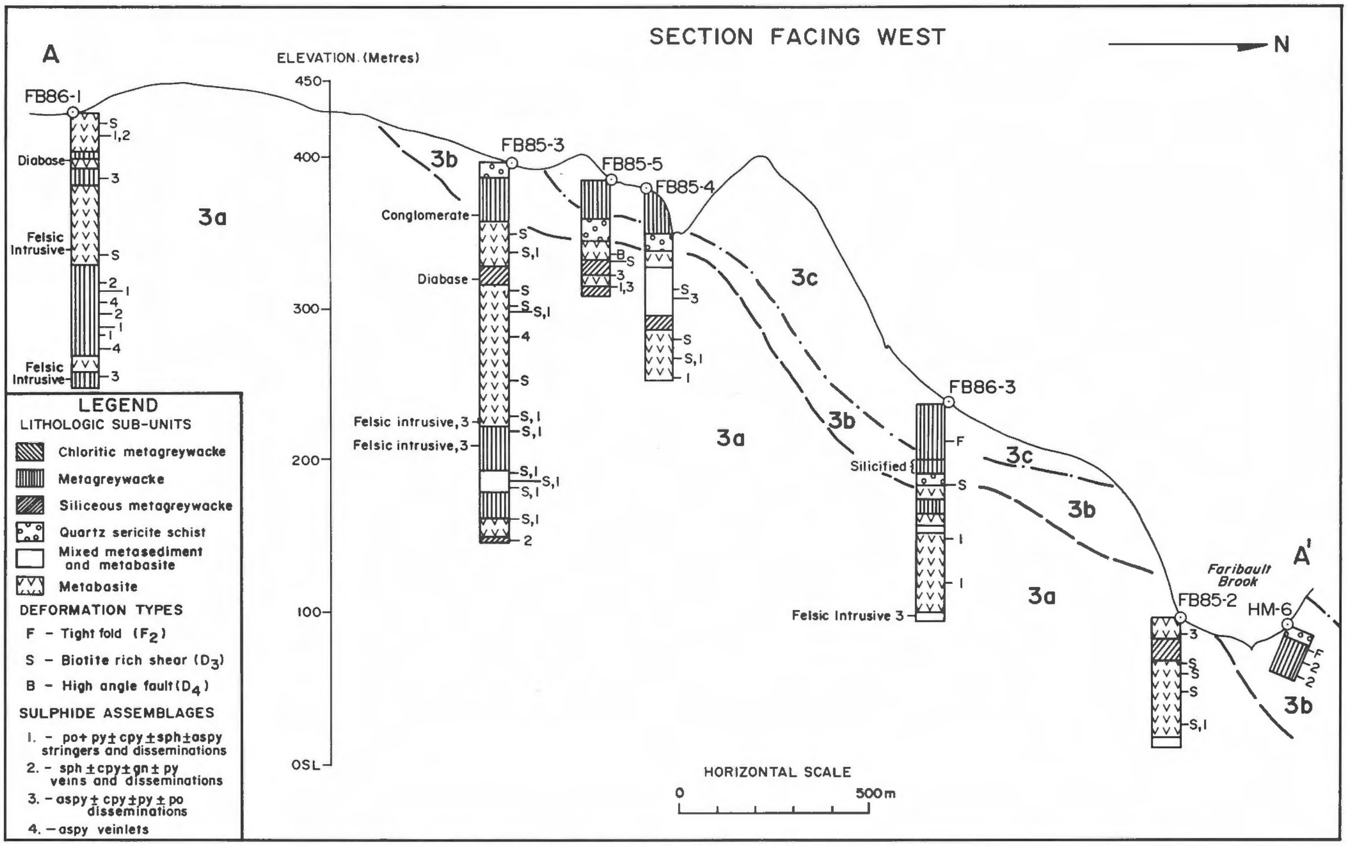

Figure 3. North-south cross-section A to A' along the west side of Faribault Brook. Vertical exaggeration is 4:1. Unit numbers correspond to text.

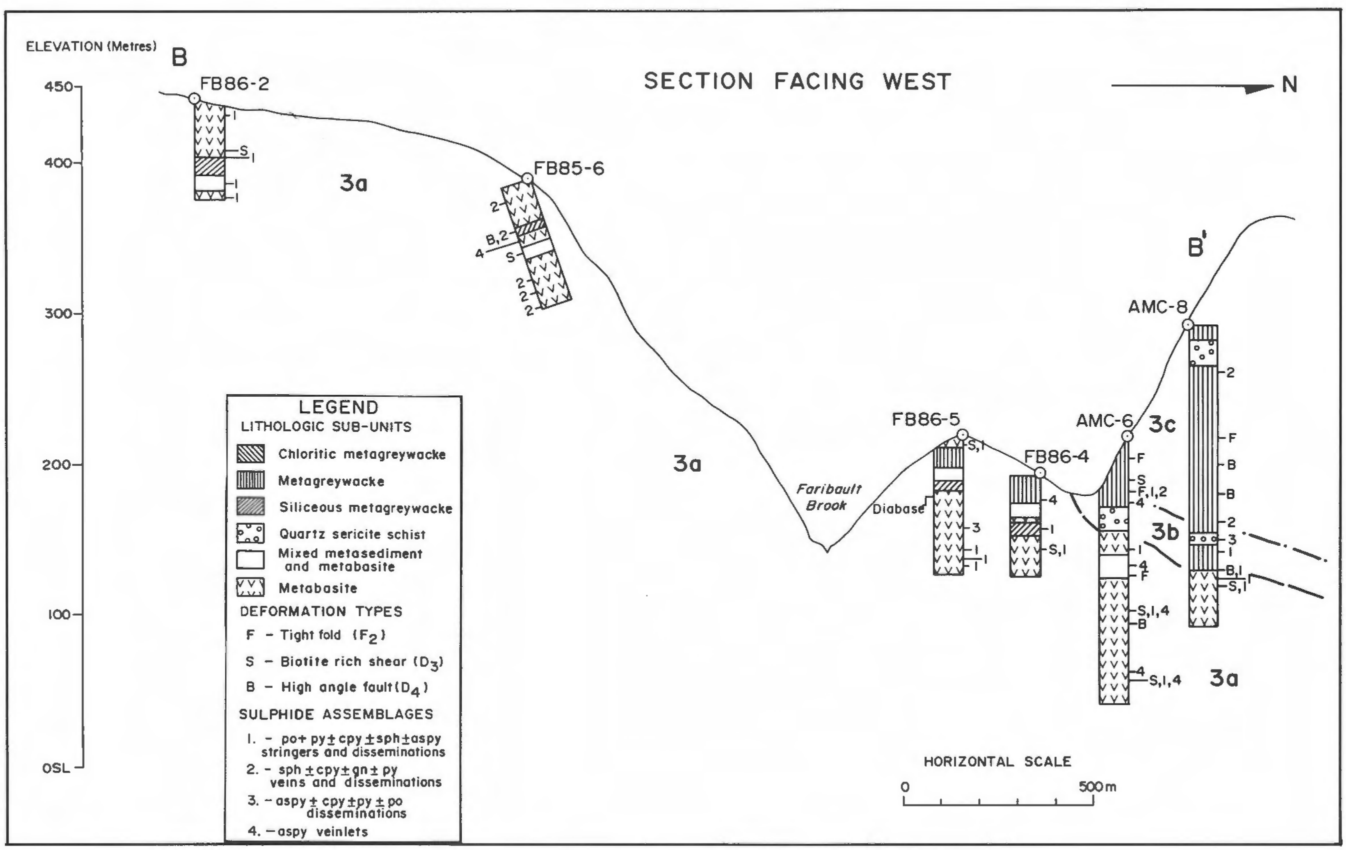

Figure 4. North-south cross-section B to B' along the east side of Faribault Brook. Vertical exaggeration is 4:1. Unit numbers correspond to text.

lithology and the dominant associations between lithology, structure, and mineralization.

MINERALIZATION

Although mineralization is widespread, sulphide assemblages are highly variable owing to the effects of deformation, remobilization, and possible later introduction, of some or all sulphides. Common associations between mineralization, structure, and lithology form the basis for the classification of mineral assemblages shown in Figures 3 to 6. Most sulphide-bearing rocks have anomalous gold ranging from a few hundred ppm to 0.5 oz. / t (unpublished data A.L. Sangster). Four principal assemblages were documented in the drill core:

1. Pyrrhotite + pyrite + chalcopyrite + sphalerite + arsenopyrite.
2. Sphalerite + chalcopyrite + galena + pyrite ± arsenopyrite.
3. Arsenopyrite + chalcopyrite + pyrrhotite + pyrite.
4. Arsenopyrite + pyrrhotite + chalcopyrite + pyrite + quartz.

There is a strong association between mineralization and shearing below the uppermost metabasite layer. The assemblage pyrrhotite + pyrite + chalcopyrite + sphalerite + arsenopyrite occurs as cleavage-parallel disseminations, and in thin, foliation-parallel quartz veinlets hosted by biotite-rich, well foliated metabasites. These are the most common styles of mineralization observed in drill core. Pyrrhotite, pyrite, chalcopyrite and sphalerite occur as ragged, anhedral grains. Arsenopyrite is subhedral to euhedral when disseminated; however, it occurs primarily in thin, isoclinally folded and transposed, arsenopyrite-rich veinlets.

The association sphalerite + chalcopyrite + galena + pyrite ± arsenopyrite is common in surface sulphide occurrences and in drill core. On the surface, this assemblage is hosted by quartz — sericite schist at the Galena Mine, Core Shack and Mountain Top showings, but also occurs at the Junction showing where it is hosted by thinly interlayered siliceous metasediment and metabasite. Anhedral, cleavage-parallel disseminations and foliation-parallel and low angle cross-cutting, sulphide-bearing quartz veins are the common form of mineralization.

In drill core this assemblage is restricted to massive clots along fractures within local F1 fold hinges in competent metasedimentary rocks, and to cross-cutting, calcite-filled fractures associated with late brittle deformation. Plagioclase-phyric diabase dykes, probably related to the Devono-Carboniferous Fisset Brook Formation, also host fracture-filling carbonate veins of this type.

The pink, sericitic, felsic intrusions host fine-grained, cleavage-parallel disseminations of anhedral to subhedral arsenopyrite, chalcopyrite, pyrrhotite and pyrite. At the Marleau showing, mineralization is preferentially concentrated in the competent chilled margins of the dyke.

Sub-massive arsenopyrite veins, from 1 cm to 25 cm wide, with accessory pyrrhotite, chalcopyrite, pyrite and quartz are associated with high strain zones in friable, chlo-

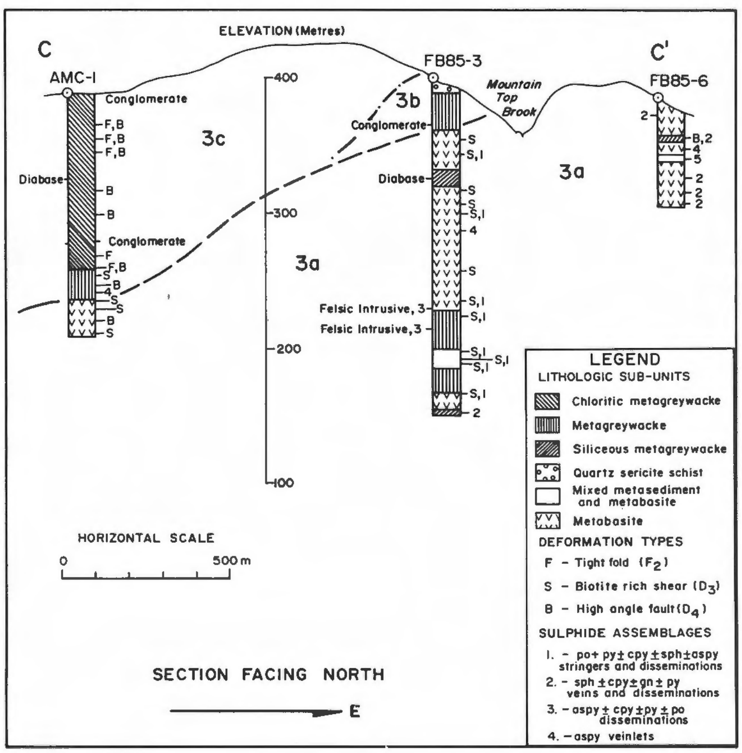

Figure 5. East-west cross-section C to C' through the Mountain Top Brook area, across the southern end of the map area. Vertical exaggeration is 4:1. Unit numbers correspond to text.

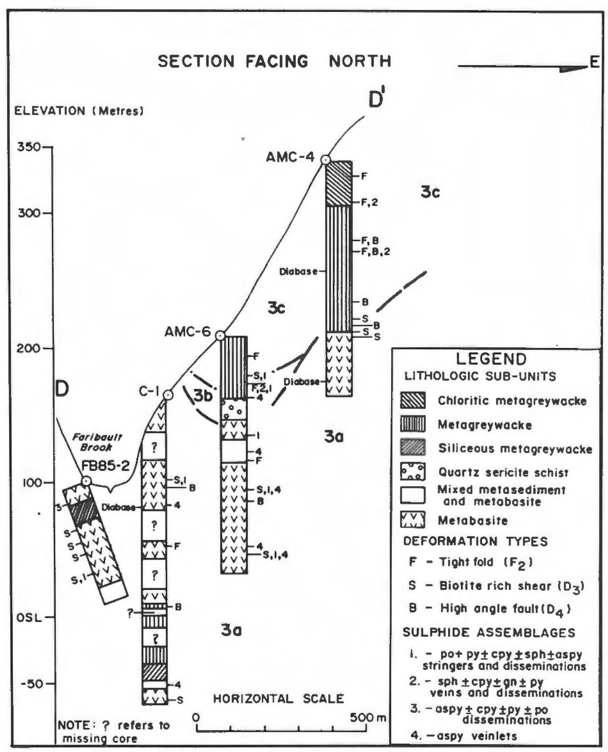

Figure 6. East-west cross-section D to D' through lower Faribault Brook, near the northern end of the map area. Vertical exaggeration is 4:1. Unit numbers correspond to text.

ritic metabasite and less commonly in sericite schist. Garnet and sericite are common along the margins of wider zones in metabasite, such as those at the Road 1a and Road 1b occurrences. Arsenopyrite veins in drill core form erratically distributed zones through the metabasite. Euhedral and acicular arsenopyrite is disseminated adjacent to some zones containing arsenopyrite veins. Similar arsenopyrite-rich veins, with surrounding disseminated arsenopyrite are found in well-foliated quartz-sericite schist; the latter unit also contains base metal mineralization at the Core Shack and Mountain Top showings.

In addition to the above, folded, boudinaged quartz veins host some or all sulphide assemblages known in the area and may be associated with sulphide remobilization. Some later, open-folded, locally boudinaged quartz veins host minor sulphides and some are barren. The relative ages of these veins and disseminated veins is not known.

DISCUSSION

The Jumping Brook metamorphic suite is interpreted as a lower Paleozoic volcanic-sedimentary sequence (Jamieson et al., 1987) that represents tholeiitic mafic volcanism (Connors, 1986) succeeded by a brief period of felsic volcanism, and followed by greywacke deposition.

The structural pattern shown by the surface geology in the Faribault Brook area(Fig. 2) and the cross-sections (Fig. 3 to 6) is strongly controlled by the F2 folding, consistent with the ubiquitous, north-northeast directed mineral, minor fold axis and boudin lineation. The plunge of the units in cross-section corresponds with the plunge of the lineations mapped on surface. Major F2 antiforms have been mapped along the Cheticamp River by Craw (1984), and the surface trace of the stratigraphic units and their dip in cross-section indicate an antiformal structure with a north-northeast-trending axis passing through the lower Faribault Brook area.

Subsequent D3 shearing of the suite has been accomodated within the comparatively ductile metabasites. This has produced an F2 fold-dominated upper metasedimentary unit, and a D3 shear-dominated metabasite unit in which previous structures have been masked.

The source of mineralization in the Faribault Brook area is obscure, complicated by the diversity of the sulphide assemblages, styles of occurrence and severe structural and metamorphic overprint. Based on the lithological and structural controls discussed above, some characterization of sources is possible.

The sphalerite + chalcopyrite + galena + pyrite assemblage is restricted to felsic volcanogenic sedimentary rocks, except where associated with carbonate. Garnet overgrows sphalerite in this lithology at Galena Mine (Connors, 1986), indicating a pre-metamorphic origin for at least some of these sulphides. The spatial restriction and pre-metamorphic origin of sulphides is consistent with polymetallic Zn-Cu-Pb mineralization associated with felsic volcanism in island-arc sequences.

The constant association of disseminated, anhedral, arsenopyrite-chalcopyrite mineralization with pink, felsic intrusions suggests a primary relationship rather than secondary introduction. The relationship between this style of arsenopyrite occurrence and the vein types discussed above is unknown.

The association of the pyrrhotite + pyrite + chalcopyrite + sphalerite + arsenopyrite assemblage with biotite-rich metabasite formed during D3 shearing implies a genetic relationship between the occurrence of these sulphides and deformation. In accord with this interpretation is the virtual absence of Pb in these zones in relation to its average concentration of less than 10 ppm in undeformed metabasite, and the presence of minor sphalerite and chalcopyrite in relation to Zn and Cu concentrations of greater than 100 ppm (Jamieson, unpublished data) in undeformed lithologies. Alternative origins for the sulphides in metabasite-hosted shear zones include introduction from outside the suite, remobilization from a sulphide-rich source within the suite, or leaching of the metabasite by metamorphic fluids. Comparably mineralized shear zones outside the Jumping Brook metamorphic suite are not known. Remobilization from a sulphide-rich source is possible; however, no evidence of such a source has been obtained. Whole rock concentrations of Fe, Zn, Cu and Pb are proportional to concentrations of sulphides found in the mineralized shear zones. This fact and the strong coincidence of mineralization with zones of ductile shear indicate that the mineralization may have been mobilized by metamorphic fluids from a pre-existing source in the unaltered, undeformed protolith.

A consideration of the origin of the sulphides is dependent on the age assigned to the Jumping Brook metamorphic suite. The Late Precambrian age which has commonly been assigned in the past permitted epigenetic derivation of the metals from either the Cambrian, Cheticamp Pluton, from the Devonian Salmon Pool Pluton, or syngenetic derivation from a volcanic-hydrothermal system associated with the extrusion of the Jumping Brook volcanic rocks. The recent assignment of a Lower Paleozoic age to the volcanic rocks(Currie, 1982; Jamieson et al., 1986) and the defined pre-metamorphic character of the earliest sulphides(Connors, 1986) eliminates the Cheticamp granite and the post-tectonic Salmon Pool granite as potential sources. The only viable genetic process, considering these relationships and the host volcanic stratigraphy, is the operation of a hydrothermal system associated with the volcanic rocks in which the occurrences are found. All stages of mineralization would then be considered to be products of remobilization from components originally deposited in the volcanic pile.

CONCLUSIONS

Two periods of felsic volcanism and related sedimentation are recorded in drill core from the Faribault Brook area. The volcanic — sedimentary sequence was folded during F_2 deformation about a north-northeast-trending axis located in lower Faribault Brook. Various styles of mineralization within this sequence have formed during different stages, including a pre-metamorphic stage for the base metal assemblage hosted by felsic, volcanic-derived metasedimentaryy rocks and for the arsenopyrite + chalcopyrite assemblage hosted by felsic intrusions. Arsenopyrite disseminations and massive veins may overprint some pre-metamorphic mineralization, and later D3 shear-hosted base metal mineralization may have been derived from unaltered metabasites by metamorphic fluids generated during ductile shear. The ultimate source of the metals is the volcanic system in which they occur.

ACKNOWLEDGMENTS

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Distribution and significance of Ordovician flysch units in western Newfoundland

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Quinn, L., Distribution and significance of Ordovician flysch units in western Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 119-126, 1988.

Abstract

Three main units of Ordovician flysch are present in western Newfoundland: The autochthonous Goose Tickle Formation (Llanvirn), dominated by silty argillite with minor sandstones; Mainland Sandstone (Llanvirn-Llandeilo), also autochthonous, and predominantly classical turbidites and thick-bedded, massive sandstones; the allochthonous Lower Head formation (Arenig-Llanvirn), characterized by thick-bedded, massive sandstones and minor conglomerates.

The Mainland Sandstone occurs on the Port au Port Peninsula, and north of Bellburns. The Goose Tickle Formation occurs east and northeast of these localities, perhaps suggesting an eastward fining of the autochthonous clastics.

Some of the detritus in all three formations can be directly related to immediately surrounding rock units. The high content of quartz and feldspar may be derived from older siliciclastic sediments. Basement of the Long Range Complex might also have provided a source as it is now thrust over allochthonous sediments and could have been structurally high during the Ordovician.

Résumé

Il existe dans l'ouest de Terre-Neuve trois grandes unités de flysch ordovicien: la formation autochtone de Goose Tickle (Llanvirnien), dominée par une argilite limoneuse accompagnée d'une petite quantité de grès; le grès de Mainland (Llanvirnien et Llandeilien), aussi autochtone et surtout composé de turbidites classiques et de couches épaisses de grès massifs; la formation allochtone de Lower Head (Arénigien et Llanvirnien), caractérisée par des couches épaisses de grès massifs et une petite quantité de conglomérats.

Le grès de Mainland se manifeste dans la péninsule de Port-au-Port et au nord de Bellburns. La formation de Goose Tickle se trouve à l'est et au nord-est de ces localités, ce qui pourrait peut-être indiquer un affinement vers l'est des roches clastiques autochtones.

On peut établir une corrélation directe entre une partie du matériel détritique présent dans les trois formations et les unités rocheuses situées immédiatement aux alentours. La forte teneur en quartz et feldspath a peut-être pour origine les anciens sédiments silicoclastiques. Il se peut également que le socle du complexe de Long Range en ait constitué la source, étant donné qu'il chevauche maintenant les sédiments allochtones, et qu'il aurait pu constituer une élévation durant l'Ordovicien.

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INTRODUCTION

The Humber Zone of western Newfoundland (Williams, 1979) represents the early Paleozoic continental margin of eastern North America (Williams and Stevens, 1974), a part of the Appalachian miogeocline (Williams and Hatcher, 1982). Western Newfoundland comprises three main tectonic elements (Fig. 1):
1) Grenvillian basement of the Long Range Complex.
2) Autochthonous Cambro-Ordovician sedimentary rocks.
3) Allochthonous terranes (the Hare Bay and Humber Arm Allochthons) which contain Cambro-Ordovician sedimentary rocks in addition to volcanic and ophiolitic rocks.

Grenvillian basement is overlain unconformably by clastic rocks of Precambrian-Lower Cambrian age which are products of Late Precambrian rifting and subsequent development of the passive margin of the Iapetus ocean. During the Cambrian period, a widespread carbonate platform developed on the newly formed continental margin. This remained stable until Middle Ordovician times when bank subsidence resulted in the deposition of deeper water limestones and shales. This passive margin sequence was blanketed in Llanvirn times by flysch deposits (Stevens, 1970). The older sedimentary strata of the Humber Arm and Hare Bay allochthons are interpreted as slope/rise deposits mainly coeval with the shallow water autochthonous succession. They are overlain by flysch slightly older at its base than the autochthonous flysch.

Allochthonous sedimentary rocks are structurally overlain by volcanic and ophiolitic rocks. Melange zones separating these structural slices are related to assembly and emplacement of the allochthons. The entire package is interpreted to have been assembled during eastward subduction (Strong et al., 1974), followed by obduction of oceanic crust (upper structural slices) and parts of the continental slope/rise (lower allochthonous slices) across the continental shelf. The Humber Arm Allochthon was in place by Middle Ordovician time since its leading edge is preserved under the neoautochthonous Caradocian Long Point Group (Rodgers, 1965; Stevens, 1970). Post-emplacement thrusting of the Long Range Complex over the Humber Arm Allochthon has occurred, possibly during the Acadian Orogeny (Williams et al., 1985; Cawood and Williams, 1986; R. Grenier, pers. comm., 1987).

Lower Paleozoic sandstone units in the Humber Zone fall into two main tectonic categories:

1. Late Precambrian or Cambrian marine sandstones deposited as a result of rifting and initial development of a passive continental margin.
2. Ordovician marine sandstones deposited in an active margin setting.

These categories may be subdivided as follows:

1a. Autochthonous ?Precambrian-Cambrian rift related sandstones.

1b. Sandstones broadly equivalent to 1a. which were deposited farther offshore in deeper water and have subsequently been transported over the ancient continental margin.

2a. Autochthonous Ordovician marine sandstones deposited in an active margin setting.

2b. Sandstones deposited in the same general tectonic setting as those in 2a and subsequently transported as part of an accretionary prism over the ancient continental margin.

Classification into these few simple types can avoid confusion over the many formational names which have arisen out of areally limited field mapping projects. The subdivision of the two main groups emphasises that differences are to be expected between units which are now in place and their equivalents which have been transported. Table 1 lists the sandstone formations of western Newfoundland and their classification according to the scheme outlined above.

This report deals with the field component of a provenance study of Lower Paleozoic sandstones of western Newfoundland, particularly those of category 2.

ORDOVICIAN FLYSCH UNITS

Three main units of Ordovician flysch can be identified in western Newfoundland. They are: the Mainland Sandstone (category 2a); the Goose Tickle Formation (category 2a); the Lower Head Formation (category 2b).

The Mainland Sandstone

The Mainland Sandstone outcrops on the Port au Peninsula (see Fig. 2). It has a directly measureable thickness of at least 700m — if the extensive covered areas are assumed to represent continuous section then the total thickness of the unit may be 1.5km (Stevens, 1970). The unit conformably overlies the Llanvirnian Cape Cormorant Formation (Klap-

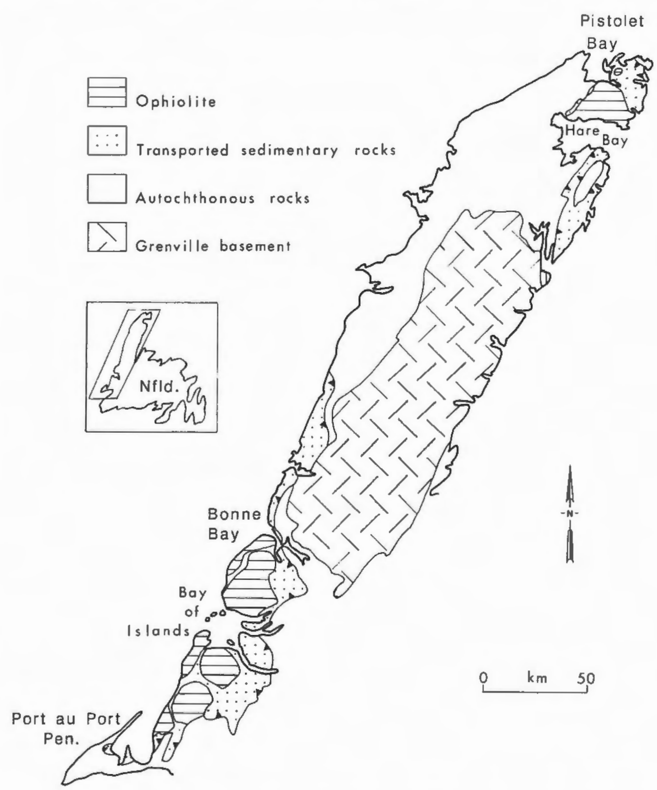

Figure 1. Map of western Newfoundland showing the major tectonic elements.

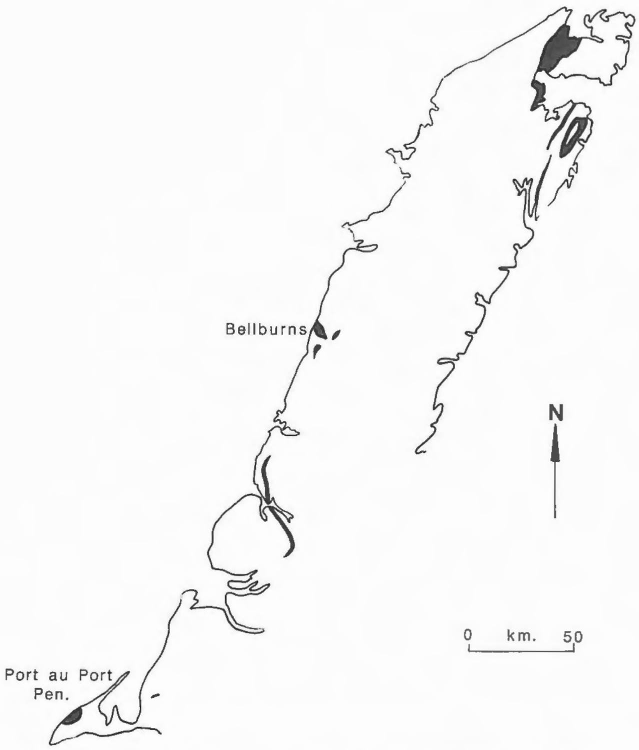

Figure 2. Sketch map showing the distribution of category 2a sandstones in western Newfoundland.

Table 1. Classification of western Newfoundland sandstone units

1a) Bradore Formation	Schuchert and Dunbar (1934)
1a) Bateau Formation	Williams and Stevens (1969)
1a) Hawke Bay Formation	Schuchert and Dunbar (1934)
1b) Maiden Point Formation	Cooper (1937), Betz (1939), Tuke (1968).
1b) Summerside Formation	Stevens (1970)
1b) Irishtown Formation	Stevens (1970)
1b) Blow me Down Brook/Sellars Formation	Stevens (1970), Quinn (1985)
2a) Goose Tickle Formation	Cooper (1937), Tuke (1968)
2a) Norris Point formation	Williams et al. (1985)
2a) Mainland Sandstone	Schillereff and Williams (1979)
2b) Lower Head Formation	Williams et al. 1985, James and Stevens, 1987

pa et al., 1979). Graptolites in the upper part of the Mainland sandstone give a Llandeilo or possibly Caradoc age (James and Stevens, 1982). On the west coast of the Port au Port Peninsula the Mainland Sandstone is conformably overlain by shallow water carbonates of the Long Point Group. Farther east it is structurally overlain by the Humber Arm Allochthon which in turn is stratigraphically overlain by the Long Point Group (Rodgers 1965; Stevens, 1970). There are faults and discontinuities within the Mainland Sandstone which render it difficult to correlate between different sections, particularly as exposure is poor in inland areas.

The Mainland Sandstone has two major facies: 1) Medium- to thick-bedded, fine- to medium-grained sandstones showing abundant complete or partial Bouma sequences (Fig. 3). The most common sequence is bce. Flutes are common on bases of beds. The sandstones are friable and have a high percentage of mud. Black and green shale detritus is common; 2) Thick bedded to very thick bedded, amalgamated, medium- to coarse-grained massive sandstones (Fig. 4). The thick-bedded facies commonly shows dewatering structures, and occasionally displays irregular tops. A crude parallel stratification is present in these sandstones which in wave washed outcrops shows low angle (15 degrees) truncations. Individual sandstone beds commonly become amalgamated or pinch out along strike. A minor facies is a coarse grained, trough crossbedded, medium-bedded sandstone. Crossbedding is more common at the top of the section, where set thickness is as great as 1m. As the outcrop is very friable, structures in recessive intervals are hard to determine. However where seen on wave washed outcrops, finer grained intervals consist of parallel laminated, rippled and convoluted

Figure 3. Thin and medium bedded turbidites in the Mainland Sandstone, Port au Port Peninsula.

Figure 4. Thick bedded massive sandstones of the Mainland Sandstone, Port au Port Peninsula.

siltstones. Some thinning and fining upward sequences are visible in the sandstones but are not common throughout the section. The Mainland Sandstone is dominated by facies of Class C2.1 or C2.2 (classical turbidites) of Pickering et al. (1986) which were deposited by high or intermediate concentration turbidity flows. The massive sands belong to facies group B, disorganised sands, and are mainly of class B1.1.

North of Bellburns (see Fig. 2) up to 100m of sandstones are exposed overlying shales of the Black Cove Formation. These sandstones are fine- to coarse-grained, thin- to thick-bedded and show acd portions of the Bouma sequence. Climbing ripple lamination is common. Mean bed thickness is about 30 cm, with a maximum bed thickness of up to 1.5m. The sandstones are commonly amalgamated with internal scour surfaces showing several centimetres relief, and commonly amalgamate or pinch out a few metres along strike. Some of the thick- to very thick bedded sandstones show crossbedding in sets up to 30 cm thick. The sandstones are interbedded with packages up to 1m thick of black and green banded argillite.

This unit shows similarities with both the Mainland Sandstone and the Goose Tickle Formation, but on the basis of the high percentage of sands versus silts, it is here included with the Mainland Sandstone.

Paleocurrent directions from both the Port au Port and Bellburns localities indicate flow to the southwest.

The Goose Tickle Formation

The Goose Tickle Formation is well exposed on the shores of Pistolet Bay and Hare Bay on the Northern Peninsula (Fig. 2). It also occurs in a north-south-trending zone in inland areas west of the Long Range Complex. A prominent unit of fine grained sandstone and argillite in the vicinity of Bonne Bay has been informally named the Norris Point formation (Williams et al., 1985). However it has been interpreted as a correlative of the Goose Tickle Formation (Williams et al., 1985), and is here included with that unit. In the Bonne Bay region rocks of this formation outcrop parallel to the thrusted base of the Humber Arm Allochthon. The Goose Tickle Formation is of Llanvirnian age (Williams and Smyth, 1983). The unit is so highly deformed that no accurate thickness estimate can be made. However, about 80m of relatively undeformed section are exposed in Pistolet Bay. At the type section in Hare Bay there is at least 100 m near the base of the section with very poor exposure towards its top. The Goose Tickle Formation overlies either the Table Point Formation or a thin, poorly developed Black Cove Formation at various different localities. Limestone breccias which might be correlatives of the Cape Cormorant Formation (Klappa et al., 1979) are interbedded with sandstones of the Goose Tickle Formation. This will necessitate a revision of the stratigraphic nomenclature as these breccias cannot be considered a part of the Table Head Group. This revision is currently being made by the author and S. Stenzel of Memorial University.

The top of the Goose Tickle Formation is everywhere faulted against the overriding allochthons. The unit is deformed in open (westwards away from the allochthon) to tight (eastwards) folding, with a strong associated cleavage. Goose Tickle lithologies commonly form the matrix to melanges at the bases of the allochthons.

Since the stratigraphic units within the allochthons are discordant with the basal thrusts, the Goose Tickle Formation is juxtaposed against different allochthonous formations at different localities.

The Goose Tickle Formation is a mainly fine-grained siliclastic unit with minor limestone breccia in its basal part. Despite the difficulties in estimating thicknesses, the Goose Tickle Formation is divisible into lithological packages.

The unit is mainly characterized by green silty argillite, with black mudstone bands up to 1 cm thick which are between 4 and 15 cm apart (Fig. 5). Massive silty horizons are up to 50 cm thick, but mean thickness is about 20cm. This facies is interbedded in the lower parts of the section with resistant, calcareous tan weathering fine- to medium-grained sandstones (Fig. 6). These commonly show ripple cross-lamination and convolute laminations, with infrequent graded bedding and parallel laminations. The beds are commonly amalgamated with an event thickness of 15-30 cm. Beds

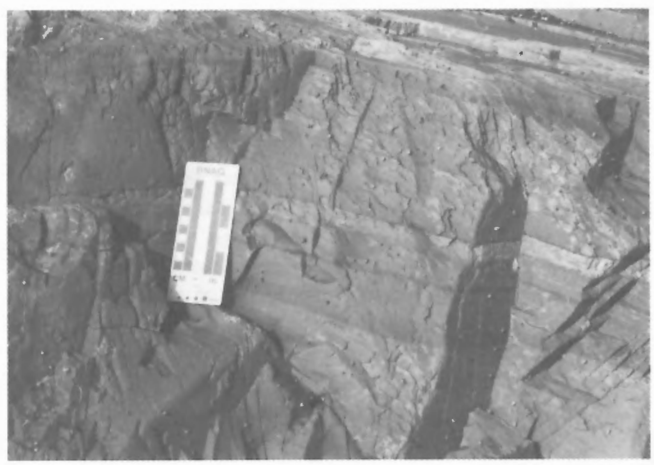

Figure 5. Silty argillite facies with black bands in the Goose Tickle Formation, Pistolet Bay.

Figure 6. Resistant sandstone beds in the Goose Tickle Formation, Pistolet Bay.

have a mean thickness of about 30 cm, but occasionally reach as much as 1m. Flutes and loads are common in this facies.

In the lower part of the section (probably within the basal 50m) limestone breccia and calcarenite horizons up to 5 m thick are interbedded with resistant sandstones and silty argillite, although the breccia is not everywhere present. In places the breccia is associated with slumping in the resistant sandstone facies. As one goes farther upsection, the resistant sandstone facies decreases in importance. At the type section of the Goose Tickle Formation in Hare Bay, it is succeeded by thin, tan weathering, fine sandy lenses with irregular ripple laminae, and convolute laminae. These sands have a mean thickness of 5 cm and are regularly interbedded at 5- to 20-cm intervals with the silty argillite facies.

Farther up section, the silty argillite facies begins to contain horizons up to 1 cm thick of sandstone which contains a large proportion of black and green shale detritus. This facies coarsens and thickens generally upsection. Near contacts with the Northwest Arm Formation this facies is represented by greywacke beds up to 40 cm thick, some of which show trough crossbedding in sets up to 30 cm thick. Near the contact with the Northwest Arm Formation there are conglomerates (up to 2 m thick where observed by the author, but reported up to 10 m thick by Williams and Smyth (1983)) (Fig. 7). The conglomerate is clast supported and is characterized by a high percentage of black and green shale pebbles, with other clasts consisting of laminated fine grained limestone, tan weathering calcareous sandstone, and pyrite nodules. Volcanic cobbles have also been reported from this conglomerate (Williams and Smyth, 1983). At Triangle Point in Pistolet Bay, this conglomerate horizon might be represented by a granule calcarenite containing black and green shale pebbles. The black and green shale chips have been interpreted by previous workers (Tuke, 1968; Knight, 1986) as having been derived from the allochthonous Northwest Arm Formation.

The sequence described above holds for the Goose Tickle Formation at Pistolet Bay, and the northwest part of Hare Bay, where the unit is interpreted as being structurally overlain by the Northwest Arm Formation (Tuke, 1968; Williams and Smyth, 1987; Knight 1986). The Northwest Arm Formation is a chaotic sequence of black and green shales, limestones and sandstones whose matrix is inferred to be of Tremadoc age (R. Stevens, pers. comm. 1987). At Triangle Point in Pistolet Bay, relationships between the Goose Tickle Formation and the Northwest Arm Formation are complex and the two units appear to be very closely intercalated. Conglomerates similar to those in the Goose Tickle Formation also occur within the Northwest Arm Formation. The possibility that some of the contacts between these two units may be conformable cannot be overlooked and closer study with better age control is required for the Northwest Arm Formation. The relationship between the Goose Tickle Formation and the Northwest Arm Formation is obviously a very close one since black and green shale chips are not detrital components in the Goose Tickle Formation, where it is juxtaposed against allochthonous formations other than the Northwest Arm Formation, for example in the eastern part of Hare Bay, and near Bonne Bay.

Paleocurrent directions in the Goose Tickle Formation indicate flow to the south and west (Williams and Smyth, 1983; Knight, 1986).

Lower Head Formation

The Lower Head Formation is repeated across an east-west zone north of Bonne Bay in a series of east dipping imbricate thrust slices, (Williams et al., 1985) (Fig. 8).

Equivalents of the Lower Head Formation exist in the Bay of Islands area, but these do not include the Blow me Down Brook Formation, which has been reinterpreted as an

Figure 7. Conglomerate containing black and green shale clasts, upper part of Goose Tickle Formation, Pistolet Bay.

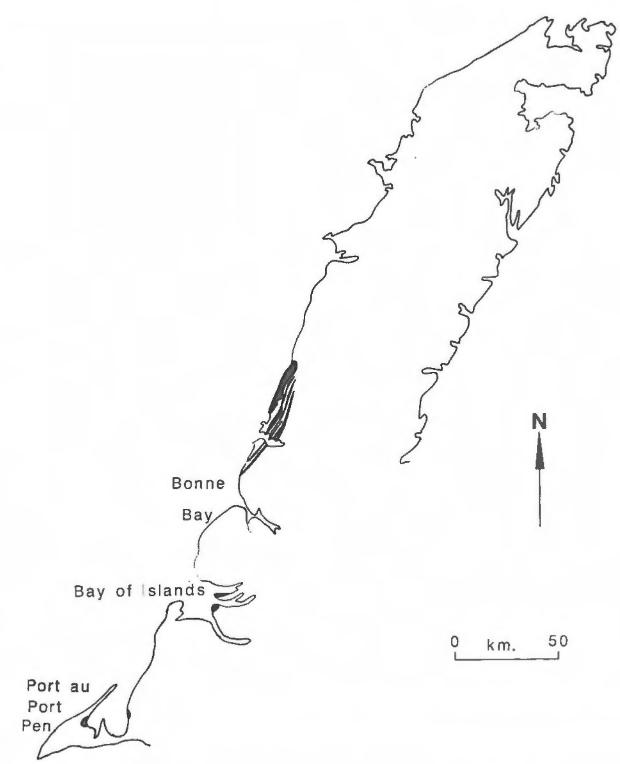

Figure 8. Sketch map showing the distribution of category 2b sandstones in western Newfoundland.

older, probably Cambrian sandstone isolated in a structurally high slice beneath the igneous slices (Quinn, 1985; Waldron, 1985). These equivalents are currently unnamed, as are those on the Port au Port Peninsula (see Fig. 8). No equivalents of the Lower Head Formation are present in the Hare Bay Allochthon. The Lower Head Formation ranges from Late Arenig in age (D. Johnstone pers. comm. in Quinn, 1985) to Llanvirn (Cawood and Williams, 1986). The unit is nowhere more than 350m thick and it conformably overlies deep water carbonates of the Cow Head Group (James and Stevens, 1986). The first influx of sandstones is sharp, but at several localities red and green shales and limestone conglomerates of Cow Head aspect are interbedded with sandstones for several tens of metres upsection from the first sandstone occurrence.

The top of the Lower Head Formation is everywhere faulted against the Cow Head Group, except on the Port au Port Peninsula where its equivalents are unconformably overlain by the Long Point Group. The main part of the section is characterised by thick-very thick bedded massive sandstones, medium-coarse grained, poorly sorted with scattered granules and pebbles of argillite (Fig. 9). These commonly nucleate spherical concretions. Fluid escape structures such as pillars and sheet structures are also common. Occasionally a crude parallel stratification is visible. The thick sandstone beds are commonly amalgamated, with patches of shale rip-up clasts and irregular scour surfaces marking the bases of depositional events. Some of the beds have more matrix-poor crossbedded tops.

Sandstone dykes and sills are common at the base of the Lower Head Formation in many localities, particularly where it overlies red and green cherts and shales of the Cow Head Group. A distinctive facies consists of channelised conglomerates containing cobbles of limestone and chert supported by a granule sandstone matrix (Fig. 10). These lithologies resemble those in the immediately underlying Cow Head Group. In the most northerly localities packages of green shale and siltstone occur which are up to 25m thick. Other facies include thin- to medium-bedded calcareous turbidites. Thin-bedded, fine-grained limestones are found interbedded with the sandstones as high as 250m above the base of the unit. Lateral continuity of beds of the Lower Head Formation is difficult to evaluate but at one locality a single conglomerate bed is traceable along strike for 1 km. The most northerly outcrops of the Lower Head Formation consist of a limestone conglomerate in a sandstone matrix, which form a resistant hill (Cawood and Williams, 1986). Approximately along strike of this on the shoreline, the sandstones contain horizons of limestone granules up to 20 cm thick. At this locality a folded olistrostromal horizon is present which is several tens of metres thick. It contains large boulders up to 5m in long diameter of conglomeratic sandstone in a pebbly mudstone matrix.

Paleocurrent indicators are few in the Lower Head Formation, but the few measurements taken indicate flow to the south or southwest.

The overall view of the Lower Head Formation in contrast to the Mainland Sandstone is that it is a unit dominated by thick to very thick bedded sands and gravels which are generally disorganized. Using the facies classification of Pickering et al. (1986) Facies B1.1 (thick-medium bedded disorganised sands) is the dominant facies with minor B2.1 parallel stratified sands. These facies are similar to facies present in the Mainland Sandstone, but they constitute a much higher percentage of the Lower Head Formation. Facies C2.2 showing classic Bouma sequences make up less than 5 per cent of the unit, but are more common in the finer grained calcareous beds. The conglomerates belong to class A1.2 or A1.4. Shales appear to be D2 organised silts, muddy silts and silt-mud couplets. All of these facies types are the product of turbidity currents and show many similarities to facies in the Tourelle Formation in Quebec (Hiscott, 1979; 1980) which has been interpreted as the product of deposition on a mid fan lobe of a small submarine fan.

Figure 9. Very thick bedded massive sandstones of the Lower Head Formation. White weathering patches are concretions around fluid escape pillars, and argillite granules.

Figure 10. Conglomerate horizon in the Lower Head Formation. Clasts are of limestone and chert.

DISCUSSION

It is clear that some of the detritus in all three Ordovician flysch formations can be directly related to immediately surrounding rock units. However other questions concerning sandstone provenance have arisen from this field investigation.

From the above descriptions it is evident that the coarsest autochthonous sandstones occur in the west and northwest, and thus there may be a fining eastwards of the autochthonous sands. This would raise a problem in that the source for these sandstones has been inferred to lie to the east (Stevens, 1970), and if this is the case sands of more distal aspect would more proximal to the source.

Stevens (1970) classified allochthonous siliclastic rocks in western Newfoundland into an upper quartzofeldspathic flysch and a lower quartzofeldspathic flysch. In fact lithologies in the category 2b rocks are so similar to those of category 1b that they are not easy to distinguish in the field. However rocks of categories 1 and 2 were supposedly deposited in radically different tectonic settings. In an accretionary prism setting such as is implied in Newfoundland, one would expect category 2 sandstones to be mineralogically immature — rich in feldspar and lithic fragments. For Quebec examples the quartz and feldspar is interpreted to have been derived from older sedimentary rocks (Hiscott et al., 1986). However in Newfoundland the possibility exists that the Long Range Complex which is now locally thrust over allochthonous sediments, could have been a morphological high during the Ordovician and could have provided primary detritus to the basins in which these sandstones were deposited.

ACKNOWLEDGMENTS

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Variations in structural style along the Long Range Front, western Newfoundland

Robert Grenier[1] and Peter A. Cawood[1]

Grenier, R. and Cawood, P.A., Variations in structural style along the Long Range Front, western Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 127-133, 1988.

Abstract

The Long Range Front is a weakly emergent thrust zone that brings Grenville basement above a Cambro-Ordovician carbonate sequence and locally over the Humber Arm Allochthon. Faults in this zone are interpreted as steep ramps above a flat sole thrust. Thrusting postdates Ordovician emplacement of the allochthon and is probably Devonian.

Faults within the structural front form a linked system. Along strike, variations in spacing and amount of displacement on faults has resulted in major variations in structural style. Between Bonne Bay and Portland Creek Pond, deformation is confined to a narrow zone involving two thrusts and an overturned footwall syncline. Between Portland Creek Pond and Hawkes Bay, thrusting is distributed over a wider zone marked by numerous high-angle reverse faults. Between Hawkes Bay and Ten Mile Lake, shortening is accommodated by displacement on two widely spaced thrusts.

Résumé

Le front de Long Range est une zone de chevauchement, faiblement émergente, qui amène le socle de Grenville au-dessus d'une séquence carbonatée datant du Cambrien et de l'Ordovicien et, par endroits, au-dessus de l'allochtone de Humber Arm (Taconique). On a interprété les failles présentes dans cette zone comme étant des pentes fortement inclinées reposant sur un chevauchement au plan inférieur plat. Le chevauchement a eu lieu après la mise en place de l'allochtone ordovicien et date probablement du Dévonien.

Les failles au sein du front structural formation réseau relié. Le long de la direction générale, des variations au niveau de l'espacement et de l'importance des déplacements associés aux failles se traduisent par des variations majeures au niveau du style structural. Entre la baie Bonne et l'étang de Portland Creek, une étroite zone de déformation englobe deux chevauchements, des plis larges et ouverts, et des synclinaux déversés. Entre l'étang de Portland Creek et la baie Hawkes, il existe une zone plus large de chevauchement comportant de nombreuses failles inverses fortement inclinées. Entre la baie Hawkes et l'étang Ten Mile, le rétrécissement est compensé par le déplacement qu'ont subi deux chevauchements largement espacés.

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INTRODUCTION

The Long Range Front of western Newfoundland forms a major thrust front bounding the western margin of the Long Range Inlier. The inlier consists of Precambrian rocks of the Grenville Orogen and represents the largest basement inlier within the Appalachian system. This report summarizes the results of field work carried out during the summers of 1986 and 1987 to delineate and describe the structural front along its entire 200-km length from Bonne Bay to Ten Mile Lake (Fig. 1).

A series of regional geological maps covering the area of the Long Range Front have recently been published (Knight, 1985, 1986a, b, c, d; Williams et al., 1984; Williams, 1985a; Williams et al., 1986; Cawood et al., 1987; Williams and Cawood, in press). These maps confirm the original work of Johnson (1941) who first pointed out that the western margin of the inlier was faulted (Long Range Thrust). These maps also provide the basis for the more detailed and focused work of the present study.

GENERAL GEOLOGY

Western Newfoundland consists of three distinct tectonic elements: Humber Arm Allochthon, latest Precambrian-Ordovician platformal sequence and Precambrian Long Range Complex. The Humber Arm Allochthon underlies the broad coastal lowland from Bonne Bay to Portland Creek Pond. The platformal sequence underlies the undulating hills farther east and the lowland to the north. Basement rocks are largely confined to the high ground of the Long Range Mountains. Outcrop throughout the western lowland is poor except for coastal and stream sections. Well exposed structural contacts are rare. Major faults are generally covered with boulders, glacial till, or series of long narrow lakes. Although often covered, the orientation of major faults can generally be inferred from associated small scale fractures developed in adjacent outcrops. Further constraint on fault orientation is provided by axial planar cleavage of thrust related, west-verging folds. Although rare, cleavage is developed in the vicinity of some of the major faults.

Precambrian granites and granitic gneisses of the Long Range Complex (Baird, 1960) are exposed in a number of structural inliers. The largest of these is the Long Range Inlier which extends along the entire eastern side of the study area, but basement rocks are also exposed in smaller inliers at the head of St. Pauls Inlet (Fig. 1A), along the northwestern margin of the Highlands of St. John, and at Ten Mile Lake (Fig. 1C).

A Cambrian-Ordovician shallow water platformal sequence unconformably overlies the Precambrian basement. The platformal sequence is composed of latest Precambrian to Middle Ordovician age carbonates and minor clastics. For this study the platformal sequence is divided into lower and upper units (Fig. 1 and 2). This division allows a clear depiction of fault offset and highlights the structural control on the distribution of rock units. The lower platformal unit consists of clastics and carbonates of the latest Precambrian to Lower Cambrian Labrador Group. The upper platformal unit consists of a sequence of Middle Cambrian dolostones of the Port au Port Group, Lower to Middle Ordovician dolostones and limestones of the St. George Group, and Middle Ordovician Table Head Group carbonates. Included in the upper platform division are clastics (Goose Tickle Formation and equivalent units) which overlie the Table Head Group and grade into a melange zone (Rocky Harbour melange, Williams et al., 1985) which marks the contact with the structurally overlying Humber Arm Allochthon.

The Humber Arm Allochthon consists of Middle Cambrian to Early Ordovician limestone breccias, ribbon limestone and shale of the Cow Head Group stratigraphically overlain by Middle Ordovician sandstones of the Lower Head Formation (James and Stevens, 1986). These units are arranged in a series of northeast-trending, east-dipping and east-facing thrust slices. Imbrication within the allochthon and its emplacement over time equivalent platformal strata took place during the Early-Middle Ordovician Taconian Orogeny.

AGE OF BASEMENT THRUSTING

The time of formation of the Long Range Front is poorly constrained. Grenville basement of the Long Range Inlier is thrust over both the platform sequence and the Humber Arm Allochthon. Thus, development of the structural front postdates the Taconian age for emplacement of the allochthon. Post-deformational cover rocks are absent from the region of the structural front, preventing a direct assessment of the upper age limit of deformation. However, little-deformed Carboniferous sediments onlap onto the southeast margin of the Long Range Inlier suggesting a pre-Carboniferous age for development of the structural front. Regional relationships provide a further constraint on the timing of deformation. Thrusting at the Long Range Front forms part of a widespread deformational event in western Newfoundland. In the Stephenville region, 125 km south of the study area, this event deforms the Siluro-Devonian Clam Bank Group but pre-dates Carboniferous sedimentation (Williams, 1985b). Thus, the deformational event in the Stephenville region, and probably also deformation at the Long Range Front, are correlated with the Devonian Acadian Orogeny.

STRUCTURAL STYLE OF THE LONG RANGE FRONT

The structural style of the Long Range Front varies along strike allowing division of the front into three structural zones (Fig. 1). The structural styles within these zones are outlined below from south to north.

Southern Zone

The southern zone extends from Bonne Bay to Portland Creek Pond (Fig. 1A and 2, cross-section XX'). Deformation is concentrated along a narrow belt delineated by the Long Range and Parsons Pond thrusts. The footwall sequence is folded into a regional asymmetric syncline with a steeply dipping to overturned eastern limb and a shallow-dipping western limb. Mesoscopic scale west-verging folds, commonly with an east-dipping axial planar cleavage, occur within the platformal carbonates in this zone. The Parsons Pond Thrust represents a splay off the main Long Range Thrust.

It disrupts the footwall sequence and brings platformal carbonates over the Humber Arm Allochthon. The Parsons Pond Thrust represents the exposed foreland extent of the thrust front in this region. However, broad-scale flexuring of the footwall sequence near the southern termination of the inlier may reflect an additional blind splay at depth.

The Long Range Thrust is exposed along West Brook, 4 km south of Parsons Pond (Fig. 3; Oxley, 1953; Williams et al., 1986). The fault is marked by a narrow, brittle fracture zone dipping 35 degrees southeast. Granite in the immediately overlying hanging-wall sequence is crushed, broken and chloritized and the footwall limestone is recrystallized and partially silicified.

At St. Pauls Inlet, platformal carbonates are thrust over a small inlier of Precambrian basement. Silicification has occurred along the fault contact which is marked by broken chloritzed granite and an overlying tectonic breccia composed of upper platformal carbonates. This contact can be traced north of the pond for about 1 km where it is delineated by a 10-m-wide quartz vein. Adjacent to the vein is a fault within the Precambrian basement which dips 30 degrees to the northeast.

Central Zone

The central section of the Long Range Front extends from Portland Creek Pond to the Highlands of St. John (Fig. 1B, 2; cross-section YY'). Deformation extends across the entire coastal lowlands and is characterized by thrusting in the area immediately adjacent to the Long Range Inlier and widespread high-angle faulting farther west. Bedding is largely subhorizontal and only gently folded about open upright structures. Small-scale folding is absent and cleavage, which is rare, was only observed adjacent to some of the major faults.

The Parsons Pond and Long Range thrusts continue northward to Blue Mountain. A number of minor splays occur within the platformal carbonates adjacent to these major structures. Dip of the Long Range Thrust and associated splays varies from around 70 degrees east of Blue Mountain to around 35 degrees just to the north (Fig. 1B). Slickensides on the latter fault surface pitch 45 degrees to the southwest, indicating an additional component of strike-slip movement.

West of the Long Range Thrust, numerous northeast-southwest trending high-angle reverse faults occur within the exposed upper platformal unit. These faults have disrupted the platform into a series of gently dipping, weakly folded blocks. Localized high-angle strike-slip faults trending east-west occur in the vicinity of Blue Mountain. The high-angle reverse faults die out to the east of Hawkes Bay. This corresponds with the commencement of the Ten Mile Lake Thrust.

Northern Zone

The northern zone extends from the southern end of the Highlands of St. John north to Ten Mile Lake (Fig. 1C, 2; cross-section ZZ'). The Long Range Structural Front within this zone is fairly broad and is marked by the Long Range and Ten Mile Lake thrusts. Deformation is largely concentrated along these fractures with little folding and faulting in the intervening rock sequences. Cleavage development in the platformal sequence is poor. Bedding is generally subhorizontal with the exception of near-vertical beds on the northern shore of Ten Mile Lake.

The Ten Mile Lake Thrust defines the western margin of the Highlands of St. John and results in thrusting of the lower platformal unit, and locally Grenville basement, over the upper platformal unit. The axis of a footwall syncline trends parallel to Ten Mile Lake. The dip of the fault varies from around 50 degrees at Ten Mile Lake to 80 degrees near its southern termination at Hawkes Bay. Compared to the southern and central zones, displacement on the Long Range Thrust within this northern zone is minor. At the Highlands of St. John, Grenville basement is thrust over the lower platformal unit and east of Ten Mile Lake the lower platformal unit thrust over itself. Small scale faulting within Grenville basement immediately adjacent to the inferred trace of the main thrust suggests a dip of around 30 degrees.

At Ten Mile Lake the Long Range and Ten Mile Lake thrusts merge into a single high-angle fault. Total displacement across the combined fault is only minor with the lower platformal unit in the hanging wall juxtaposed against the lower sections of the upper platformal unit in the footwall.

DISCUSSION

The Long Range Front marks the western limit of deformation associated with uplift and west-directed thrusting of the Long Range Inlier. Footwall collapse during overthrusting of this large basement massif led to the development of a series of foreland directed thrusts which share a common subhorizontal basal detachment. Thus, these thrusts form a linked system (cf. Dahlstrom, 1970). Variation in the relative amount of displacement and spacing of the thrust planes has lead to major along strike variations in structural style.

In the southern zone, deformation is concentrated along the Long Range Thrust, although at least one splay off this fracture, the Parsons Pond Thrust, developed in the immediately adjacent footwall sequence. Concentration of deformation over this narrow zone was associated with folding of the footwall sequence into a regional asymmetric syncline. Relatively widespread mesoscopic folding also developed in this footwall sequence.

In the northern zone, the structural front is considerably wider extending between the Long Range and Ten Mile Lake Thrusts. North of St. Margaret Bay deformation is distributed over an even broader zone and involves an additional thrust lying on the foreland side of Ten Mile Lake. A consequence of this widening of the deformational front is that folding of the sedimentary sequence is limited to broad regional warping. Maximum displacement along the thrust front in the northern zone is displaced towards the foreland. Offset across the Long Range Thrust in this zone is only relatively minor with most of the regional foreshortening taken up along the Ten Mile Lake Thrust. However, displacement across the latter fracture decreases toward the northern end of the zone where it merges with the Long Range Thrust. In this region there is a further foreland-directed transfer of the site of maximum displacement to the thrust lying north of St. Margaret Bay. Thus, there is an overall foreland directed stepping out of the deformational front in this northern zone.

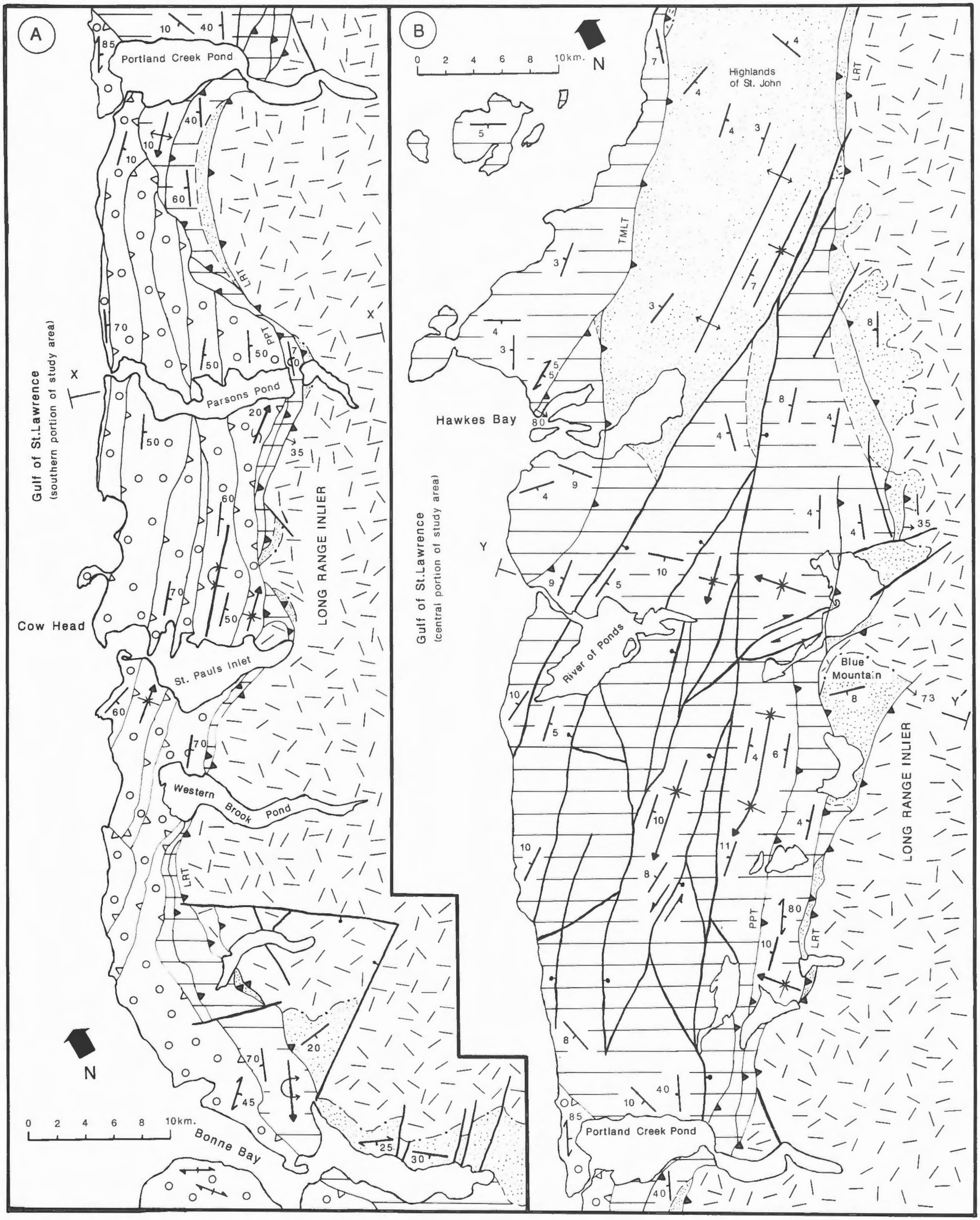
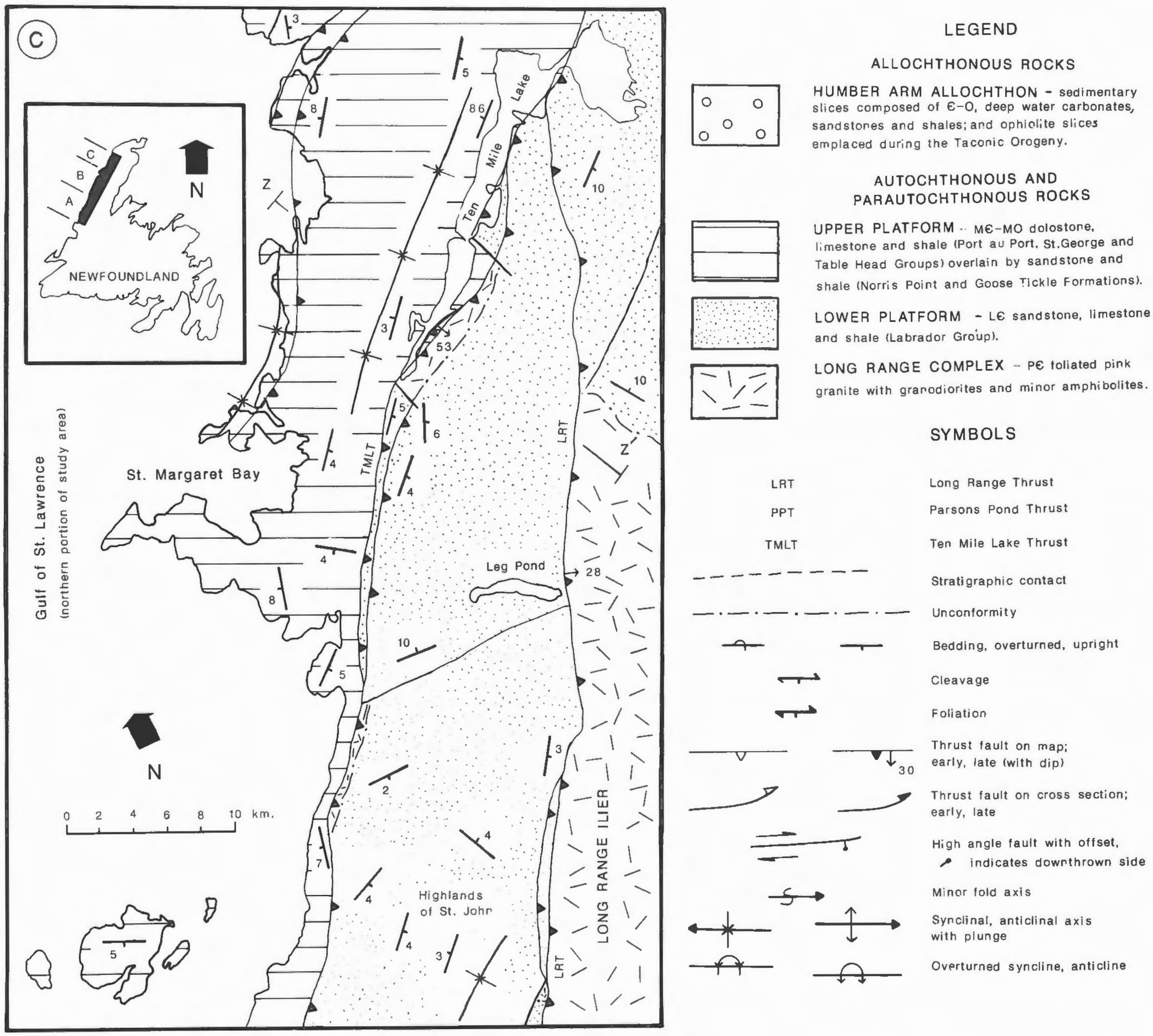


Figure 1. Simplified geological map of the Long Range Front; A) southern, B) central and C) northern portions of the study area. This map is based on field work by the authors and a compilation of previous work referred to in the text.

The relatively complex structural style of the central zone with its numerous high-angle reverse faults reflects its intermediate structural position between the contrasting southern and northern zones. The region acted as a transfer zone between the concentrated, narrow thrust front of the southern zone and the more dispersed, broader thrust front of the northern zone. Shortening occurs by small displacements on numerous high-angle faults (Fig. 2, cross-section YY')

The northern and southern terminations of the Long Range Inlier are marked by Grenville basement plunging to the north and south, respectively, below the Paleozoic cover sequence (Williams et al., 1984; Bostock, 1983). The northern termination reflects the northward decrease in displacement on the Long Range Thrust, and hence, the northward decrease in the amount of basement uplift. The southern termination corresponds with a major reversal in structural vergence from west-directed along the Long Range Front to east-directed south of the Bonne Bay area which is characterized by east-verging folds and thrusts (Waldron, 1985; Bosworth, 1985; Williams and Cawood, 1986). Reversal in structural vergence reflects the change from an emergent thrust front along the western margin of the Long Range Inlier to a buried thrust front farther south (Morley, 1986). This reversal may be accommodated along an east-west strike slip fault located in the subsurface east of Bonne Bay.

The preservation of cover sequences on Grenville basement at a number of localities within the Long Range Inlier (Knight, 1985, 1986a, b; Williams, 1985a; Williams et al., 1985; Cawood et al., 1987) provide a direct tie with stratigraphic units in the footwall sequence and suggest a maxi-



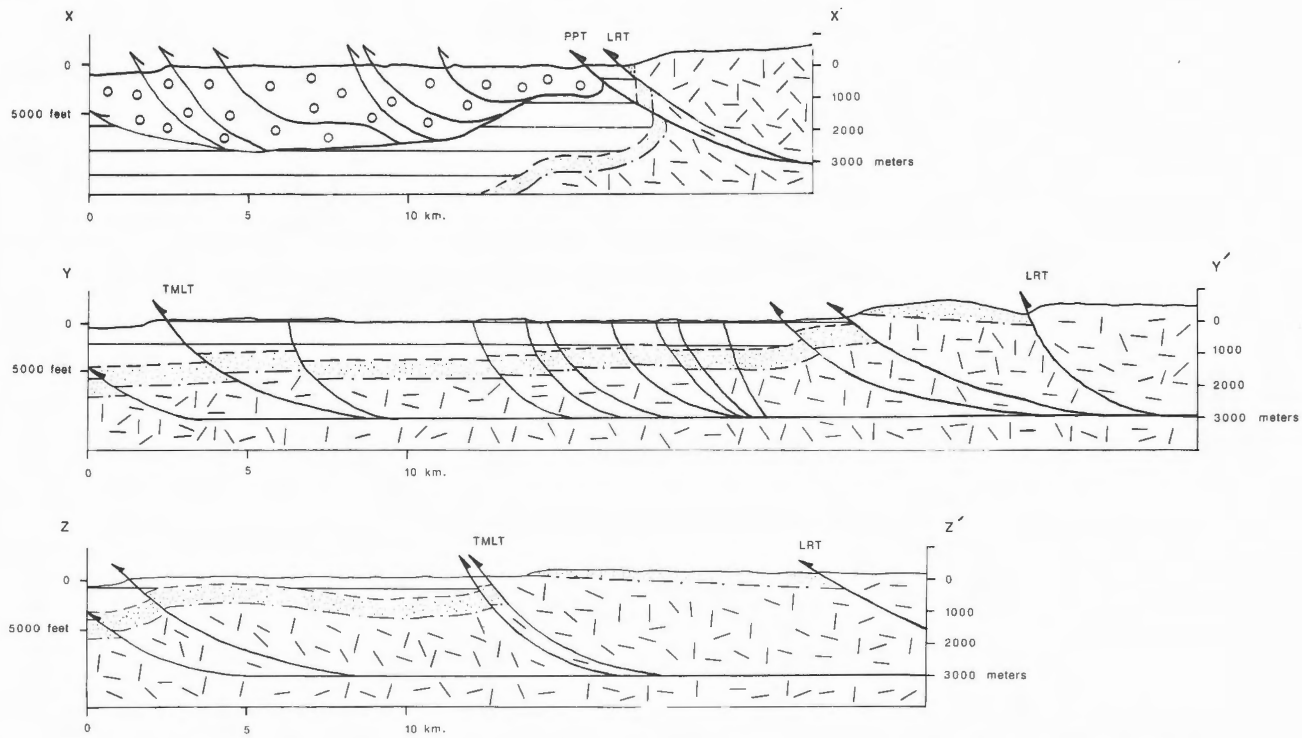


Figure 2. Simplified structural cross-sections for Figure 1. XX' corresponds to XX' in Figure 1A. Faults in the Humber Arm Allochthon are due to imbrication during transport and emplacement during the Taconian Orogeny. YY' corresponds to YY' in Figure 1B. ZZ' corresponds to ZZ' in Figure 1C.

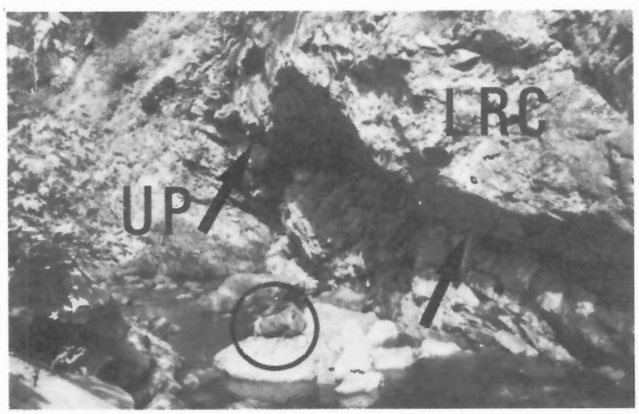


Figure 3. The Long Range Complex (LRC) is thrust over Middle Ordovician limestones of the upper platformal unit (UP) in West Brook, 4 km south of Parsons Pond. Arrows point to the Long Range Thrust which dips 35 degrees to the southeast. Backpack (circled) is for scale.

mum lateral transport of the inlier of only a few kilometres (Fig. 2; Cawood and Williams, 1986).

In summary, the Long Range Front represents a narrow, weakly emergent thrust front (Morley, 1986) dominated by the Long Range Thrust and associated splays with only minor folding. The hanging wall sequence consists of a single large and internally coherent block of Grenville basement. This is indicative of abrupt abandonment of the sole thrust and its sharp rapid ramping to the surface with few intervening flats.

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Evidence for D_1-related thrusting and folding in the Bathurst-Millstream River area, New Brunswick[1]

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van Staal, C.R., Winchester, J., and Cullen, R., Evidence for D_1-related thrusting and folding in the Bathurst-Millstream River area, New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 135-148, 1988.

Abstract

The Tetagouche Group rocks exposed north of Bathurst were previously interpreted as a north younging homocline. Detailed mapping and new fossil discoveries this summer show a complex repetition and inversion of stratigraphy, which is explained by earliest D_1- related, east or southeast directed thrusting and folding (F_{1A}). A large downward facing F_2 fold in the footwall of the Bathurst thrust fault is explained as the result of refolding the overturned limb of a large F_{1B} fold. The presence of another generation of F_1 folds (F_{1B}) that refold the Bathurst thrust fault rather than invoking a giant F_1 fold nappe, is necessary to account for the apparently consistent facing direction towards the north or west of the footwall rocks of the Bathurst-thrust fault for tens of kilometres along strike. The thrusts, F_{1A} and F_{1B} folds are interpreted as the result of one progressive deformation (D_1).

Résumé

On avait auparavant interprété les roches du groupe de Tétagouche exposées au nord de Bathurst comme constituant un homoclinal dont les strates deviennent progressivement plus jeunes vers le nord. Les travaux de cartographie détaillés et la découverte de nouveaux fossiles durant l'été, indiquent une répétition complexe et une inversion de la stratigraphie, qui s'expliquent à l'aide des plus anciens chevauchements et plissements (F_{1A}) dirigés vers l'est ou le sud-est, liés à la déformation D_1. Un grand pli F_2, orienté vers le bas, dans la paroi de la faille chevauchante de Bathurst résulte d'un nouveau plissement du flanc inverse d'un grand pli F_{1B}. La présence d'une autre génération de plis F_1 (F_{1B}) qui ont à nouveau plissé la faille chevauchante de Bathurst, plutôt que d'une nappe de charriage géante F_1, peut seule expliquer l'orientation apparemment cohérente vers le nord ou l'ouest, sur des dizaines de kilomètres en direction de la paroi, le long de la faille chevauchante de Bathurst. On a interprété les chevauchements et les plis F_{1A} et F_{1B} comme étant le résultat d'une seule déformation progressive (D_1).

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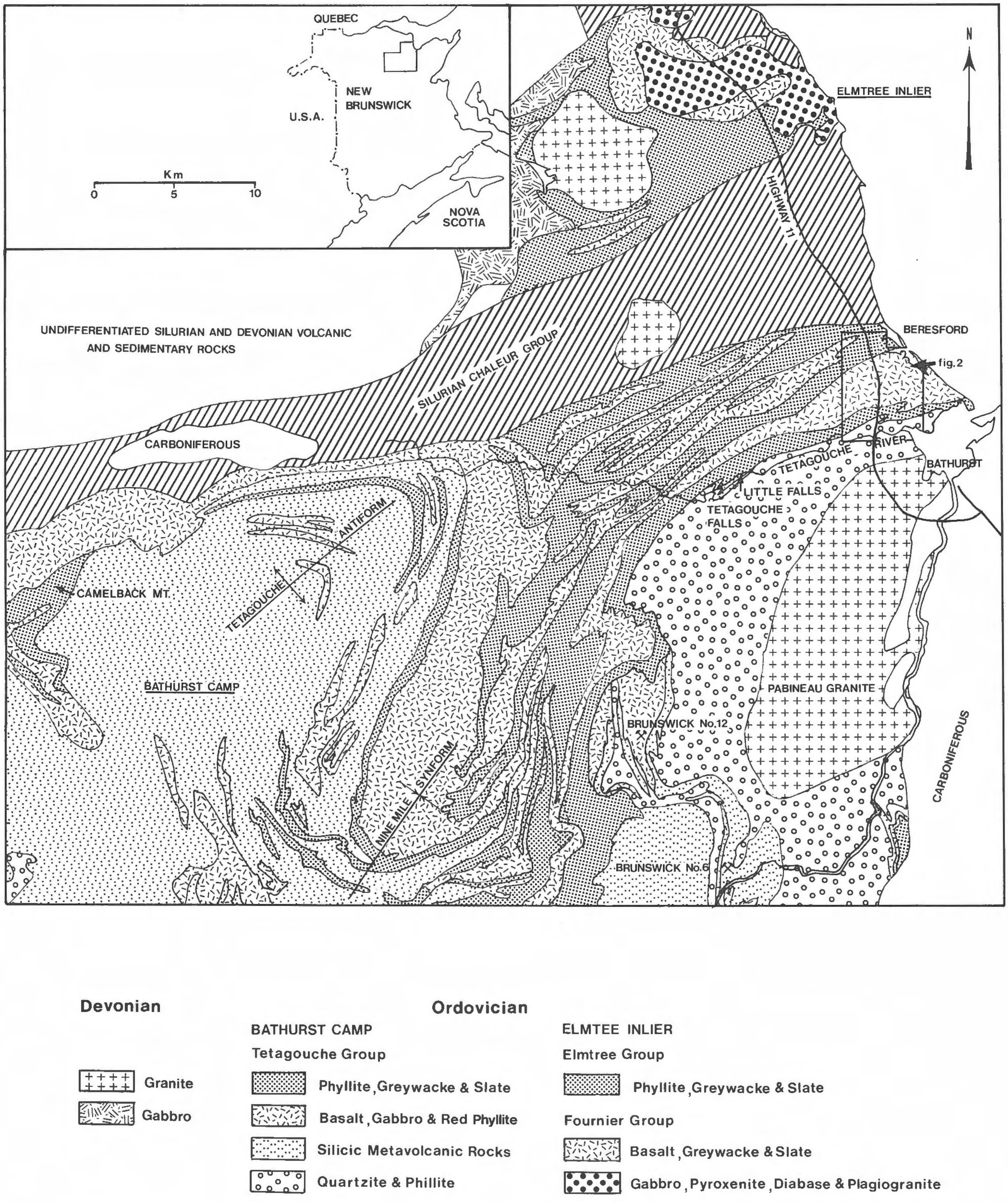


Figure 1. Geological map of the northern part of the Bathurst Camp and Elmtree Inlier in northern New Brunswick. Geology adapted from Davies (1979) and van Staal (1987). Outlined area in the northeastern part of the Bathurst Camp, north of Bathurst, is area discussed in this paper.

INTRODUCTION

Scarcity of fossils, sparse exposure and complex structural history generally prevent satisfactory understanding of the geometry and detailed stratigraphy of the Cambro- Ordovician Tetagouche Group in northern New Brunswick. Local stratigraphic interpretation commonly relies heavily on the superposition principle. However, two generations of tight to isoclinal folds and the presence of early, F_1 related thrusts and/or slides (van Staal and Williams, 1984, 1986; van Staal, 1986, 1987) show the danger of the use of the superposition principle in this area. One important example of the application of this principle, with regional implications, is the rock section exposed in the Tetagouche River (Young, 1911) and along Highway 11 north of Bathurst (Rast and Stringer, 1980). This area (Fig. 1) is purported to show an almost complete section of the Tetagouche Group (Fyffe and Noble, 1985; Fyffe, 1987). The sequence contains black graphitic shales with graptolites characteristic of the Dicranograptus clingani Zone (Dean in Skinner, 1974, p. 43; Fyffe and Noble, 1985; J. Riva, pers. comm., 1987) of the late Caradocian, and represents an important marker unit.

Based on a few younging indications and the superposition principle, the sequence exposed in this area (Fig. 2) was previously interpreted as a more or less north-facing homocline (Rast and Stringer, 1980; Fyffe, 1982, 1987). This has important implications since the basalts that structurally occur above the Late Caradocian black shales should then be younger; contrary to the interpretation of Skinner (1953, 1974) who mapped the presence of a large anticline with the basalts in the core. Pajari et al. (1977) and Rast and Stringer 1980) went even further and assumed the existence of an angular unconformity, marking the classical Taconian orogeny, at the base of these basalts. Following McBride (1976), the evidence for the existence of this unconformity was mainly based on a supposedly lesser degree of deformation present in the rocks north of this unconformity, i.e. F_1 is absent in these rocks.

On the other hand, van Staal (1986,1987) interpreted the contact between the main package of basalts and associated sediments of the upper part of the Tetagouche Group and the sediments and silicic volcanics of the lower part as an F_1 related thrust. These various interpretations were critically investigated and tested during the course of our regional mapping project. Mapping and structural analysis was facilitated during the 1987 summer by the drought and low water levels in the rivers and brooks which locally provided almost continuous stream-bed outcrop (e.g. Grants Brook) and numerous new exposures.

Lithologies and facing of the geological domains

The area discussed in this paper (Fig. 2) is divided into four geological domains which are described below in order of occurrence from south to north.

Bathurst Domain

The oldest part of the Tetagouche Group exposed in the Tetagouche River near Bathurst consists of a sequence of siltstones, shales and quartz wackes that were transformed into phyllites and quartzites during regional metamorphism and deformation. These psammitic and semi-pelitic rocks occur throughout the Miramichi Zone (Fyffe, 1982; Fyffe et al., 1983) and form part of the Nepisiquit formation.

The phyllites are commonly reddish whereas the quartzite beds are dominantly green. The red is generally discontinuous and irregular and the colour boundaries cut across bedding and all other structures. The red is therefore probably due to percolation from the locally overlying Carboniferous red beds of the Bathurst Formation (Young, 1911; Skinner, 1974). The quartzite beds, reach thicknesses of 2 m, and are locally feldspathic. The thick quartzite beds locally cut down through into silty and finer grained phyllites, suggesting that they represent, at least in part, channel deposits. This part of the Nepisiquit formation correlates best with unit 2 of Helmstaedt (1971).

The quartzite-phyllites of the Nepisiquit formation are stratigraphically overlain by a sequence of thin bedded calcareous siltstones, mudstones and minor crossbedded limestone, which are capped by a light green arenaceous or rudaceous limestone. These rocks are mapped as the Vallée Lourdes formation, named after the area enclosed by the big bend in the Tetagouche River in north Bathurst (Fig. 2). The rudaceous limestone is represented mainly by a polymictic conglomerate (Young, 1911) in a calcareous matrix. The pebbles were derived from the underlying calcareous siltstones and mudstones and fine- to medium-grained biotite-bearing granitoids. Granitoid pebbles are as large as 60 cm in diameter. No indication of predepositional deformation structures was found in any of the clasts. The arenaceous limestone contains numerous quartz crystal fragments.

We agree with Fyffe and Noble(1985) and Fyffe(1987) who correlated these rocks with lithologically similar rocks that occur along strike 10km to the west at Little Falls and Tetagouche Falls (Fig. 1), which contain brachiopods and conodonts of Late Arenigian - Early Llanvirnian age (Fyffe, 1976; Nowlan, 1981; Neuman, 1984).

Younging indicators are generally rare or absent in most of the metasediments in part owing to the strong structural overprint. Even where they are present, the younging directions change across the outcrop due to the complex folding. When the geometry of this complex folding cannot be worked out because of discontinuous outcrop or an advanced stage of transposition, younging indicators prove to be useless. However, crossbedding and grading in the calcareous siltstones and limestones as well as fragments of the latter in the rudaceous limestone at the top of the Vallée Lourdes formation (Fig. 2) indicate that this unit is consistently north facing and support a general younging of the Bathurst Domain towards the north (Fyffe, 1982; Fyffe and Noble, 1985). This strengthens the earlier assumptions that the Nepisiquit formation is Early Ordovician or older (Helmstaedt, 1971; Fyffe et al., 1983; van Staal, 1987). Mesoscopic folding is rarely observed in this unit, despite the presence of 4 generations of foliations, i.e. S_1, S_2, S_3, and S_4 (see below). Apparently this rock unit deformed more homogeneously than the surrounding rocks, possibly because of a lower competency contrast between the beds as is indicated by the absence of boudinage typically present in the other metasedimentary units. Scarcity of mesoscopic folds probably led Young (1911) to speculate that these calcareous siltstones and mudstones could represent post-Ordovician, i.e. Silurian, strata infolded with the Tetagouche Series.

Beresford Domain

The formations of the Bathurst Domain are followed to the north by late Caradocian, graphitic black shales, siltstones and cherts that were transformed into black phyllites and chert (Young, 1911; Skinner, 1953). This black phyllite unit lies conformably above red and green phyllites. The outcrop pattern of the red and black phyllites with respect to the basalts (Fig. 2), indicates the presence of a large Z-shaped fold (see below and Fig. 2).

Graptolites discovered this summer in the black phyllite unit on the Tetagouche River (GSC Loc. 103032), south of

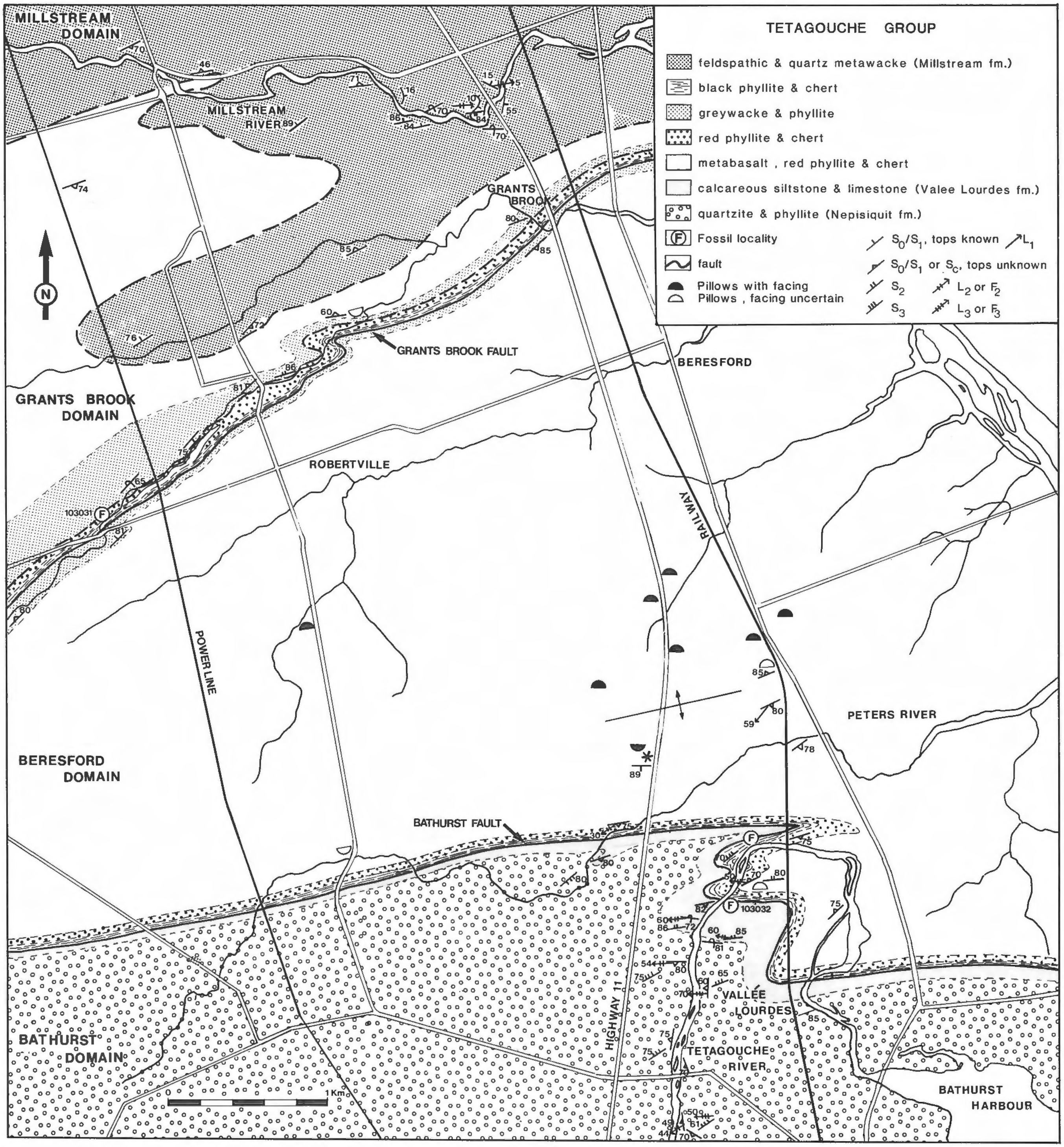


Figure 2. Geological map of the Bathurst-Millstream River area. Asterisk marks first outcrop of basalt along Highway 11 discussed in text. These basalts young towards the south.

the well known railway bridge locality, confirm the late Caradocian age (J. Riva, written commun., 1987). The red and green phyllites form part of the ferromanganiferous red and green shales, siltstones and cherts that are characteristic of the Tetagouche Group in the northern part of the Miramichi Zone (Young, 1911; Helmstaedt, 1973; Fyffe, 1982; van Staal, 1986). The phyllites are interbedded with mafic volcanics. The mafic volcanics are a heterogeneous mixture of vesicular pillow basalts, plagioclase-phyric basaltic dykes and sills and locally pillow breccias. The pillow basalts, characteristically green or bluish green, predominate. White, red, and turquoise interpillow chert is common and is locally mineralized by chalcopyrite and pyrite. The basaltic dykes grade locally into the pillow lavas indicating that they are feeders to the lavas. Plagioclase phenocrysts have been altered to albite, epidote and bluish green pumpellyite pseudomorphs.

The pillowed basalts exposed along Highway 11 and parallel country roads north of the Tetagouche River in the Beresford Domain (Fig. 2) were interpreted by Skinner (1974) as occupying the core of a large anticline. On the basis of this simple structural interpretation and assuming a simple deformation history, the basalts were thought to underlie and consequently be older, than the late Caradocian black phyllites. Rast and Stringer (1980) and Fyffe (1982) refuted Skinner's interpretation on the basis of their observation that all pillows young northwards. We agree in general with the observations of Rast and Stringer and Fyffe that the younging is northwards in the majority of basalt outcrops where tops can be determined.

However, we stress that outcrop is far from continuous and there is sufficient room (Fig. 2) for numerous younging reversals due to isoclinal folding. This is particularly pertinent if the mesoscopic slides seen at the outcrop scale (Fig. 3) are also present at the macroscopic scale. That such speculation is justified is illustrated by the first outcrop of basalts along Highway 11, north of Bathurst, which youngs dominantly towards the south. The pillowed flows, locally vesicular, show chilled margins on both top and bottom of the flows but large mudflame structures occur only on one side of the flows, where the hot pillows flowed into cold and unconsolidated mud. These mudflames suggest a southwards younging direction. Grading and channelling in the interlayered red phyllites confirm that younging is to the south. Thus these sedimentary structures represent important evidence that refutes the consistent north younging homoclinal stratigraphy of Rast and Stringer (1980) and Fyffe (1982).

The pillow basalts and red and green shales, siltstones and cherts are interbedded throughout the northern Miramichi Zone. The sedimentary rocks are therefore interpreted as equivalents of the red Fe/Mn-rich sediments commonly found in the upper part of oceanic crust, since the basalts resemble ocean floor basalts chemically (Whitehead and Goodfellow, 1978; van Staal, 1987; Winchester and van Staal, unpubl. data). The age of the basalts is so far only constrained by conodonts from interbedded limestone at Camel Back Mountain (Fig. 1). The conodonts indicate an early to middle Caradocian age equivalent to the upper parts of the Nemagraptus gracilis Zone or the lower part of the Diplograptus multidens Zone (Nowlan, 1981). The basalts can be traced as one continuous package around the Tetagouche antiform and Nine Mile synform pair to Bathurst. This suggests that the stratigraphy of the rocks immediately to the north of the Bathurst Domain faces southward, opposite to the general north facing of the Bathurst Domain, since the black shales contain graptolites of the late Caradocian Dicranograptus clingani Zone (J. Riva, written commun., 1987). North of the south-facing basalts the younging direction changes to the north indicating the presence of an anticline in the basalts, as previously suggested by Skinner (1974). The basalts are bounded to the north by thin bedded greyish green greywacke and phyllite, although the contact between the greywacke, and basalt is not exposed. At present this contact is assumed to be conformable, which suggests that the greywacke and phyllite unit is younger than the basalts. If this interpretation is correct the evidence for an anticline within the Beresford Domain implies that this unit should pinch out along strike since it is absent on the southern side of this domain.

Grants Brook Domain

A similar rock sequence to that found in the Beresford Domain occurs in the Grants Brook Domain (Fig. 2). Graphitic and graptolite-bearing black shales, siltstones and cherts are bounded to the north by red Fe/Mn-rich shales and cherts, which are interbedded with basalt or a fine grained basaltic sill close to the intersection between Grants Brook and the powerline (Fig. 2). A few graptolites recovered this summer from the black phyllites (GSC loc. 103031) do not allow a definite assignment to a specific zone although they indicate a Caradocian age, possibly older than the black phyllites in the Beresford Domain (J. Riva, pers. comm., 1987).

The sequence from south to north of black phyllite to red phyllite and chert (see Fig. 2) precludes an interpretation that the Grants Brook Domain is simply a repetition of the Beresford Domain by large scale anticlinal folding as proposed by Skinner (1974). Between this package of rocks and the basalts of the Beresford Domain (Fig. 2) there is a narrow band of thin bedded, greenish grey tuffaceous greywacke, and grey phyllite. Lithologically similar rocks, but here also including conglomerate beds and dark grey to black phyllites, occur structurally above the red phyllite-chert unit. The greywacke- phyllite unit is in turn overlain by fine grained basalts and pillow breccias.

The general facing of the Grants Brook Domain is impossible to determine from younging directions. Although individual beds locally show good grading the deformation in the black phyllites and greywacke and phyllites is generally very high. Individual beds were disrupted and intensely boudinaged accompanying isoclinal F_1 folding (see below) such that lenticular fragments of competent greywacke beds float in a matrix of grey-black phyllonite. However by analogy with the Beresford Domain, the sequence from south to north of black phyllite to red phyllite (Fig. 2) suggests a general facing towards the south in the southernmost part of the Grants Brook Domain. The alternative interpretation that the black phyllite-chert unit is older than the red phyllite unit in this domain is at present rejected. Because these two units are regionally very extensive, this interpretation would imply an alternating deposition of red and black shales, which is thought unlikely considering their markedly different state of oxidation.

Millstream Domain

The mafic volcanics of the Grants Brook Domain are bounded to the north by rocks of the Millstream formation. The Millstream formation consists mainly of alternating greywacke or sandstone with minor greenish grey slate. Locally this formation also includes thin pink felsite sills. The greywackes are feldspathic, locally arkosic, and contain lithic clasts, calcareous concretions and detrital muscovite low in the succession, but grade up into quartz wacke. The greywacke beds are commonly, though not invariably, graded and thick compared to the slate beds. In places the slate beds

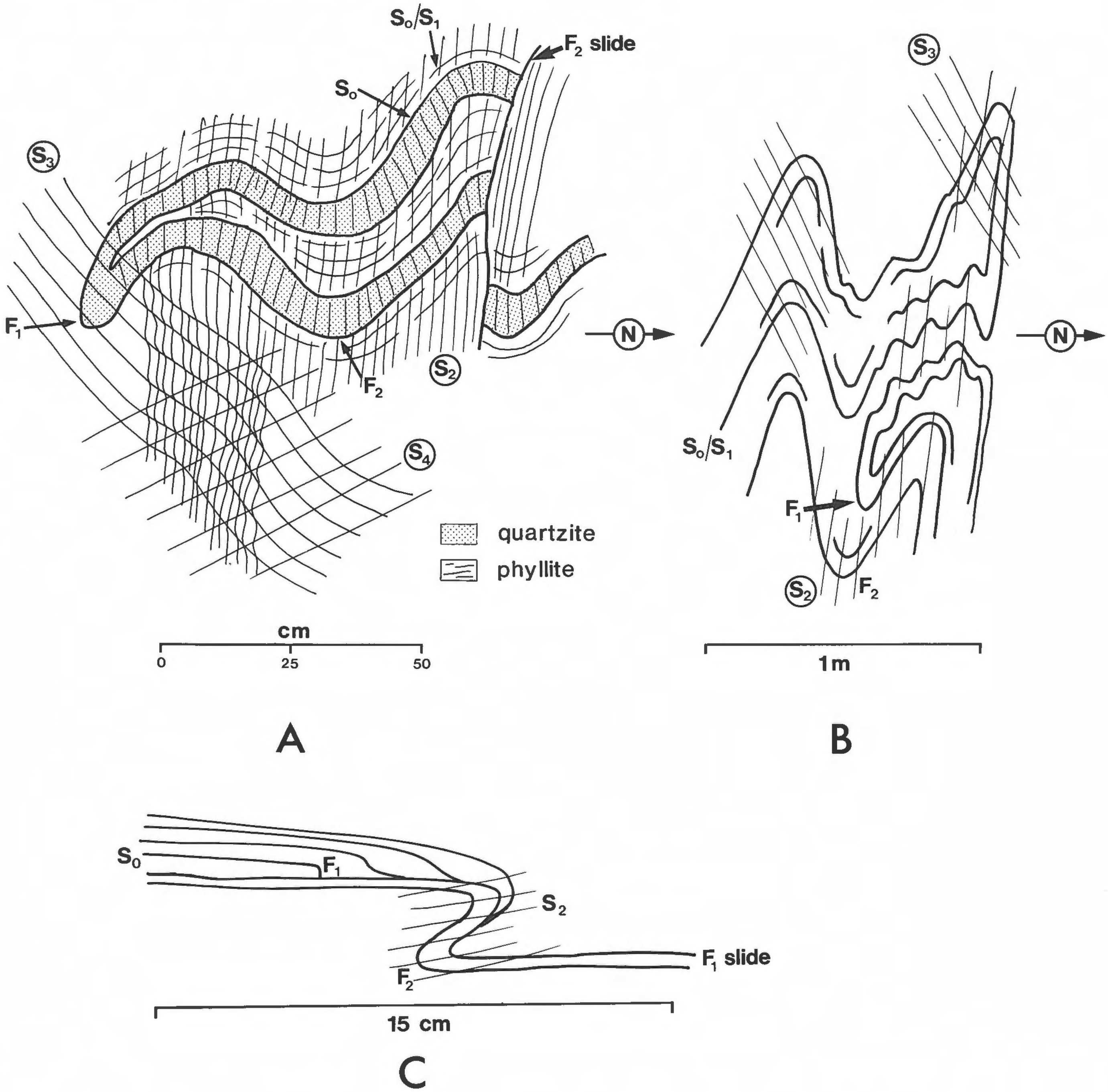


Figure 3. Field sketches showing overprinting relationships. A, in quartzite-phyllite unit on the Tetagouche River. B, in red phyllites in Grants Brook. C, F_2-refolded F_1-thrust in greywacke-phyllite unit exposed to the west of the area discussed in this paper.

are missing and the greywacke beds are amalgamated. Cross-bedding, rip up clasts and bottom structures are common in these rhythmites, which may represent turbidite deposits. These sedimentary structures suggest that this domain has a general facing towards the north.

The age of the Millstream formation is problematic. Lithologically similar rocks occur in the Elmtree inlier (see also Rast and Stringer, 1980) where they are part of the Pointe Verte Formation (Fyffe, 1985). If this correlation is correct, it suggest that the Elmtree inlier and Bathurst area rocks are connected in the subsurface beneath the Silurian Chaleur Group (Fig. 1).

The Pointe Verte Formation alkali-basalts have been dated (conodonts in interpillow limestone) as late Llandeilian-early Caradocian (Nowlan, 1981; Fyffe, 1986) but fossils have not yet been found in the greywackes included in this formation. The precise age of the latter is thus still unconstrained, particularly since the contact between the basalts and greywackes is probably tectonic (Davies et al., 1983; van Staal and Langton, unpubl. results).

STRUCTURAL HISTORY

The structural history of the Tetagouche Group is polyphase and complex. Detailed descriptions of the relative structural succession have been given elsewhere (e.g. van Staal and Williams, 1984; van Staal, 1986, 1987). Nevertheless the respective structures present in this area will be described briefly below.

On basis of overprinting relationships (Fig. 3), at least four conspicious generations of structures have been recognized in this area. The first group of structures are represented by a bedding parallel foliation (S_o/S_1), stretching lineation (L_1) and isoclinal folds (F_1). The second group of structures are a finely spaced crenulation cleavage (S_2) and tight to isoclinal folds (F_2). Discontinuities, i.e. slides, have locally developed along the axial planes of small scale F_1 and F_2 folds (Fig. 3). There is a style overlap between F_1 and F_2 folds (van Staal, 1986) and where overprinting relationships are not well defined or missing, these two groups cannot everywhere be distinguished from one another with certainty.

S_2 is parallel or subparallel to S_o/S_1 along the limbs of F_2 folds and is therefore commonly difficult to separate from the earlier foliation(s). The regional foliation on the limbs of large F_2 folds is therefore generally a composite structure, mapped as S_C.

F_2 folds and S_2 are overprinted by a differentiated crenulation cleavage (S_3), axial planar to open to tight folds (F_3). A characteristically strongly developed crenulation lineation (L_3) on S_1, S_2, or S_C as well the rather consistent strike of S_3 between northeast and east (Fig. 4) in this area make this group of structures commonly easy to recognize although S_3 closely resembles S_4 in style locally. F_1, F_2 and F_3 folds vary in size from millimetre to kilometre scale. The last conspicuous structure is a penetrative, generally widely spaced fracture-crenulation cleavage that is axial planar to open kinks and folds. S_4 generally has a strike between northwest and north (Fig. 4) in this area, although a conjugate set of kinkbands or fracture cleavage with an east-trend is locally present.

Open, centimetre scale, recumbent F_5 kinks or folds (van Staal and Williams, 1984) have only been observed at a few places due to the lack of three dimensional outcrop. These structures are, however, abundant farther west along the Tetagouche River, in high, precipitous cliffs. S_5 generally dips shallowly towards the southeast.

Assessment of the existence of an unconformity in the Tetagouche Group

McBride (1976), Pajari et al. (1977) and Rast and Stringer (1980) claimed that the rocks of the upper part of the Tetagouche Group exposed in the Grants Brook and Beresford domains are less deformed than the lower part of the Tetagouche Group, i.e. the Bathurst Domain. This claim, which represents the main evidence for the existence of an angular unconformity between the lower and upper parts of the Tetagouche Group, i.e. the Nepisiquit and Vallée Lourdes formations versus the rocks exposed in the Beresford, Grants Brook and Millstream domains, implies a distinct episode of pre-late Ordovician deformation. Moreover, it also implies significant volcanism in post-late Caradocian times, a feature not seen in similar rocks in the Dunnage Zone of Newfoundland (e.g. see Dunning et al., 1987).

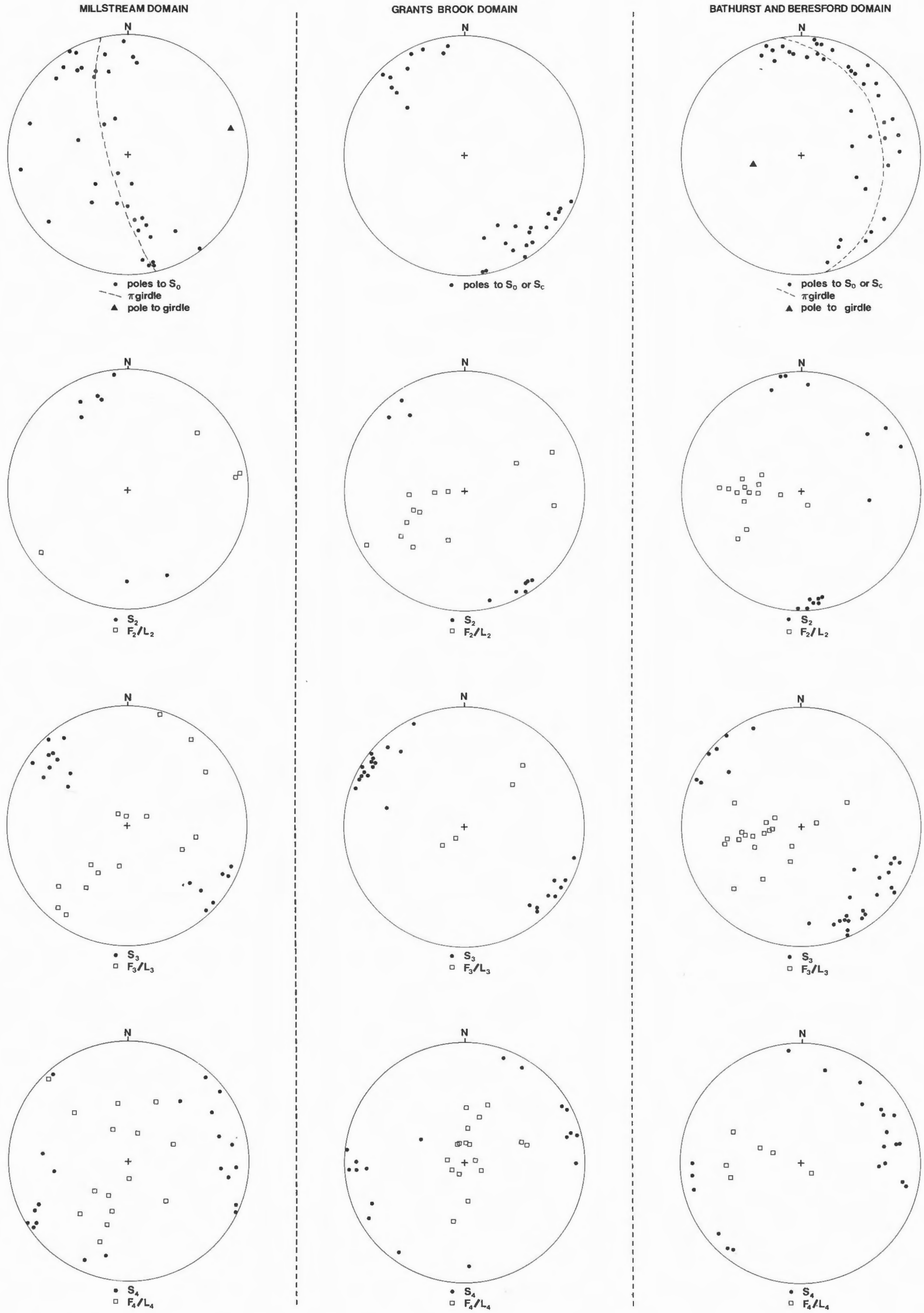
Two lines of evidence overrule the existence of this angular unconformity:

1) Structural analysis showed that the structural succession is identical below and above this boundary between the Bathurst and Beresford domains (Fig. 3, 4 and 5). The amount of strain recorded in the rocks, admittedly varies considerably, but this is a feature generally typical of the Tetagouche Group (van Staal, 1986). Deformation is obviously highly heterogeneous and concentrated in relatively soft rocks, in this case phyllosilicate- rich sediments and volcanics.

2) Graptolites indicate that the Grants Brook black phyllites and cherts are at least coeval or possibly older (J. Riva, pers. comm., 1987), but certainly not younger than the black phyllites and chert in the Beresford Domain as is necessary with a simple north facing homocline above an unconformity.

Geometry of the Bathurst and Beresford domains

The spatial distribution of the black shales, red and green shales and basalts along the northern part of the Tetagouche River and Peters River in the Beresford Domain suggest the presence of a large Z-shaped fold (Fig. 2). The basalts exposed in the Tetagouche River are therefore better explained as being continuous with the lithologically similar basalts occurring along Highway 11, and not as a separate basalt lens as previously interpreted (Davies, 1976; Fyffe and Noble, 1985). This interpretation is supported by the absence of these basalts in the part of the Bathurst Domain shown in Figure 2. Analysis of the outcrop of basalt shown on the map of Davies (1976) in the Bathurst Domain shows it to be a small andesitic dyke intruding the black shales, possessing distinctly different chemistry than the basalts. The relative age of the large Z-shaped fold is interpretative since the hinges are not


MILLSTREAM DOMAIN
GRANTS BROOK DOMAIN
BATHURST AND BERESFORD DOMAIN
N
poles to S_0
π girdle
pole to girdle
poles to S_0 or S_c
S_2
F_2/L_2
S_3
F_3/L_3
S_4
F_4/L_4

well exposed. Structural analysis indicates that this fold predates F_3, since S_3 maintains the same relationships to S_o/S_1 or S_C on both limbs of this structure, i.e. S_3 cuts across both limbs (Fig. 2). The asymmetries of mesoscopic F_2 folds, S_2-S_o/S_1 relationships (Fig. 2) as well as the coincidence in general orientation between S_2 and the trend of the axial plane of this fold are consistent with a formation during F_2. However, the sparse data do not completely rule out the possibility of an F_1-age. An identical structure that occurs at Tetagouche Falls (see Fig. 8 of Fyffe and Noble, 1985 or Fig. 2b of Davies et al., 1985) is probably also a large F_2 fold (van Staal, unpub. results). If the F_2 interpretation is correct, this large Z-shaped fold is probably downward facing (Shackleton, 1958) where it involves the formations of the Bathurst Domain since the S_2 - S_o bedding intersection as well as the plunge of small scale F_2 folds indicate a westerly plunge (Fig. 4), whereas the general facing of the Bathurst Domain is to the north (see above). In fact, the orientation of S_2, with respect to S_o and younging in the cross-bedded calcareous siltstones and limestone of the Vallée Lourdes formation show consistently downward-facing relationships (Fig 2). Form surface mapping and the relationships between S_o/S_1,S_2 and S_3 in the Tetagouche River suggest that the F_2 and older structures are refolded by a large NE-trending F_3-fold (see Fig. 2, 4 and 8).

Another large F_2 fold has been mapped in the eastern part of the Grants Brook Domain (Fig. 2), using the black and red phyllites as marker horizons. The hinge of this fold is not exposed and there are alternative ways to explain the outcrop pattern. However, a fold is the most simple explanation and most consistent with available data. This structure is a S-shaped fold on the surface but its plunge remains indeterminable. However, most small scale F_2 folds and L_2 intersection lineations on the limbs of this stucture plunge generally to the west (Fig. 4), suggesting a westerly plunge for this fold as well.

Figure 5. Isoclinal F_1 fold in psammitic layer in red phyllite unit along Grants Brook. Note how S_2 and S_3 cut across both limbs of F_1. In fact on the scale of the outcrop, F_1 is clearly refolded by F_2. Strike of S_2 is ENE-WSW.

Interpretation of the contacts between the various geological domains

Mapping shows that the calcareous rocks of the Vallée Lourdes formation of probable late Arenigian – early Llanvirnian age (see above) are juxtaposed directly with highly strained late Caradocian black shales and cherts of the Beresford Domain (Fig. 2). Thus, Middle Ordovician rocks, which usually consist of volcanic rocks, appear to be absent (Fyffe, 1982; van Staal, 1987). Also the Bathurst Domain and the southern part of the Beresford Domain have opposite facing directions.

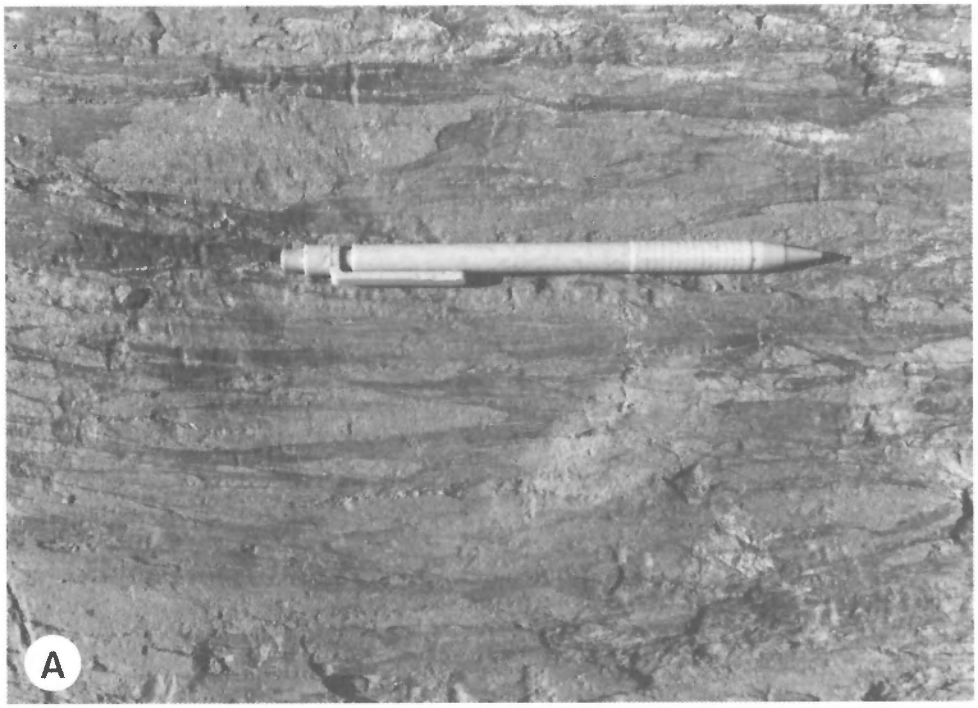
The simplest interpretation is that the contact between these two domains is a fault, here named the Bathurst fault, an interpretation supported by the presence of a narrow zone of phyllonite along the contact between the black shales and the arenaceous limestone. A more extensive phyllonite, characterized by intense transposition and boudinage, marks the contact between the Beresford and Grants Brook domains. There is a strong contrast in finite strain between the basalts and greywackes of the Beresford Domain and the adjacent rocks in the Grants Brook Domain. For example, greywacke beds in the Beresford Domain are rarely boudinaged and the basalts are weakly cleaved. Two regionally extensive marker units (i.e. the black and red phyllites) are repeated north of the base of this phyllonite zone, which is therefore also interpreted to be a major fault, here named the Grants Brook fault. These phyllonite zones are folded by large scale F_2 and F_3 folds and now have a steep to vertical dip (Fig. 2). Small scale F_2 folds refold disrupted and boudinaged bedding in these phyllonite zones and S_2 cuts across both limbs of rootless intrafolial F_1 folds (Fig. 6B). The phyllonitic fault zones are therefore interpreted as D_1 structures. However, the possibility of reactivation during F_2 or younger deformation cannot be ruled out.

Figure 4. Equal area, lower hemisphere projections of structural elements of the various domains described in this paper. The girdles defined by S_o or S_C in the Millstream, Bathurst and Beresford domains are mainly due to F_2 and F_3 folds respectively. These data were used to construct the block diagram of Figure 8.

Figure 6. D_1 high strain zones in tuffaceous greywacke-phyllite unit.

A. Internal fabric in a thrust zone in the core of Nine Mile synform (Fig. 1) with complete transposition of the more competent greywacke beds. Short limbs of F_1 folds are commonly dismembered by boudinage. Strike of S_0/S_1 is E-W.

B. Dismembered intrafolial F_1 folds in greywacke beds. Layering is very lenticular due to boudinage and pinch and swell. Note that F_1 structures are overprinted by F_2 in right lower corner of photograph. S_2 has an E-W strike.



There are few, if any reliable kinematic indicators in the phyllonites, and the overprint by F_2 and younger deformation has resulted in a complicated geometry, which inhibits determination of the sense of movement along these faults in this area. However, folding of the fault zones by upright and generally moderately plunging F_2 folds (Fig. 4) suggests that the pre-F_2 dip of the phyllonite zones was rather shallow with a strike at a high angle to S_2 since they would need to be fairly close to the F_2 shortening direction for refolding to occur.

These faults are therefore probably not strike-slip faults. Furthermore, since the faults are parallel to bedding, transcurrent movement would require large horizontal displacements in the order of tens of kilometres to explain the repetition of the red and black phyllites in the Grants Brook Domain.

Van Staal (1985, 1986, 1987) proposed that the contact between the main sheet of basalts and the underlying silicic volcanics and quartzite-phyllite unit in the Bathurst camp is probably a D_1 – related thrust. The pillow basalts become progressively more strained and altered to chlorite schists towards this contact until the pillows are no longer discernible. The highly strained, phyllonitic basalts also include lenses of glaucophane schist (Trzcienski et al., 1984; van Staal, 1987). Between these two distinct groups of volcanic rocks, i.e. silicic volcanics toward the south versus mafic volcanics towards the north (Fig. 1), there is a distinct package of rocks of variable thickness that consists of lenses of tuffaceous greywacke, red and green ferromanganiferous shales and cherts, basalts, massive sulphides, limestone and silicic volcanics surrounded by grey to black phyllites. This package is characterized by numerous zones of F_1- related phyllonites (Fig. 6A) or low grade mylonites that appear to anastomose around elongated pods of less highly strained

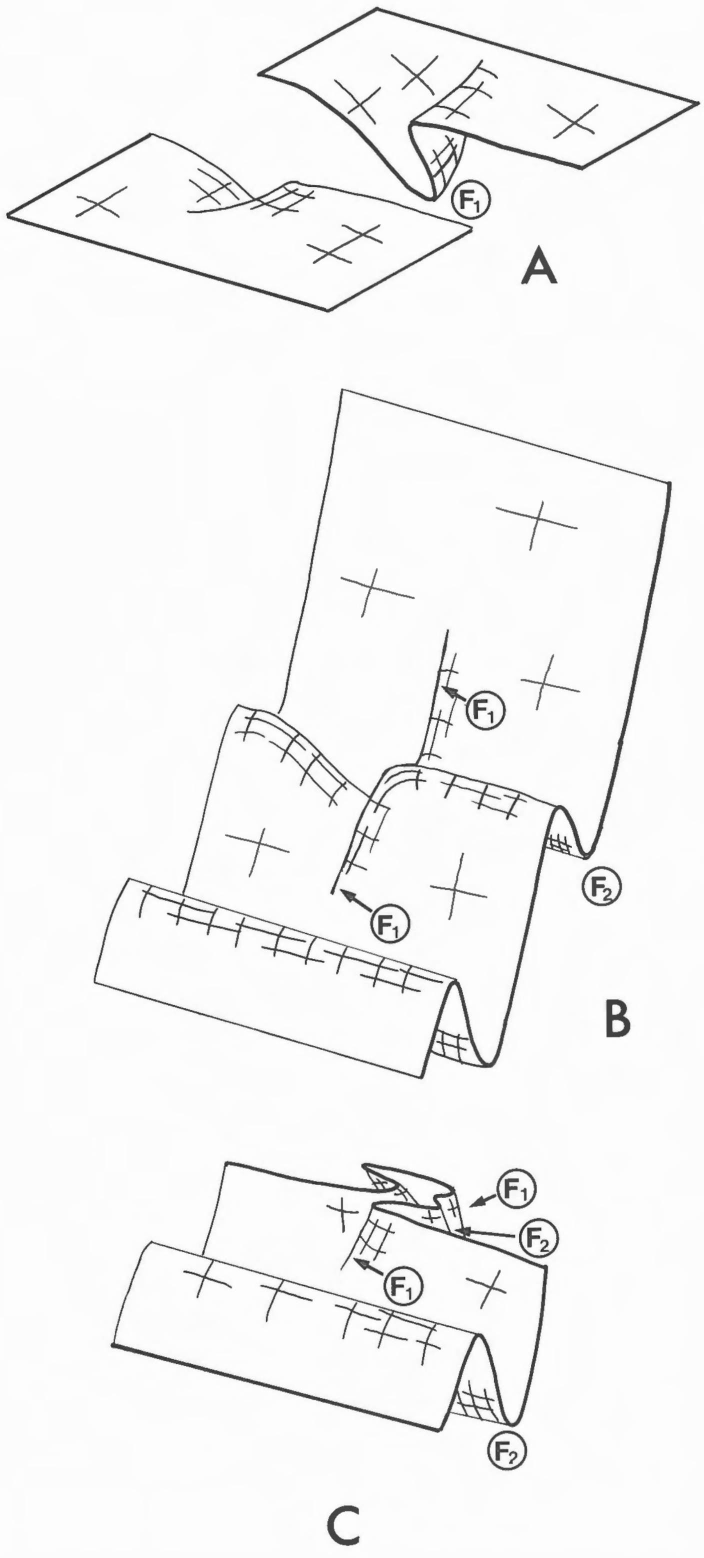


Figure 7. Sequential development of the F_1-F_2 interference structures observed in the Brunswick no. 6 and no. 12 mines. Figure adapted from van Staal and Williams (1986). Part of structure is removed in C to show geometry of a similar structure thought to be present in the big bend of the Tetagouche River after erosion.

rocks. This geometry is probably responsible for the lenticular and discontinuous nature of the various units within the complex. This heterogeneous complex of soft, phyllosilicate-rich, locally graphite-bearing rocks probably took up much of the strain during the juxtaposition of the two distinct volcanic assemblages and represents a major thrust zone rather than a low angle normal fault, since it brings glaucophane schist on top of lower pressure greenschist facies rocks (van Staal, 1987). Van Staal (in press) interpreted this movement zone as the basal thrust of an accretionary wedge.

The phyllonite zones described in this paper can be traced towards the west where they coalesce with this laterally extensive movement zone. The Bathurst and Grants Brook faults are therefore also interpreted as thrusts rather than low angle normal faults.

Van Staal (1987) argued that the main thrusting motion was in an easterly or southeasterly direction. Since no reliable kinematic indicators have been found, such a thrusting direction could not be confirmed in the Bathurst-Millstream River area; neither does the geometry of this small area by itself rule out the possibility of northwest directed movement. At present, following the regional scale arguments of van Staal (1987), an east or southeast directed thrust motion is preferred. However, such a movement picture demands explanation of the downward facing nature of the large Z-shaped F_2 fold in the Bathurst Domain and the large younging reversal in the Beresford Domain.

Interpretation of the geometry of the fault bounded domains

Large downward facing, Z-shaped F_2 folds occur at several places along the extension of the contact between the Bathurst and Beresford domains towards the west. Van Staal and Williams (1984, 1986) described two such structures in the Brunswick Mines area and the authors also mapped such a structure in the Tetagouche Falls area (see Fig. 8 of Fyffe and Noble, 1985 or Fig. 2b of Davies et al. 1985) this summer. The Z-asymmetry as well as the downward facing nature is explained by van Staal and Williams (1984, 1986) by superposition of F_2 folds on earlier, strongly overturned F_1-folds. The presence of such F_1 folds could unequivocally be demonstrated in the Brunswick No. 6 and No. 12 mines because of the extensive three dimensional exposure. Similar structures have been discovered recently in the Heath Steele Mines (C. Morreton pers. comm., 1987). Figure 7 shows schematically the movement picture of the Brunswick No. 12 and No. 6 structures as depicted by van Staal and Williams (1986). The same interference pattern is an attractive mechanism to explain the downward facing F_2 fold in the Bathurst Domain. The Bathurst Domain and its lateral extension toward the west maintain a northward facing direction for at least 15 km. Beyond this point the Vallée Lourdes formation has not been recognized with certainty, possibly because of poor exposure. Nevertheless the Nepisiquit formation can be traced tens of kilometres farther towards the southwest or south, apparently maintaining a westward facing direction (taking into account the change in strike from east-west to north-south towards the west (Fig. 1)). The Bathurst Domain is therefore not thought to form part of a giant F_1 fold nappe, rather the F_1 fold invoked to explain the downward facing F_2 fold has a short overturned limb as depicted in Figure 7 and should therefore fold the earlier Bathurst thrust fault.

The large scale younging reversal in the Beresford Domain (Fig. 2) is most simply explained as a hangingwall anticline above the Bathurst thrust fault (Fig. 8). The absence of the red and black phyllite units on the northern limb of this anticline in the investigated area could be the result of the Grants Brook fault.

A consequence of this interpretation is that the thrust and hangingwall anticline are folded by a later F_1 fold (Fig. 8). Although such an interpretation is very complex, it is nevertheless consistent with the progressive nature of D_1 seen on a small scale (van Staal and Williams, 1984, 1986; van Staal, 1986, 1987). The thrusts and related folds comprise the earliest structures, i.e. S_0/S_1 and F_{1A} whereas the strongly overturned F_1 folds, represent F_{1B}. The latter structures resemble the overturned folds described by Moore and Karig (1976) in the Nankai Trough in Japan. Such a deformation sequence is thus consistent with the accretionary wedge environment proposed by van Staal (in press). The southward facing of the southern part of the Grants Brook Domain could be explained by a horse (e.g. see Boyer and Elliot, 1982) provided its northern boundary is also a thrust. This interpretation depends on the nature of the contact between the Grants Brook and Millstream domains; whether it is a thrust or normal stratigraphic contact and whether the greywacke-phyllite unit is older or younger than the basalts in the Grants Brook Domain. In the Beresford Domain the greywacke-phyllite unit may be younger than the basalts, thus implying that the whole Grants Brook Domain is southward facing and a horse, and further that a thrust is present between the generally north facing Millstream Domain and Grants Brook Domain. This model could not be tested in the field since the contact between these two domains is not exposed, although strongly cleaved slates occur close to the approximate position of the domain boundary. Alternatively the greywacke- phyllite unit may be interbedded with the basalts

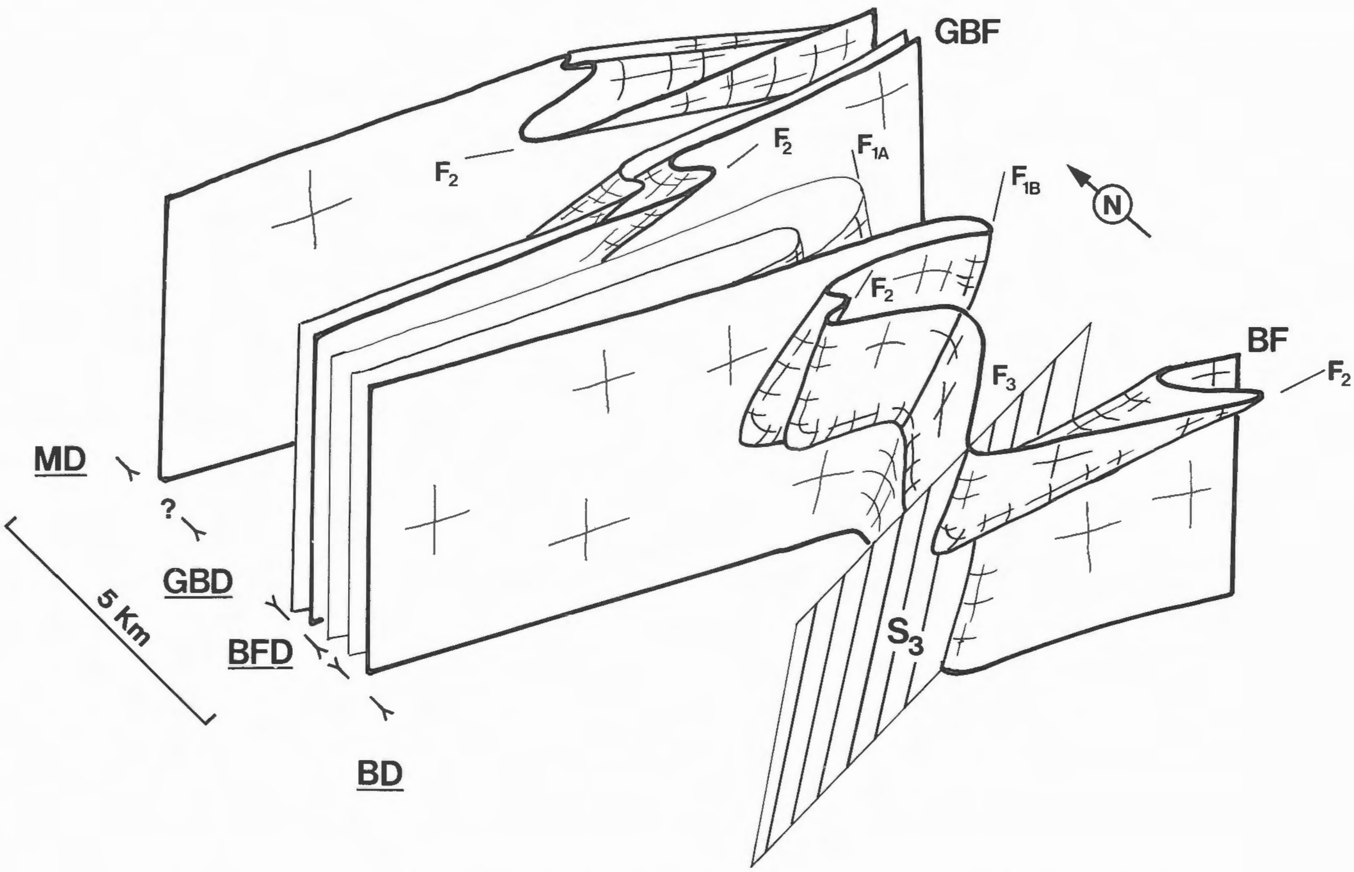


Figure 8. Interpretative block diagram of the Bathurst-Millstream River area. Thick lines represent thrusts or domain boundaries; MD, Millstream Domain; GBD, Grants Brook Domain; GBF, Grants Brook fault; BFD, Beresford Domain; BD, Bathurst Domain, BF, Bathurst fault, Younging (arrows) is indicated. F_1-F_2 interference structure depicted in southeastern corner of block refers to the interference pattern shown in Figure 7. Note that this large structure itself is refolded by F_3. The S-shaped F_2 fold at the eastern end of the Bathurst Domain is inferred to illustrate the three dimensional geometry of the F_{1B}-F_2 interference pattern. However, it is not a requirement of this model that such a fold should be exposed on surface just east of Bathurst since this depends on the level of erosion with respect to the F_{1B} fold and whether or not another F_{1B} fold is present to the east of the F_{1B} fold discussed here.

and if correct the northern part of the Grants Brook Domain may face north and possibly be in conformable contact with the Millstream formation. This model requires a thrust between the red phyllites and the greywackes, a model supported by the intense deformation in the greywackes and red phyllites exposed along Grants Brook. More work is needed to better understand these relationships.

Significance of the Millstream Domain

The turbiditic greywackes and sandstones that are exposed in the Millstream River (Fig. 2) generally young towards the north. Large F_2 folds mapped in the Millstream River (Fig. 2) are upward facing and Z-shaped looking down plunge. No apparent break has been observed between the Millstream formation and rocks mapped as Silurian by Davies et al. (1969), Skinner (1974) and Williams (1976). Both the age of the Millstream formation and its apparent low grade of metamorphism are problematic compared with elsewhere in the Tetagouche Group. Lithologically correlative rocks occur in the Pointe Verte Formation in the Elmtree inlier to the north (see above). However no fossils have been found either in these greywackes, which resemble the Late Caradocian to Ashgillian turbiditic greywackes that overlie the Caradocian black shale in northern Newfoundland (e.g. Horne, 1976; Karlstrom et al., 1982; van der Pluijm et al., in press). If this latter correlation is correct the late Ordovician hiatus, commonly quoted as evidence for the Taconian orogeny (Rast and Stringer, 1980; Fyffe, 1982) may not be present in northern New Brunswick.

CONCLUSIONS

The following conclusions can be derived from our work:

1) The superposition principle should be used with the greatest care, and only where the structure is understood in such a severely deformed area as the Bathurst Camp.
2) No angular unconformity need exist within the Tetagouche Group and neither is there evidence for significant late Ordovician - pre-Silurian volcanism in the Miramichi Zone. Our mapping indicates that it is much simpler to interpret all basalts as Middle Ordovician in age analogous to the situation in the Central Mobile Belt of Newfoundland.
3) A distinct package of tectonically mixed rocks can be traced from Camel Back Mountain (Fig. 1) to Bathurst. This package is interpreted as part of a large movement zone along which rocks of oceanic affinity, i.e. MORB and oceanic within-plate basalts (van Staal; 1987, Winchester and van Staal, unpubl. data), were thrust over the lower Tetagouche Group rocks that are probably underlain by continental crust (van Staal, 1987). The presence of numerous lenses of massive sulphides within this movement zone indicates that a good understanding of this package of rocks is vital for mineral exploration in this area.

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The Portage Brook troctolite, a layered intrusion in the New Brunswick Appalachians[1]

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Abstract

The Portage Brook troctolite is a small layered intrusion, of possible Devonian age, occurring in strongly folded Ordovician sedimentary and volcanic rocks in northern New Brunswick intrusion is made up dominantly of troctolite with subordinate peridotite, anorthosite and anorthositic gabbro, most of which can be considered as mesocumulates. Layering is defined by modal variation of olivine and plagioclase and planar lamination of plagioclase. The intrusion has a basin-like structure and youngs toward the north. Features suggestive of magmatic current activity during the formation of the layered sequence are present. Sulphide concentrations around 3-5 % are confined to some thin anorthosite layers with limited lateral continuity. The Portage Brook sulphides are richer in pentlandite and chalcopyrite than those of the Goodwin Lake intrusion. The sulphides appear to have a magmatic genesis.

Résumé

La troctolite de Portage Brook est une petite intrusion en couches, pouvant dater du Dévonien, que l'on trouve dans des roches sédimentaires et volcaniques fortement plissées d'âge ordovicien du nord du Nouveau-Brunswick. L'intrusion se compose surtout de troctolite avec de petites quantités de périodotite, d'anorthosite et de gabbro anorthositique, la plupart pouvant être considérés comme des mésocumulats. La stratification est définie par une variation modale d'olivine et de plagioclase ainsi que par une lamination plane de plagioclase. L'intrusion présente une structure en forme de bassin et devient progressivement plus jeune vers le nord. On a constaté la présence d'éléments qui permettent de croire à une activité de courants magmatiques pendant la formation de la séquence stratifiée. Des concentrations de sulfures d'environ 3-5 % se restreignent à quelques minces couches d'anorthosite qui font preuve d'une continuité latérale limitée. Les sulfures de Portage Brook sont plus riches en pentlandite et en chalcopyrite que ceux de l'intrusion de Goodwin Lake. Les sulfures semblent avoir pour origine un événement de nature magmatique.

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

INTRODUCTION

Numerous mafic-ultramafic intrusions occur in the Appalachian-Caledonian orogen extending from northeastern United States through Canada, Greenland, Ireland, and Scotland to Norway. Although the ages of the intrusions range from Ordovician to Devonian, they are mostly synorogenic in nature (Boyd and Mathiesen, 1979; Wilson, et al., 1981; Thompson, 1984; Paktunc, 1986a) and have many common characteristics; therefore, they appear to represent a distinct class of mafic-ultramafic intrusions. Examples include the Moxie, Katahdin, Union and Warren intrusions in Maine, St.Stephen, Goodwin Lake, Portage Brook and Mechanicsville in New Brunswick, Huntly and Insch in Scotland, and Rana and Fongen-Hyllingen in Norway. The study area lies in the northwestern part of the Miramichi tectonic zone, approximately 100 km west of Bathurst. This zone forms the New Brunswick portion of the Gander Terrane of the Appalachian Orogen and contains other mafic intrusions such as the Goodwin Lake, Moxie and Katahdin (Fig. 1) which host significant Ni-Cu sulphide resources.

The Portage Brook intrusion was mapped at 1:10 000 scale as part of a study of platinum group element potential of the mafic-ultramafic rocks in the New Brunswick Appalachians. The results of the field studies on the intrusion are presented here.

GENERAL GEOLOGY

In northern New Brunswick, three regional geological units were defined by Smith and Skinner (1958) and Davies et al. (1969): 1) the Ordovician folded belt, 2) the Silurian-Devonian folded belt, and 3) Pennsylvanian cover. The Portage Brook troctolite lies in the northwestern edge of the Ordovician folded belt. This belt is commonly referred to as the Tetagouche Group (Smith and Skinner, 1958; Davies et al., 1969; Skinner, 1974) and consists of metamorphosed and strongly deformed sedimentary rocks, basalt, rhyolite, gabbroic sills and dykes. These rocks have been intruded by mafic and granitic rocks of Devonian age.

Devonian mafic rocks occurring in the Portage Brook region near Popple Depot were considered by Helmstaedt (1971) and Whalen (1987) as parts of the Mount Elizabeth granitic batholith. Hill (1971) mapped the mafic rocks as belonging to two lithologically distinct bodies; the Portage Brook intrusion near Popple Depot and the West Portage Brook intrusion farther to the north latter had been called the Portage Brook intrusion by Wolfe (1961). To avoid confusion, the body near Popple Depot is referred to as the Portage Brook troctolite in this paper instead of the Portage Brook intrusion (Paktunc, 1986b). The relationship of the troctolitic body to the less mafic rocks in the north is not known yet.

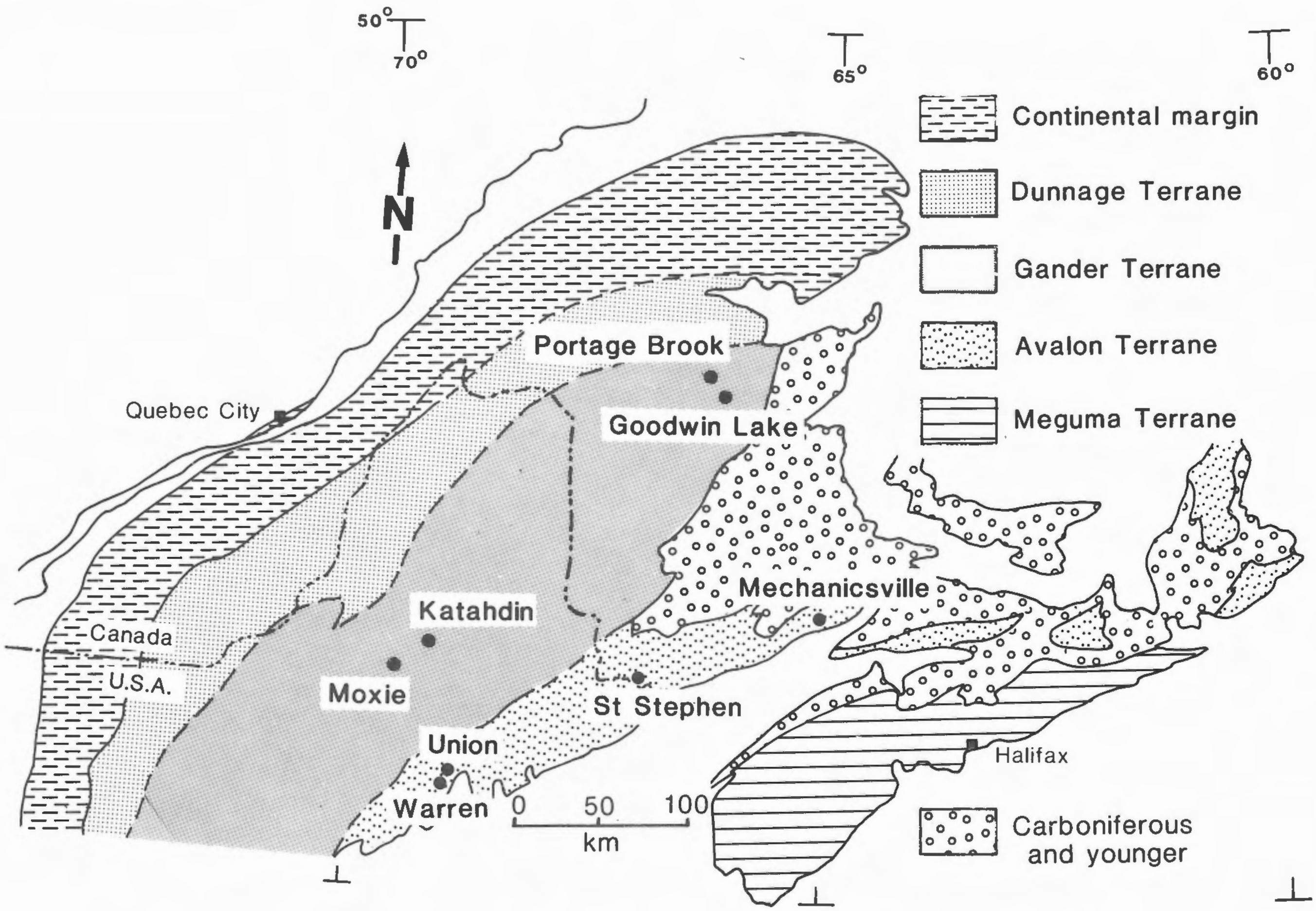

Figure 1. Map showing the location of mafic-ultramafic intrusions cited in text in tectonic zones of the Appalachian Orogen. Major plutons related to and younger than the Appalachian Orogen are omitted and some geological features are simplified for clarity.

The country rock is not well exposed near the intrusion; however, Helmstaedt (1971) postulated that the body lies largely within Ordovician rocks and is discordant in nature.

PORTAGE BROOK TROCTOLITE

Field relations

The intrusion is small (approximately 10 km^2) and well layered. In the northern half of the intrusion layering has a basin-like structure with inward dips of 20-30° (Fig. 2) structurally highest exposed member of the intrusion lies somewhere near its north-central portion. Successively lower members appear to the west, east and south. This suggests that the intrusion plunges northerly. Outcrops of gabbroic and dioritic rocks to the north, in the younging direction of the troctolite body, may represent the upper differentiated levels of the Portage Brook troctolite, and indicate the northern extension of the troctolite body remains concealed. Promi-

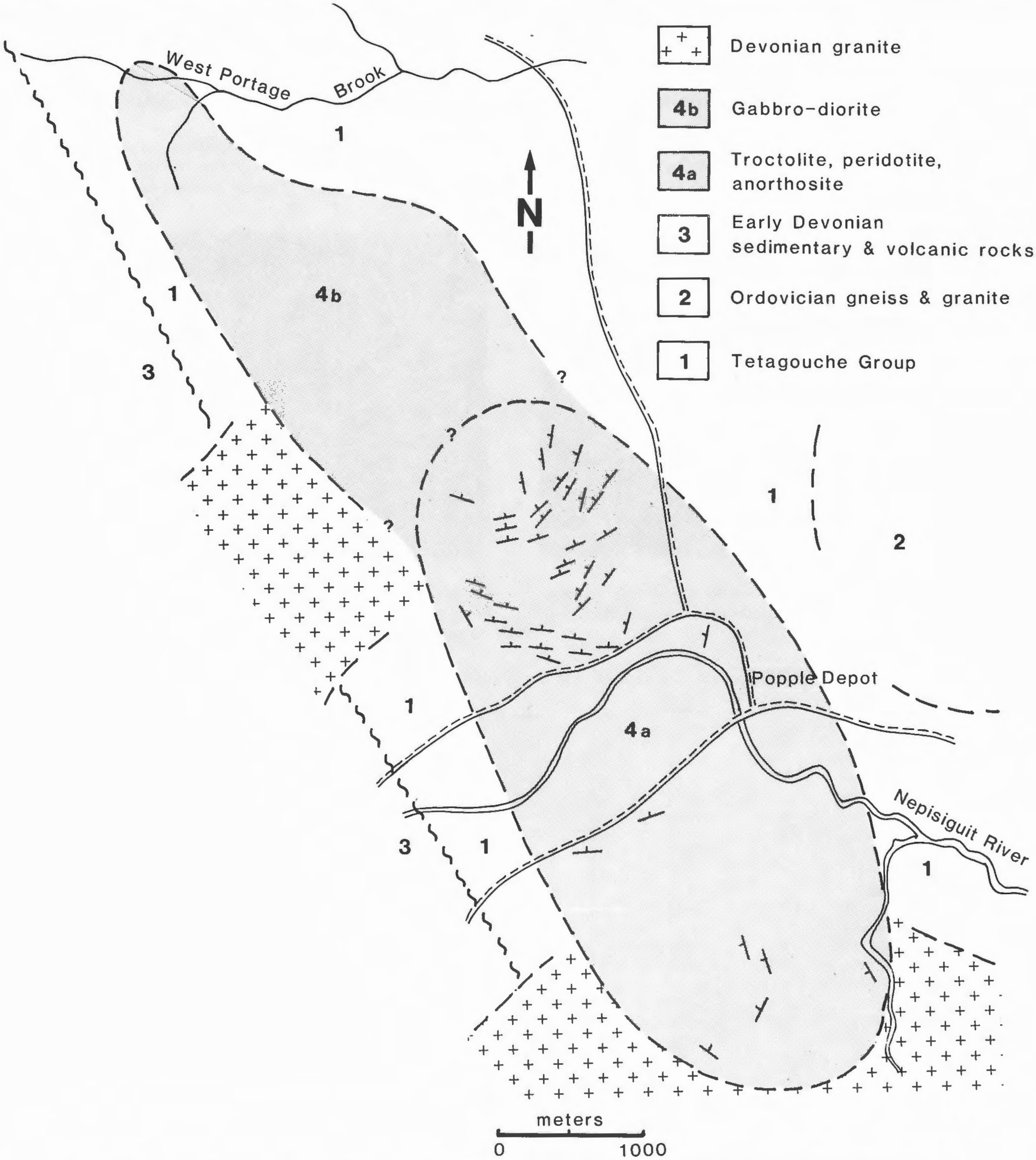

Figure 2. Geological map of the Portage Brook intrusion. Geology of the surrounding rocks partly based on Helmstaedt (1971).

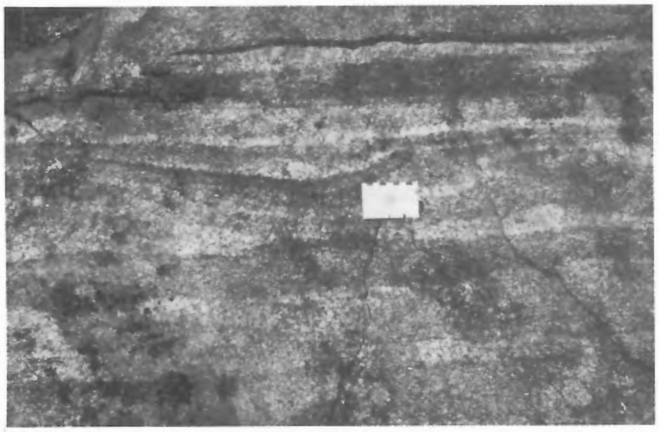

Figure 3. Modal layering in troctolitic rocks. Note the area near the scale where the layers have been downwarped, presumably due to slumping or represent a scour structure. Scale in centimetres.

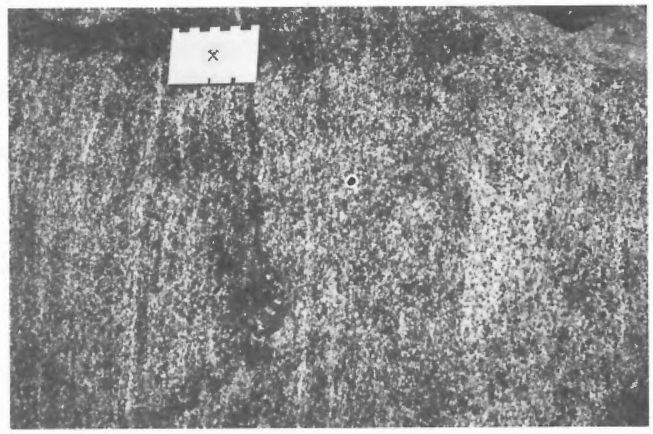

Figure 5. Modal lamination in troctolite. Note the sharp concordant contact (one third of the distance from the top) between overlying anorthositic gabbro and troctolite.

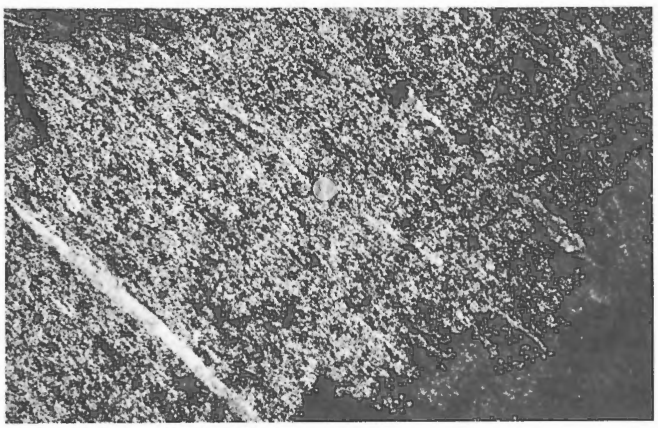

Figure 4. Schlieren layering formed by discontinuous plagioclase-rich and olivine- rich laminae in troctolite. Coin is 2.4 cm across.

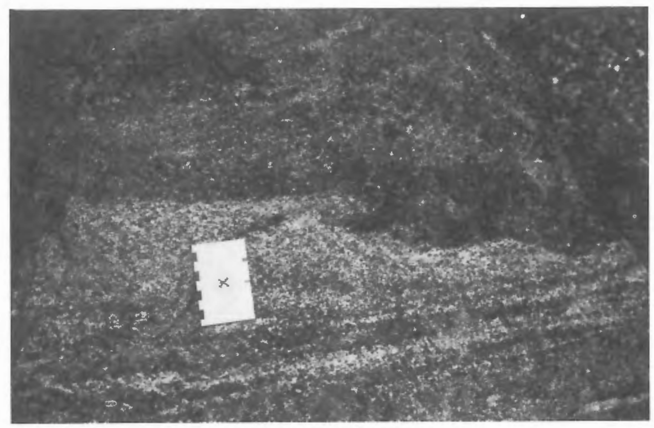

Figure 6. Discordant contact between peridotite and overlying troctolite. This appears to be a scour feature formed by magmatic currents.

Table 1. Pt, Pd, Au, Cr, Cu, Co, Ni and S analyses of the sulphide-bearing samples

	rock	Pt	Pd	Au	Cr	Cu	Co	Ni	S
P860301	P	<15	<2	<1	na	338	98	793	0.48
P860401	T peg	<15	<2	10	na	715	72	794	0.92
P860501	P	<15	<2	<1	na	205	76	843	0.27
P862002	A	<15	<2	<1	na	52	28	221	0.13
P862501	T	<15	<2	<1	na	86	51	232	0.13
P862801	T	<15	<2	<1	na	183	48	559	0.15
P863301	T	<15	<2	<1	na	83	41	291	0.14
P863601	T	<15	<2	<1	na	10	11	33	0.03
P863701	T	<15	<2	<1	na	145	50	380	0.20
P863901	T	<15	<2	<1	na	38	43	130	0.18
P864002	T	<15	<2	<1	na	80	47	365	0.16
P864301	P	<15	<2	<1	na	252	80	596	0.14
P8701	OG	<15	<2	3	171	550	44	501	0.84
P8721	PT	<15	<2	<1	219	173	38	334	0.36
P8734	T	<15	<2	2	303	18	9	21	0.33
P8709	OG	<15	<2	4	522	222	22	293	0.25
P8737	T	<15	<2	<1	162	195	50	391	0.35
P8755	P	<15	<2	<1	783	218	73	821	0.45
P8758	T	<15	3	<1	379	104	28	176	0.17
P8772	P	<15	<2	4	2522	139	94	744	0.27

Pt, Pd, Au in ppb; Cr, Cu, Co, Ni in ppm; S in wt%
A: anorthosite; OG: olivine gabbro; P: peridotite;
PT: picritic troctolite; T: troctolite
peg: pegmatitic; na: not analyzed

nent peridotitic and anorthositic units are common in the southern portion of the intrusion, corresponding to the lowest exposed members. These are usually medium- to coarse-grained massive rocks without any internal fabric, unlike the troctolites. Scattered outcrops of massive peridotite in the northern half of the intrusion may represent and upper level thick peridotite layer.

Layering is defined by modal variation of olivine and plagioclase (Fig. 3) and planar lamination defined by the arrangement of tabular plagioclase crystals parallel to layering (Fig. 4, 5). Size-grading is not a common feature. Thickness of the layers is generally in the centimetre range. Some layers pinch out over short distances. However, lateral continuity is difficult to assess because of the small size of outcrops and the absence of a marker horizon. Trough-like structures and discordances among the layers are present (Fig. 6). Also present are some depositional deformation structures such as slump and layer disruption (Fig. 3).

There is no indication of metamorphism and/or deformation present in the intrusion rocks. They are generally very fresh except for some fine serpentine veins in olivines and expansion cracks in plagioclase grains around olivine grains.

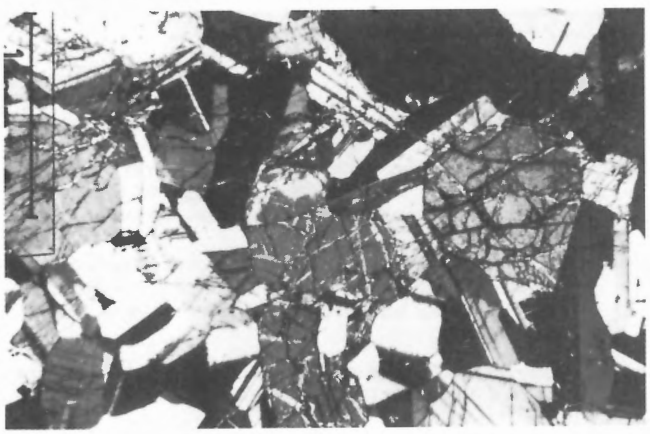

Figure 7. Photomicrograph of olivine-plagioclase mesocumulate. Transmitted light. Crossed polarizers. Scale bar: 1mm.

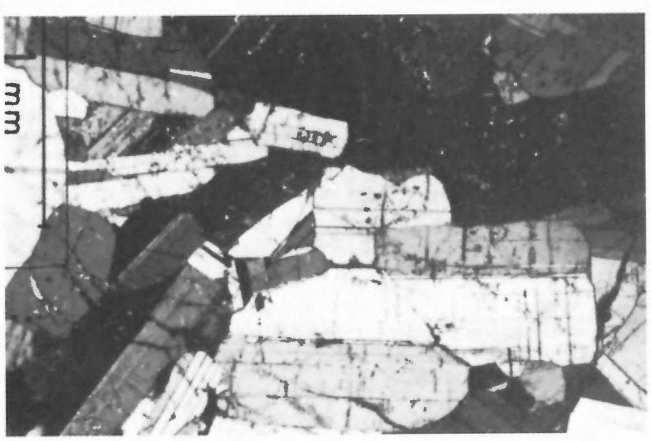

Figure 8. Photomicrograph of a typical plagioclase adcumulate. Transmitted light. Crossed polarizers. Scale bar: 1mm.

Petrology

The intrusion is made up dominantly of troctolite. Peridotite, olivine gabbro, anorthosite and anorthositic gabbro are the subordinate rock types observed. All units are cumulates rock which can be considered as a marginal facies or feeder was encountered. Peridotite is made up of millimetre-sized cumulus olivine, centimetre-sized pyroxene and phlogopite which usually form oikocrysts, and postcumulus plagioclase. Cumulus chromite is an accessory phase in peridotites. Peridotites in general can be classified as olivine-chromite mesocumulates[1]. Troctolites are made up of closely packed cumulus plagioclase and olivine. A wide range in the olivine to plagioclase ratio exists. Troctolites with more than 50 % olivine are referred to as picritic troctolite, those with less, troctolite. Most of the troctolites can be considered as olivine-plagioclase mesocumulates (Fig. 7). With an increase in the clinopyroxene content, they become orthocumulates. Anorthosites are largely adcumulates (Fig. 8).

Magmatic structures and the petrology of the intrusion suggest that it formed by accumulations of olivine and plagioclase crystals carried by convective currents. Since the intrusion is small, the effect of sidewall cooling may be im-

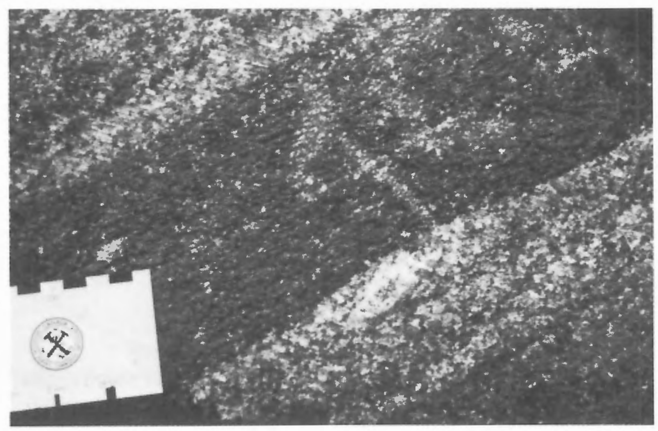

Figure 9. Intermittent layering. Modally graded layers alternating with uniform peridotite layers.

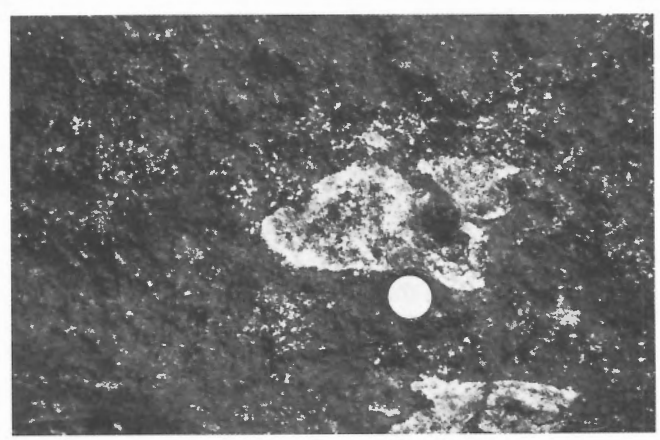

Figure 10. Anorthositic cognate xenoliths in peridotite. Coin is 2.4 cm across.

[1] Irvine's (1982) cumulus terminology is adopted in this paper.

Figure 11. Photomicrograph of interstitial sulphide. Scale bar is 0.5 mm.

portant. Thus, crystallization occurred at both the floor and the walls of the chamber. The presence of modal lamination and scour-like features indicate magmatic current activity during formation of the layers. Thin modally graded troctolite layers alternate with uniform peridotite layers (Fig. 9). This type of layering is similar to intermittent layering described in the Stillwater Complex (Hess, 1960), Skaergaard (McBirney and Noyes, 1979) and the Jimberlana intrusion (Campbell, 1978). Intermittent layering is conspicuously absent from the two major peridotite horizons of the Portage Brook intrusion, as in the Stillwater Complex and the St.Stephen intrusion.

NICKEL-COPPER SULPHIDE MINERALIZATION

Minor amounts of sulphides are present in the intrusion rocks. Their abundance reaches 5 % in some anorthositic cognate xenoliths (Fig. 10) and thin anorthositic layers. Sulphides are composed of pyrrhotite, pentlandite and chalcopyrite. They usually occupy intersticies between cumulus crystals (Fig. 11).

Twenty fresh surface samples containing several per cent of sulphides were selected for analysis, the results of which are given in Table 1. Pt, Pd and Au contents are mostly below detection limits. Maximum Ni, Cu and Co abundances are 843 ppm, 715 ppm and 98 ppm respectively. The metal (Ni, Cu, and Co) to sulphur ratio (average value about 0.25) is considerably higher than that of the St.Stephen and Goodwin Lake sulphides (Paktunc 1987). Consequently, if significant sulphide concentrations were discovered in the Portage Brook troctolite, their unit base metal value would be greater than the others mentioned, but the low PGE levels encountered to date offer little additional encouragement.

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Nickel-copper sulphide mineralization associated with the Goodwin Lake intrusion, northern New Brunswick[1]

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Paktunc, A.D., Nickel-copper sulphide mineralization associated with the Goodwin Lake intrusion, northern New Brunswick; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 155-161, 1988.

Abstract

The Goodwin Lake intrusion, possibly Devonian, occurs in strongly folded Ordovician sedimentary and volcanic rocks and consists entirely of cumulate rocks ranging from norite and gabbro to peridotite. Centimetre scale layering is defined by the modal variation of phases. Mineralization, predominantly pyrrhotite with subordinate pentlandite and chalcopyrite, is widespread in the intrusion. Sulphides occur interstitially to cumulus phases and as massive to semi-massive concentrations. Drilling has demonstrated in excess of 5 million tonnes with 0.34 % Cu, 0.28 % Ni and 0.03 % Co. Platinum-group element concentrations are generally low (less than 366 ppb total Pt+Pd+Au). Cu/Ni ratios of the sulphides are similar to those of flood basalt related ores. The nature of the mineralization shows a strong resemblance to the St Stephen and the Moxie ores. The Goodwin Lake sulphides are more depleted in Ni and Cu compared to those of the St Stephen and the Portage Brook.

Résumé

L'intrusion probablement dévonienne de Goodwin Lake se trouve dans des roches sédimentaires et volcaniques fortement plissées d'âge ordovicien et se compose entièrement de roches accumulées dont la nature varie de la norite et du gabbro à la péridotite. Une stratification à l'échelle du centimètre est définie par la variation modale des phases. Une minéralisation, soit surtout de la pyrrhotite avec de petites quantités de pentlandite et de chalcopyrite, est très répandue dans l'intrusion. Les sulfures se rencontrent sous des formes qui vont du mode interstitiel aux phases de cumulats et sous forme de concentrations massives à semi-massives. Un forage a démontré la présence de plus de 5 millions de tonnes titrant 0,34 % de Cu, 0,28 % de Ni et 0,03 % de Co. Les concentrations d'éléments du groupe du platine sont généralement faibles (moins de 366 parties par milliard pour tout Pt + Pd + Au). Les rapports Cu/Ni des sulfures sont analogues à ceux des minerais apparentés aux basaltes de plateau. La nature de la minéralisation révèle une forte ressemblance aux minerais de St. Stephen et de Moxie. Les sulfures de Goodwin Lake sont plus épuisés en Ni et en Cu comparativement à ceux de St. Stephen et de Portage Brook.

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

INTRODUCTION

The Goodwin Lake intrusion resembles closely other mafic-ultramafic intrusions occurring in the Appalachian Orogen. Examples with significant nickel-copper sulphide mineralization include the St.Stephen, Goodwin Lake, Mechanicsville, Union, Warren and Moxie intrusions (Fig. 1). The nickel-copper sulphide mineralization in the Goodwin Lake intrusion forms the second most important Ni-Cu sulphide resource after the St.Stephen intrusion in the Canadian Appalachians mineralization has been intermittently explored by private companies since its discovery in 1954. This report briefly describes the Ni-Cu resources in the Goodwin Lake intrusion and presents some preliminary results of an investigation which forms part of a study of the economic potential of mafic-ultramafic intrusions in New Brunswick.

GEOLOGY

The area is underlain by the Tetagouche Group, which consists of strongly folded Ordovician sedimentary and volcanic rocks (Smith and Skinner, 1958; Davies et al., 1969; Skinner, 1974). Sedimentary rocks consist mainly of intercalated slate and quartz greywacke and their metamorphosed equivalents (Skinner, 1974). Basalts and rhyolites form the volcanic rocks. Intruding this sedimentary and volcanic package are subvolcanic mafic sills and dykes. The Tetagouche Group rocks were metamorphosed and strongly deformed during the Ordovician Taconian and the Devonian Acadian Orogenies (Skinner, 1974).

The Goodwin Lake intrusion consists of two bodies. One of these is an elongate lens (approximately 4 km long and 500 m wide) extending from Goodwin Lake to Maliseet Mountain in a northwesterly direction (Fig. 2). The other, a circular body about 700 m in diameter, is located near the confluence of the Northwest Miramichi River and its South Branch about 2 km northeast of Goodwin Lake. These are referred to as the western and eastern bodies respectively. The western body consists of gabbronorite, norite and gabbro with subordinate olivine gabbro, troctolite and peridotite. The eastern body consists of peridotite, troctolite, norite and gabbronorite. Igneous structure of the intrusion is obscure due to extremely poor exposure. All rocks are cumulates. They are mostly medium grained, but numerous pegmatitic portions are encountered along the core. Centimeter-scale layering and modal lamination are well developed in troctolites (Fig. 3).

The intrusion does not appear to be metamorphosed. The rocks are generally fresh. The country rocks are schistose sedimentary and volcanic rocks of the Tetagouche Group. Hornfelsic rocks are present around the intrusion and as xenoliths. The intrusion is cut by fine- to medium-grained granitic rocks.

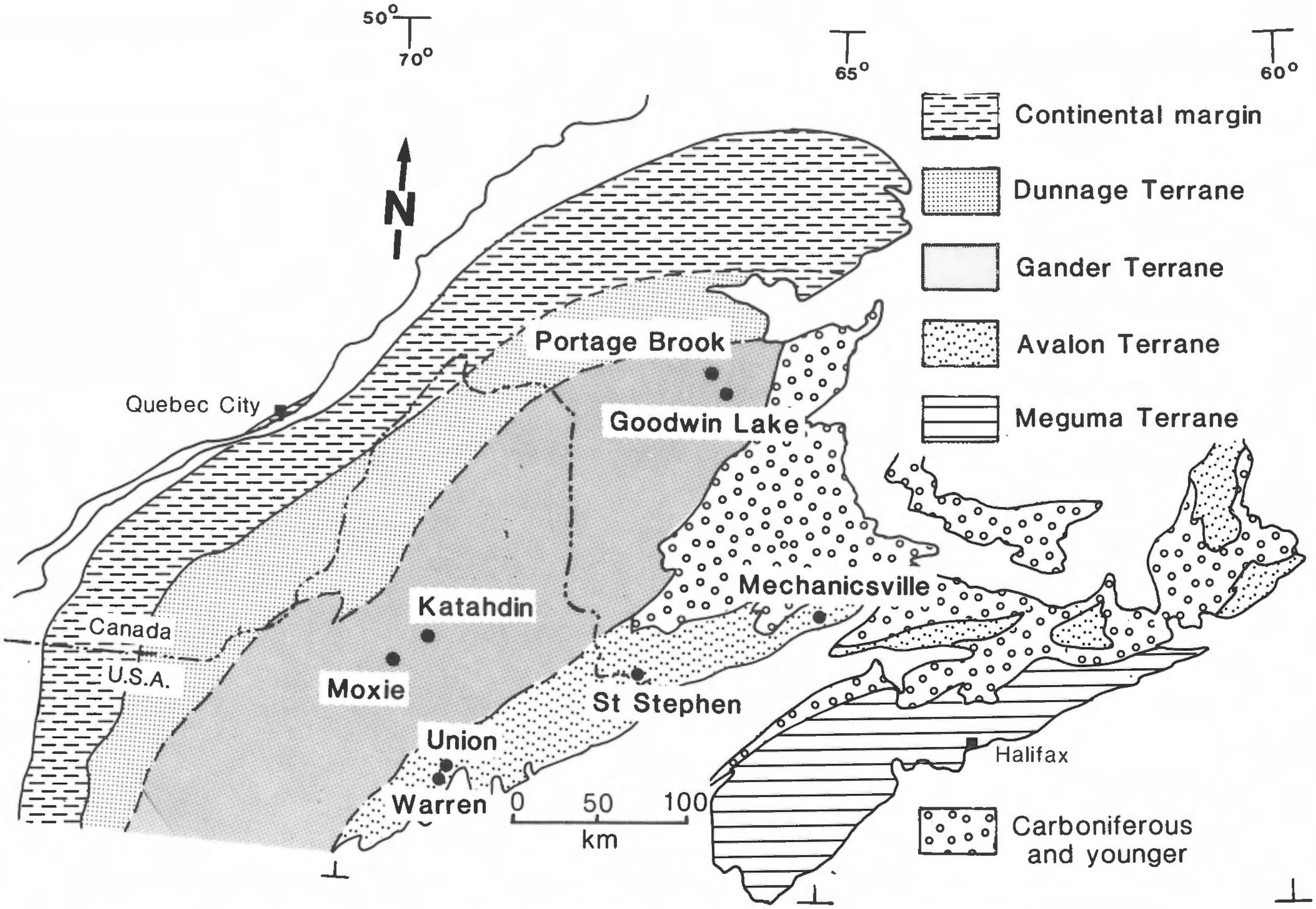

Figure 1. Map showing the location of mafic-ultramafic intrusions cited in the text in tectonic zones of the Appalachian Orogen. Major plutons younger than the Appalachian Orogen are omitted

MINERALIZATION

Mineralization consisting mainly of pyrrhotite and minor chalcopyrite and pentlandite is widespread throughout the intrusion. Although their occurrence is sporadic, sulphides appear to be confined to noritic and peridotitic rocks and occur as interstitial grains to the cumulus olivine, orthopyroxene and plagioclase grains (Fig. 4) and as massive to semi-massive concentrations. Interstitial sulphides suggest a magmatic intercumulus origin for the sulphides whereas irregular and vein- like character of the massive and semi-massive sulphides is suggestive of post-depositional deformation and/or mobilization of the sulphides. Pentlandite occurs as subhedral grains and as anhedral flame-like exsolution bodies in pyrrhotite (Fig. 5). Chalcopyrite is found as anhedral grains in association with pyrrhotite and as discrete thin veinlets indicating preferential remobilization of chalcopyrite. Commonly associated with the sulphides are composite magnetite and ilmenite grains. They usually form subhedral to anhedral grains and exhibit exsolution lamellae. They also form spectacular symplectites with silicates (Fig. 6) in some sulphide and oxide-bearing noritic rocks.

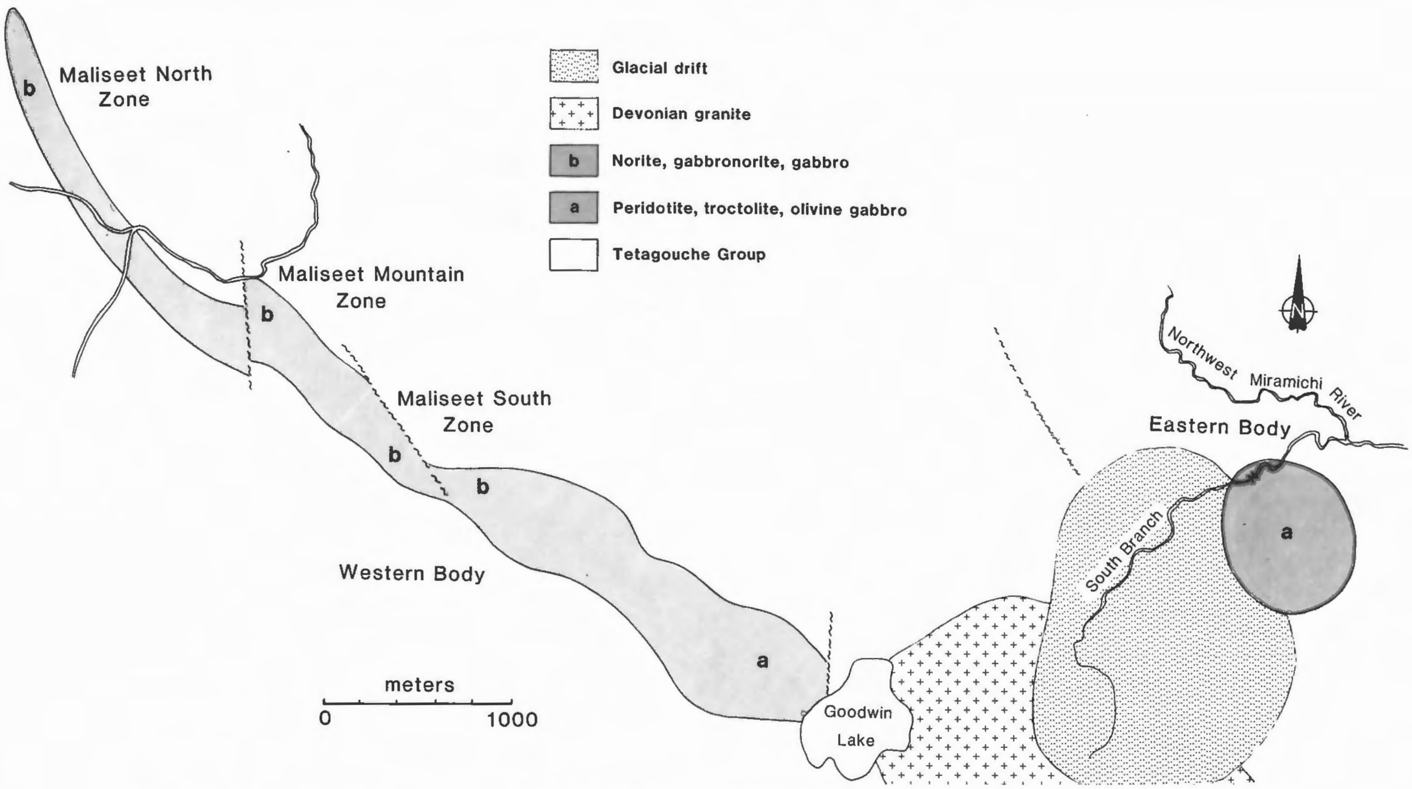

Figure 2. Geological map of the Goodwin Lake intrusion and location of the mineralized zones. Partly based on McNutt (1961).

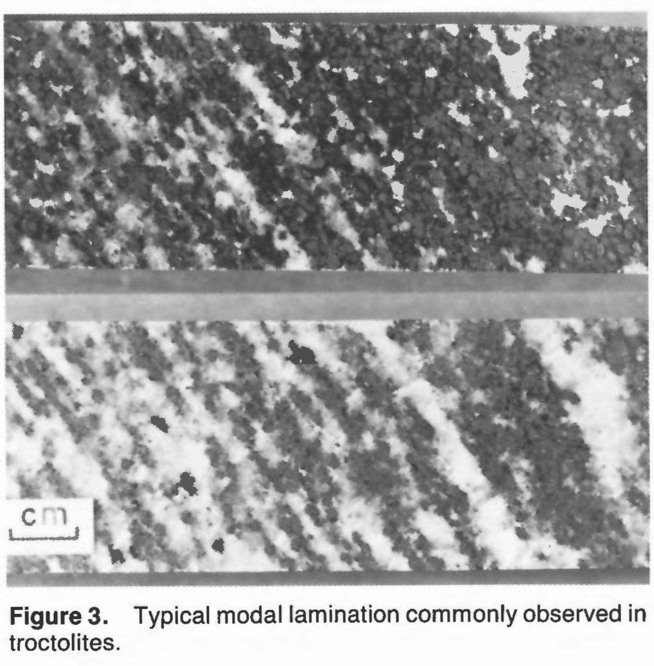

Figure 3. Typical modal lamination commonly observed in troctolites.

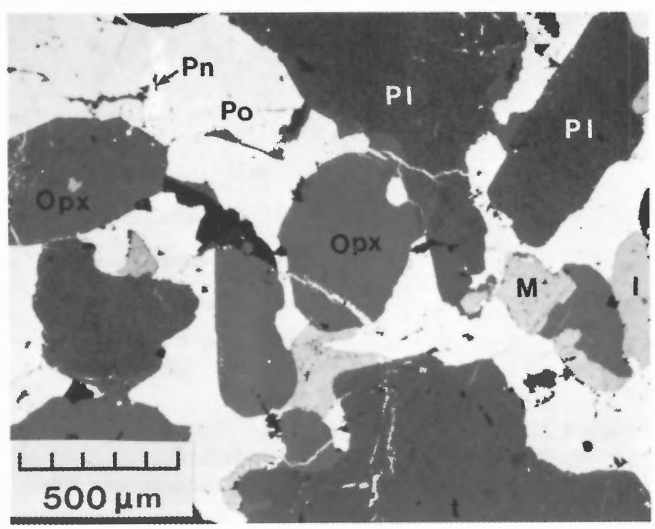

Figure 4. Photomicrograph of matrix ore. Sulphides occupy intersticies between cumulus orthopyroxene (Opx), plagioclase (Pl), and magnetite (M). I: ilmenite; Pn: pentlandite; Po: pyrrhotite. Reflected light

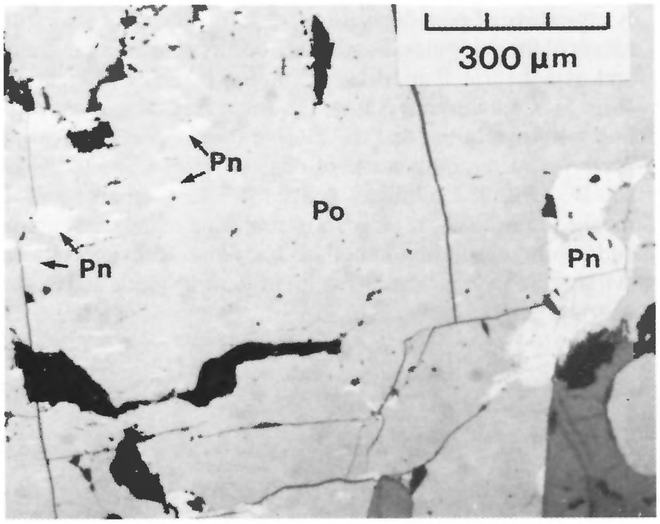

Figure 5. Photomicrograph showing discrete grains and flame-like exsolutions of pentlandite (Pn) in pyrrhotite (Po). Reflected light.

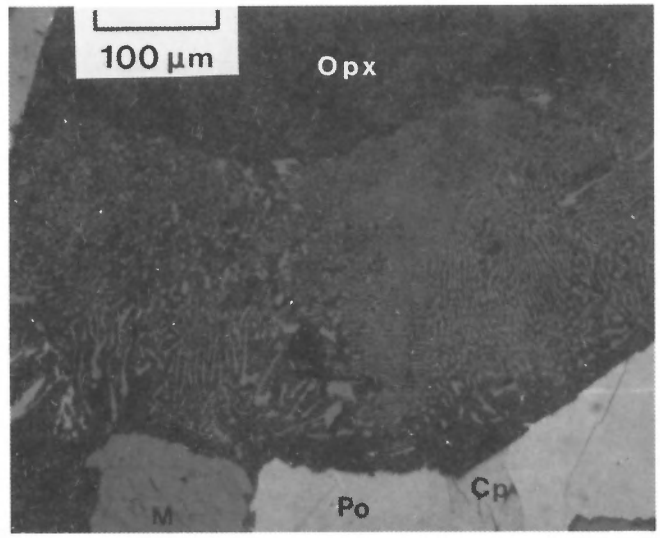

Figure 6. Photomicrograph of oxide — silicate symplectites. Opx: orthopyroxene; M: magnetite; Po: pyrrhotite; Cp: chalcopyrite. Reflected light, partially crossed polarizers.

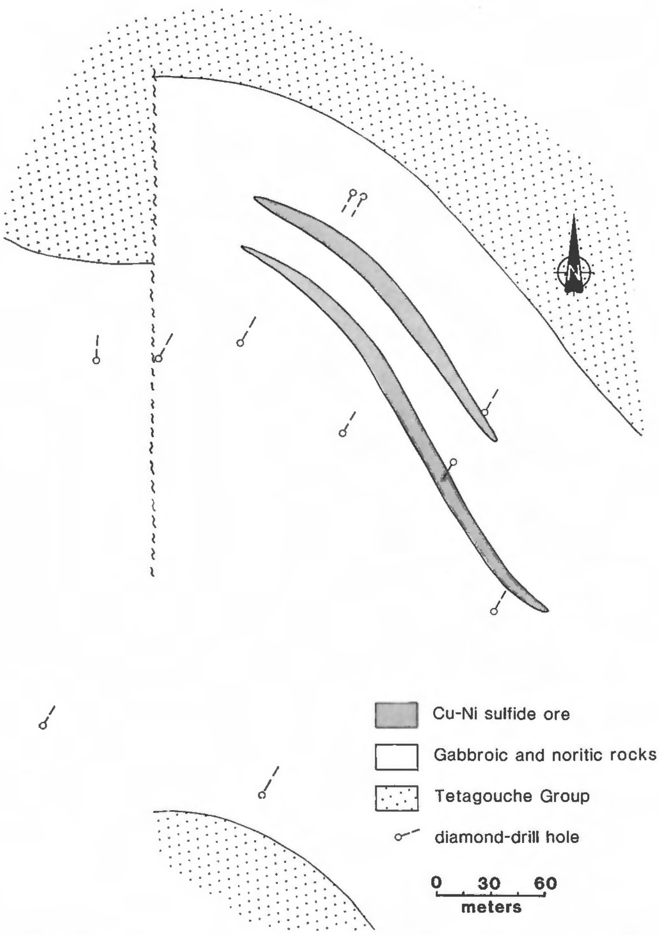

Figure 7. Maliseet Mountain zone showing outline of ore bodies near the surface. Prepared from the diamond-drill hole logs of this study and those from the assessment files.

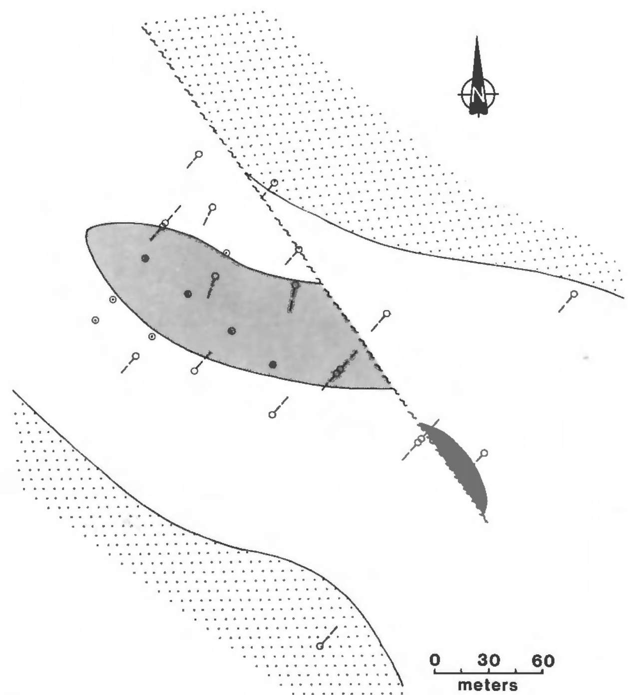

Figure 8. Plan of the Maliseet South ore zone. Ore body projected from diamond-drill hole intersections near the surface. Prepared from the logs of this study and those of companies. Refer to Figure 7 for legend.

Table 1. Ni, Cu and Co contents (compiled from the company assays) and Cu/(Cu + Ni) ratios of the mineralized zones

Zone	n	Ni (wt%) range	Ni (wt%) mean(s.d)	Cu (wt%) range	Cu (wt%) mean(s.d)	Co (wt%) range	Co (wt%) mean(s.d)	Cu/(Cu+Ni) mean(s.d)
Maliseet Mountain	11	0.02-0.88	0.26(0.28)	0.05-0.72	0.31(0.22)	0.01-0.16	0.06(0.05)	0.61(0.18)
Maliseet South	435	0.01-0.64	0.20(0.15)	0.01-3.14	0.26(0.22)	0.00-0.10	0.02(0.02)	0.58(0.15)
Eastern	23	0.01-0.71	0.25(0.18)	0.02-0.89	0.32(0.19)	0.01-0.07	0.03(0.02)	0.57(0.15)

n: number of samples; s.d. standard deviation

Table 2. Pt, Pd, Au, Cr, Cu, Co, Ni, and S analyses of the sulphide-bearing samples (this study)

	rock	sulphide	Pt	Pd	Au	Cr	Cu	Co	Ni	S
Maliseet South										
70-11-340			<15	<2	<1		47	17	28	0.15
70-13-550	N	mx	<15	4	6	56	2385	231	2400	5.71
70-17-299	N	ds	<15	<2	<1	209	55	10	52	0.21
70-17-300	N		<15	6	6	270	33	10	25	0.10
70-17-305	N	mx	<15	<2	3	147	173	30	141	0.44
70-2-242	N	mx	<15	6	8	44	2125	342	4240	10.77
70-6-261	N	mx	<15	5	8	33	4460	313	3210	9.28
70-7-492	N	ds	<15	<2	<1	36	183	29	180	0.57
70-7-64	N	mx	<15	<2	<1	65	1925	364	2850	7.44
71-10-161	GN		45	9	27	36	1645	566	9280	16.31
71-10-200	GN	ds	<15	8	11	56	3420	281	4940	7.82
71-10-276	GN	ds	<15	7	15	44	3010	121	1980	2.99
71-10-432	GN	ds	<15	5	6	62	4010	79	940	2.78
71-10-597	GN	ds	<15	<2	<1	28	70	25	16	0.49
71-8-329	OG	ds	<15	<2	<1	106	195	33	138	0.55
71-8-356	OG	ds	<15	3	2	48	677	130	860	2.70
71-8-363	OG	ds	<15	<2	2	82	651	151	754	2.65
71-8-407	OG	ds	<15	<2	<1	52	778	152	1240	3.21
71-9-382	GN		<15	<2	<1		159	28	23	2.34
71-9-390	GN		34	12	<1		400	42	210	0.71
71-9-393	GN		<15	<2	<1	59	65	21	42	0.72
71-9-445	GN		<15	<2	2		905	65	46	2.93
71-9-493	GN		<15	<2	4	43	2540	54	444	1.64
71-9-494	GN		<15	7	14	34	1700	300	4120	12.19
71-9-649	GN		<15	4	12	55	2730	135	1490	3.84
Maliseet Mt										
71-4-169	G	bb	<15	3	5		427	120	832	3.22
71-4-179	G	bb	<15	5	8	131	1610	205	2000	4.70
71-4-181	G	bb	<15	6	4	108	1272	178	1820	4.23
71-4-293	G	sm	<15	18	333	269	1214	504	881	20.42
71-4-299	G	ds	<15	3	29	173	171	88	145	2.61
71-4-304	G	ds	<15	<2	9	101	174	60	129	1.99
71-4-381	G alt	rm	53	20	8		220	100	170	3.62
71-4-386	G	mx	<15	4	5	365	276	87	212	3.99
71-4-399	G peg		<15	<2	<1	361	78	16	64	0.28
71-4-402	G	ds	<15	5	7	289	1152	171	1051	4.36
71-4-404	G	bb	<15	3	3		819	109	660	3.02
71-4-441	G	sm	<15	9	14		589	507	1284	26.50
71-5-137			<15	<2	4	143	121	75	95	2.00
71-5-263			<15	<2	3	320	49	20	56	0.74
Eastern body										
71-1-104	GN	ds	<15	<2	<1	71	132	16	128	0.52
71-1-145	PT		<15	<2	<1	788	1250	180	1910	1.87
71-1-387	GNpeg		<15	<2	<1	53	20	12	13	0.16
71-1-392	GN	ds	<15	<2	27	106	247	25	139	0.41
71-1-91	PT	mx	<15	<2	<1	681	882	154	1820	1.85
71-7-377	GNpeg	mx	49	160	<1		580	110	240	1.89
71-7-384			<15	<2	2	43	130	26	64	0.18
71-7-416			<15	<2	4	2650	1375	167	3820	1.60
71-7-450	N peg		<15	<2	4	555	40	12	95	0.05

Pt, Pd, Au in ppb; Cr, Cu, Co, Ni in ppm; S in wt%; G: gabbro; GN:gabbronorite; N:norite; OG: olivine gabbro; PT: picritic troctolite; alt: altered; peg: pe

Table 3. Cu/(Cu + Ni) and (Ni + Cu + Co)/S ratios of the mineralized zones (this study)

		Cu/(Cu+Ni)		(Ni+Cu+Co)/S	
	n	x	sd	x	sd
Maliseet South	25	0.56	0.19	0.08	0.04
Maliseet Mt	14	0.50	0.09	0.03	0.02
Eastern body	9	0.49	0.16	0.15	0.10

n: number of samples; x: average; sd: standard deviation

Three mineralized zones have been identified in the western body (Fig. 2, Tables 1, 2). Potter (1985) indicated reserves, in excess of 15 million tonnes averaging 0.31 % Cu and 0.30 % Ni. The Maliseet Mountain and the Maliseet South zones are the most extensively explored zones. The Maliseet North zone does not seem to contain significant mineralization. The Maliseet Mountain zone consists of two lens-shaped ore bodies that appear to be continuous over 200 m along strike (Fig. 7). The ore is semi-massive to disseminated in character. One diamond-drill hole intersected about 2 meters grading 0.88 % Ni, 0.53 % Cu and 0 % Co. Another intersected 36 meters grading 0.22 % Ni and 0 Cu. Grades of other intersections are low. Company reports indicate 450,000 tonnes of low grade Cu-Ni mineralization in this zone (Tables 1,2). The Maliseet South zone consists of a northwesterly trending and southerly dipping ellipsoidal body (Fig. 8). This appears to be the most promising zone, several holes having 2 to 6 m intersections grading as high as 0 % Ni and 1.30 % Cu (Table 1). According to company reports about 5,000,000 tonnes of low grade (0.34 % Cu, 0.28 % Ni and 0.03 % Co) material have been outlined.

The eastern body contains Ni-Cu mineralization which consists mostly of matrix sulphides associated with peridotitic and noritic rocks. Assay results of MicMac Mines from limited mineralized zones indicate a low grade (i.e 0.36 % Ni and 0 Cu, see Table 1). The distribution of mineralization is not known.

On the basis of 48 analyses (this study, Table 2) and 469 assay results (MicMac Mines and Atlantic Coast, Table 1), the mineralized material of the Goodwin Lake intrusion clearly has low Ni and Cu abundances (Table 1, 2). This is also reflected in the pyrrhotite-rich nature of the ore. Maximum concentrations in samples containing up to 80 % (by volume) sulphide are about 0 % Ni and 3.14 % Cu. Co abundances up to 0.16 wt % were reported by MicMac Mines. Maximum Co value obtained in this study is 0 wt % . Somewhat elevated Bi abundances (e.g 450 ppm) reported by MicMac Mines have not been reproduced in this study. Twelve samples analyzed for Bi have abundances below the detection limit (i.e. 3 ppm). Pt, Pd and Au abundances are generally low (Table 2). Maximum concentrations obtained among 48 samples are 53 ppb Pt, 160 ppb Pd, and 333 ppb Au. Other platinum group elements (i.e. Os, Ir, Ru, Rh) are also low (e.g. 3 ppb Os, 0.1 ppb Ir, 25 ppb Ru, and 1 ppb Rh).

Cu/(Cu+Ni) ratio of the mineralized samples ranges from 0 to 0.8, averaging about 0.58 (Table 1). This is slightly higher than the Cu/(Cu+Ni) ratios of the St.Stephen (Paktunc, 1987) and the Moxie sulphides (Thompson, 1982). Overall, the Goodwin Lake sulphides display Cu/(Cu+Ni) ratios similar to those of flood basalt-related ores (see Naldrett and Cabri, 1976). The lower end of the range appears to correspond to the Sudbury range (0.44-0 given by Cabri and Naldrett (1984). Metal (Ni, Cu and Co) to sulphur ratios are low as expected from the pyrrhotite-rich nature of the sulphides. The Eastern body has the highest (Ni+Cu+Co)/S ratio, probably reflecting the fact that the mineralization is predominantly in peridotite (Table 3). The composition of the sulphide melt existing in various parts of the intrusion, perhaps at different times apparently was dependent on the composition of the hosting silicate magma.

Se/S ratios of the sulphides display a range from 26×10^{-6} to 110×10^{-6}. These values are considerably lower than those representative of mantle values (230 to 350×10^{-6}, Eckstrand and Hulbert, 1987). The $d^{34}S$ values ranging from 2.7 to 18.9 per mil also deviate significantly from the range defined for igneous or mantle derived sulfur (Ohmoto and Rye, 1979). The overall low metal: sulphur ratio and preliminary sulphur isotope analyses and Se/S ratios of the sulphides suggest assimilation of country rock sulphides as a likely process in the formation of low grade Cu-Ni sulphide ore.

The nature of the mineralization in the Goodwin Lake intrusion, in general, shows a strong resemblance to sulphides in the St.Stephen intrusion (Paktunc, 1987) and the Burnt Nubble sulphides of the Moxie intrusion (Thompson, 1982).

ACKNOWLEDGMENTS

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Fluid-saturation textures and Rb-Sr isotopic data from the East Kemptville tin deposit, southwestern Nova Scotia

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Richardson, J.M., Bell, K., Blenkinsop, J., and Watkinson, D.H., Fluid-saturation textures and Rb-Sr isotopic data from the East Kemptville tin deposit, southwestern Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 163-171, 1988.

Abstract

Fluid-saturation and unidirectional solidification textures at the East Kemptville tin deposit provide a strong genetic link between the granitic magma and the stanniferous aqueous fluid phase that produced the tin-bearing greisen. The apparent lack of multiple comb layers and aplite banding suggests that fluid pressures in the magma were not high enough to cause hydraulic fracturing. Consequently, ore-forming fluids were retained within the granite below either the relatively impermeable Meguma Group metawacke or eroded stockscheider. Whole-rock Rb-Sr isotopic data on massive greisen with $^{87}Rb/^{86}Sr < 100$ *yield an age of 337 ± 5 Ma (MSWD=2.87) and an initial* $^{87}Sr/^{86}Sr$ *ratio of 0.729 ± 0.001. These textural and isotopic data reflect closed-system conditions that were preserved during the transition from magmatic to hydrothermal conditions and the formation of tin-bearing greisen. Samples with* $^{87}Rb/^{86}Sr > 100$ *show a loss of* ^{87}Sr *that can be attributed to regional post-300 Ma tectonothermal activity.*

Résumé

Des textures de solidification unidirectionnelles causées par la saturation en fluides, identifiées au gîte d'étain d'East Kemptville, fournissent un lien solide de nature génétique entre le magma granitique et le fluide aqueux stannifère responsable de la formation du greisen d'étain. Le manque évident de couches de filons multiples et de rubanement à aplite semble indiquer que les pressions exercées par les fluides dans le magma n'étaient pas assez élevées pour provoquer une fracturation hydraulique. Des fluides minéralisateurs ont été plutôt retenus à l'intérieur du pluton, soit au-dessous du métawacke relativement imperméable du groupe de Meguma, soit en bas du stockscheider érodé. Les données isotopiques obtenues par la méthode Rb-Sr appliquée à la roche totale pour un greisen massif dont $^{87}Rb/^{86}Sr < 100$ *donnent un âge radiométrique de 337 ± 5 Ma [MSWD (carré moyen des écarts pondérés) = 2,87] et un rapport initial* $^{87}Sr/^{86}Sr$ *de 0,729 ± 0,001. Ces résultats montrent que les conditions d'un système fermé ont été préservées pendant la transition d'un contexte magmatique à un contexte hydrothermal et la formation de greisen stannifère. Des échantillons dont* $^{87}Rb/^{86}Sr > 100$ *montrent une perte de* ^{87}Sr *imputable à l'activité tectonothermique régionale survenue depuis 300 Ma.*

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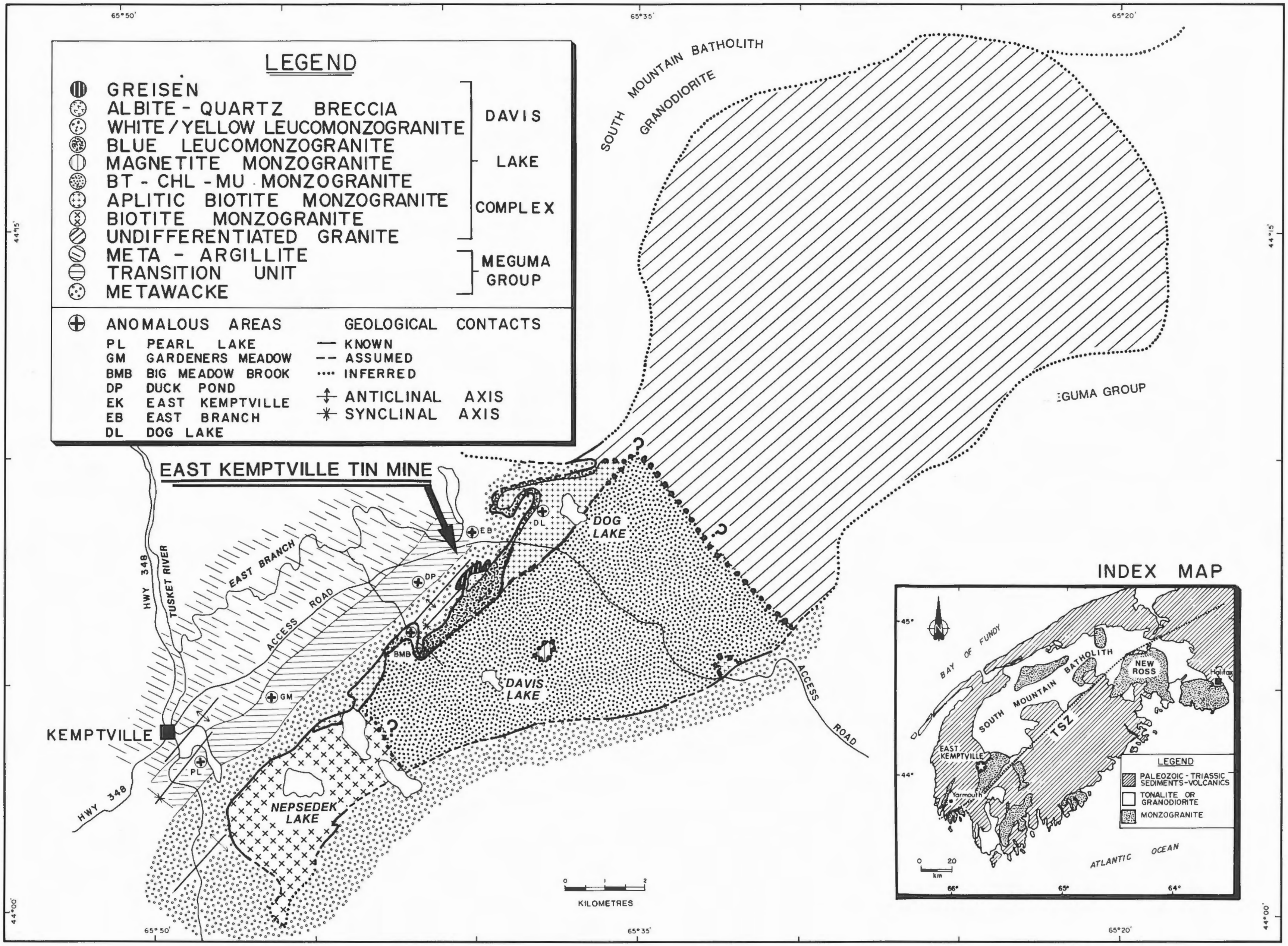

Figure 1. Location and regional geology of the Davis Lake complex and East Kemptville tin deposit, southwestern Nova Scotia. T.S.Z. Tobeatic Shear Zone (Giles, 1985).

INTRODUCTION

The East Kemptville cassiterite deposit (58 million tonnes of 0.165 % Sn), North America's only producing primary tin mine, is located in southwestern Nova Scotia (Fig. 1) below the contact of granitic rocks of Carboniferous Davis Lake complex (DLC) with metawacke of the Cambro-Ordovician Meguma Group (Fig. 2). The mine, owned by the East Kemptville Tin Corporation and managed by Rio Algom Ltd., was acquired from Shell Canada Resources Limited in 1982. Recent open-pit mining has produced new exposures that help clarify textural and structural relationships within the deposit.

NEWLY-EXPOSED GEOLOGICAL RELATIONSHIPS

Late magmatic textural features

Late magmatic features such as unidirectional solidification textures (USTs), stockscheider (marginal pegmatite or pegmatite caps), miarolitic cavities, orbicules, pegmatite segregations and dykes commonly form during the transition from magmatic to hydrothermal conditions (Moore and Lockwood, 1973; Lofgren and Donaldson, 1975; Shannon et al., 1982; Kirkham and Sinclair, in press). These features place constraints on pressure and temperature variations that occur during this transition and the chemical compositions of coexisting magmatic aqueous fluid and silicate magma. Such features and fluid- saturation textures have not been reported from the East Kemptville deposit, probably because early geological observations were based dominantly on diamond drilling and limited surface exposure of the actual contact zone. Exposures provided by recent open-pit mining reveal late magmatic textures such as pegmatitic segregations and a probable UST in addition to the felsite dykes reported by Richardson et al. (1982).

Elliptical pegmatitic segregations (1-15 cm) (Fig. 3) were observed in the deposit below the southern roof pendant (Fig. 2). These consist of coarse- grained, pink perthitic microcline, albite, quartz and/or muscovite, pyrite, apatite, cassiterite, molybdenite, chalcopyrite, sphalerite, pyrrhotite and fluorite. Three types of segregations that are probably gradational were classified according to mineralogy and by the presence or absence of an alteration halo. The margins of each segregation type are characterized by coarsening of the grain size of the associated leucomonzogranite over 1 to 4 mm. These features are considered pegmatitic segregations and not miarolitic cavities because at present, they do not have open space in the central region.

Type 1 segregations contain pink perthitic microcline, white albite, bluish-grey quartz, minor amounts of green muscovite and accessory pyrite (Fig. 3a). They do not have an associated alteration envelope. Type 2 segregations consist of white, inwardly terminating albite subhedra at the margin and, within the central region, quartz, subhedral zoned cassiterite and green muscovite masses (Fig. 3b). Accessory minerals include pyrite, chalcopyrite and rosettes of molybdenite. Microcline was not observed. Individual segregations may crosscut greisen-bordered zones or be crosscut, offset and sericitized by these zones. Type 2 segregations also lack alteration envelopes. Type 3 segregations are composed mainly of pyrite, sphalerite, cassiterite, chalcopyrite, fluorite and minor pyrrhotite. Quartz and muscovite are interstitial to the sulfides. Type 3 segregations have a faint halo of hematite that extends 2 to 20 cm outward into surrounding leucomonzogranite (Fig. 3c). Small fractures emanating from Type 3 segregations are greisenized (Fig. 3d).

Probable USTs were found near the northern roof pendant where the granite- metawacke contact is very convoluted (Fig. 2, Site 1), near the "pegmatite locality" mapped by Kontak et al. (1987). There, the granite-metawacke contact is exposed on the stoss side of a partly covered roche moutonnée. Icicle-like bluish quartz crystals 2-5 cm in length, are oriented perpendicular to the contact. Locally, the leucomonzogranite is strongly greisenized and minerals replaced by green muscovite (Fig. 4a, 4b). This feature may be similar to the marginal pegmatite or "stockscheider" described from other granite-related deposits (Taylor, 1979; Kirkham and Sinclair, in press). Minerals in these pegmatites may be oriented perpendicular to the contact and are thus a type of UST (Aubert, 1969; Kirkham and Sinclair, in press; P.J. Pollard, pers. comm. 1987).

Numerous pale yellow white or green felsite dykes of quartz and potassic feldspar or albite cut leucomonzogranite. They range from 3 to 25 cm wide and up to 75 cm long. Commonly they bifurcate and may include leucomonzogranite. In contrast to the pegmatitic segregations, these dykes commonly have welldefined contacts with the leucomonzogranite. Locally, pink perthitic microcline crystals

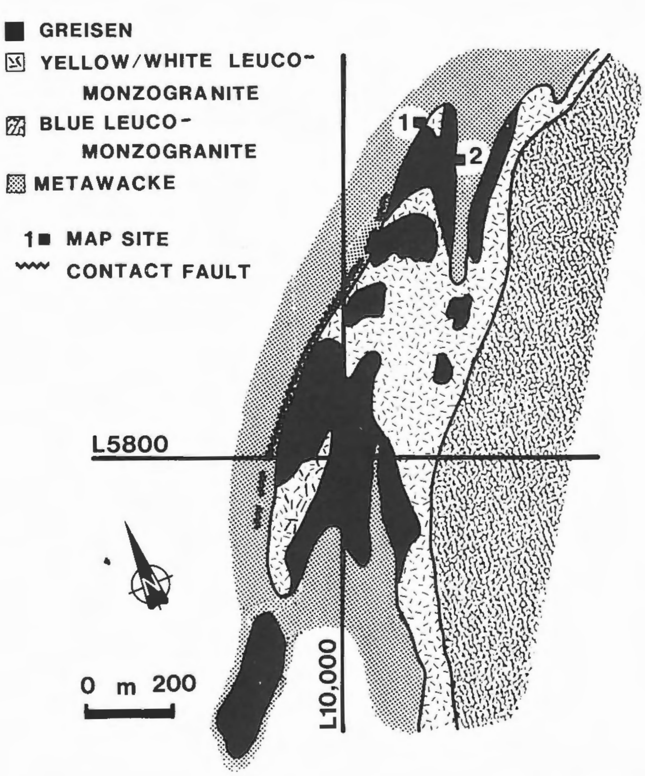

Figure 2. Surface geology, East Kemptville tin deposit. The stockscheider is located at Site 1. The area illustrated in Figure 5 is located at Site 2.

occur on the dike margins and terminate inward. These felsite dikes lack both metallic minerals and alteration envelopes, and are crosscut by greisen-bordered zones and later white quartz veins (Fig. 4c).

Extent and distribution of greisen zones

New pit exposures within the deposit reveal massive greisen zones (MGZ) up to 7 to 10 m wide, 50 m or more long and at least 15 m deep. MGZ are generally smaller at the granite-metawacke contact (Fig. 4d, 5). Greisen-bordered zones (GBZ) emanate from MGZ (Fig. 5, 6). Discrete GBZ that were traced 20 m along strike in the decline exceed 15 m in height in the pit. GBZ vary in width from several millimetres to 40 cm. The conjugate set of GBZ described by Richardson et al. (1982) was confirmed in three dimensions. GBZ parallel to the metawacke contact are approximately 1 m apart; those perpendicular to the contact, 2 m apart. Neither set appears to be preferentially enriched in metallic minerals. Elliptical concentrations of topaz and metallic minerals may occur at the intersection of these zones. Flat-lying greisen sheets such as those at the Anchor, Australia (Groves and Taylor, 1973) and the Cinovec, CSSR-GDR (Baumann, 1970) mines were not observed.

The Contact Fault and related shear zones

The Contact Fault, a large, well-developed fault-gouge zone 15 cm wide (Fig. 7a), is located near the granite-metawacke contact (Fig. 2, 7b). It is probably related to the "variable penetrative shearing event" documented by Kontak et al. (1987). In the deposit (L6230N 9925E), the fault is oriented

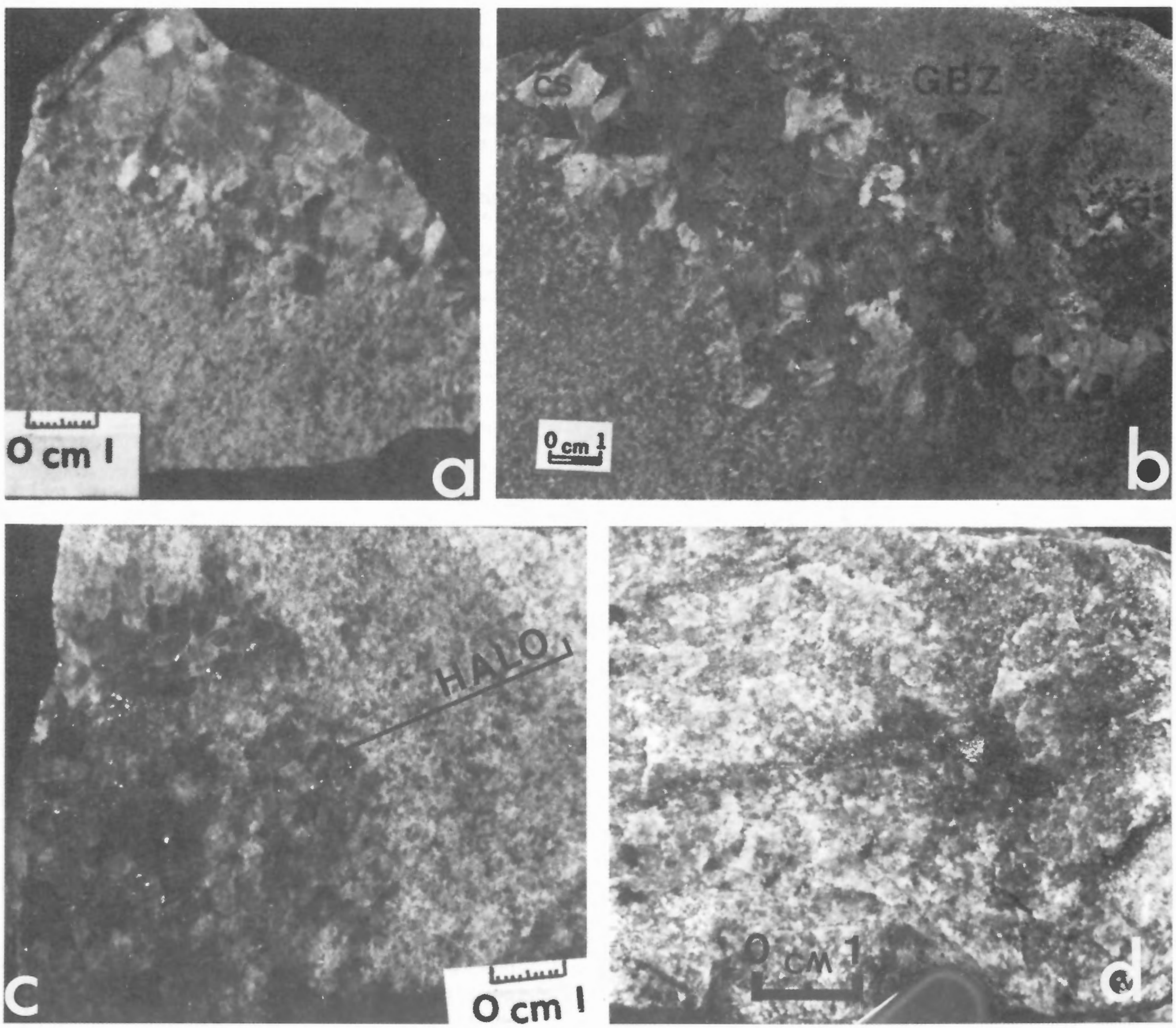

Figure 3. a) Type 1 segregation. b) Type 2 segregation (note cassiterite) crosscut by a greisen-bordered zone. c) Small Type 3 segregation with hematite alteration envelope. d) Small greisen-bordered zone emanating from Type 3 segregation.

046°/84W and clearly crosscuts the granite-metawacke contact (079°/70°W) (Fig. 8b). It therefore postdates the intrusion of the Davis Lake Complex and tin-bearing greisen. Orientation of shearing in the metawacke and cleavage in the leucomonzogranite and greisen near the granite-metawacke contact suggest that the fault parallels the granite-metawacke contact (and also therefore one set of GBZ) from at least L6000N to L6700N. The local cataclasis that caused brecciation, broken twin lamellae, subgrain development and extreme undulose extinction characteristic of the minerals in the granite, greisen and all segregation types is probably also related to this deformation.

Rb-Sr ISOTOPE GEOCHEMISTRY

Analytical methods

After the removal of weathered exteriors, large samples (2-5 kg) of greisen were crushed to <2 cm using a Chipmunk jaw crusher. Powders were prepared using a stainless-steel Bleuler mill. Rb and Sr were separated using standard ion-exchange techniques. Sr fractions were eluted a second time to remove excess Rb. Isotopic measurements were made using a high-precision Finnigan-MAT 261 multi-collector solid-source mass spectrometer at Carleton University. At the 1σ level, precision was 0.3 % and 0.8 % for Sr and Rb contents, 1 % for $^{87}Rb/^{86}Sr$ and 0.002 % for $^{87}Sr/^{86}Sr$. Blanks contained 0.4 ng Rb and 10.5 ng Sr. The measured $^{87}Sr/^{86}Sr$

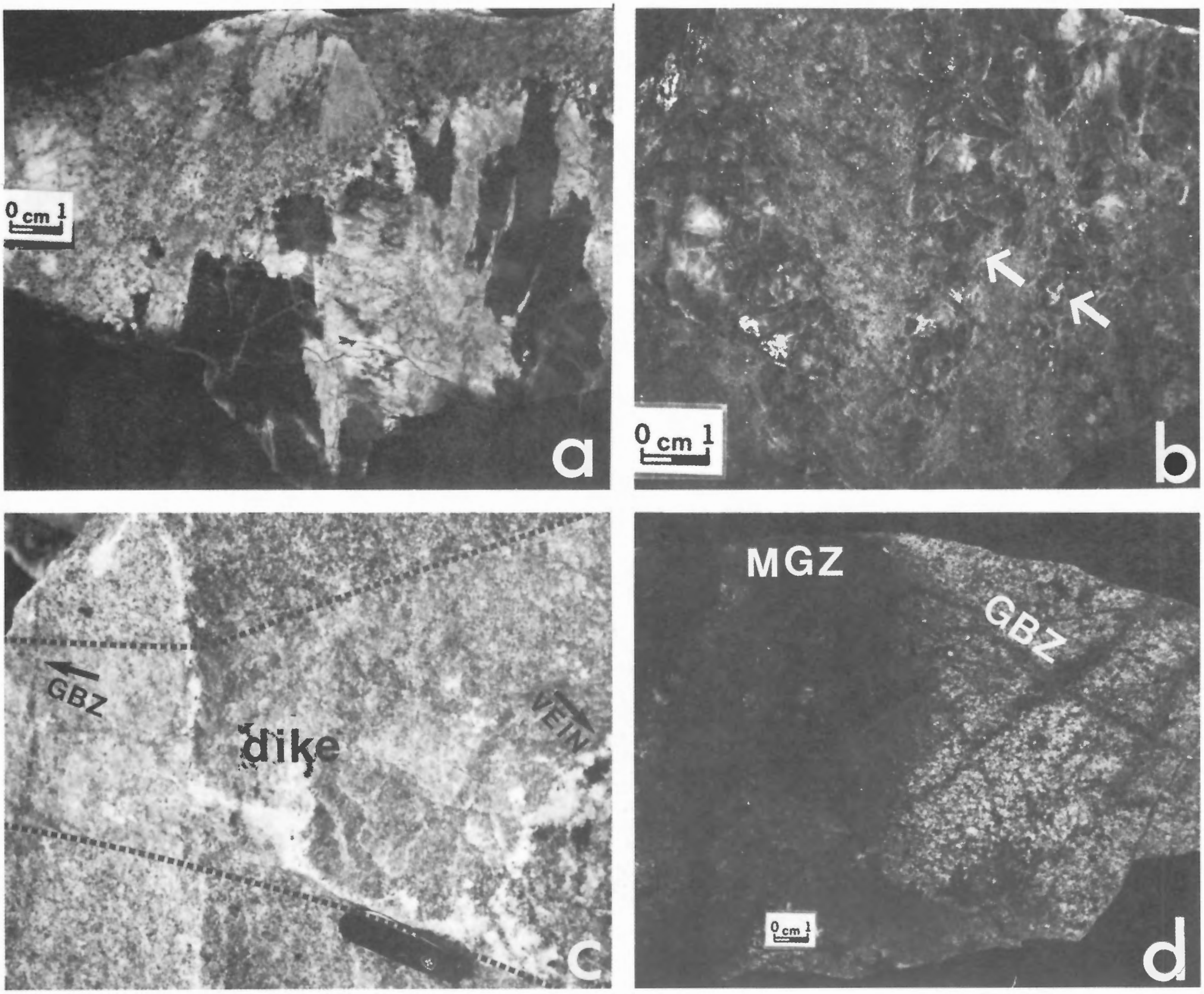

Figure 4. a) Ungreisenized unidirectional solidification texture, Site 1. b) Greisenized unidirectional solidification texture, Site 1. c) Felsite dyke cut by greisen-bordered zone and later white quartz veins. d) Greisen-bordered zone emanating from small massive greisen zone. Sample from immediately beneath granite-metawacke contact.

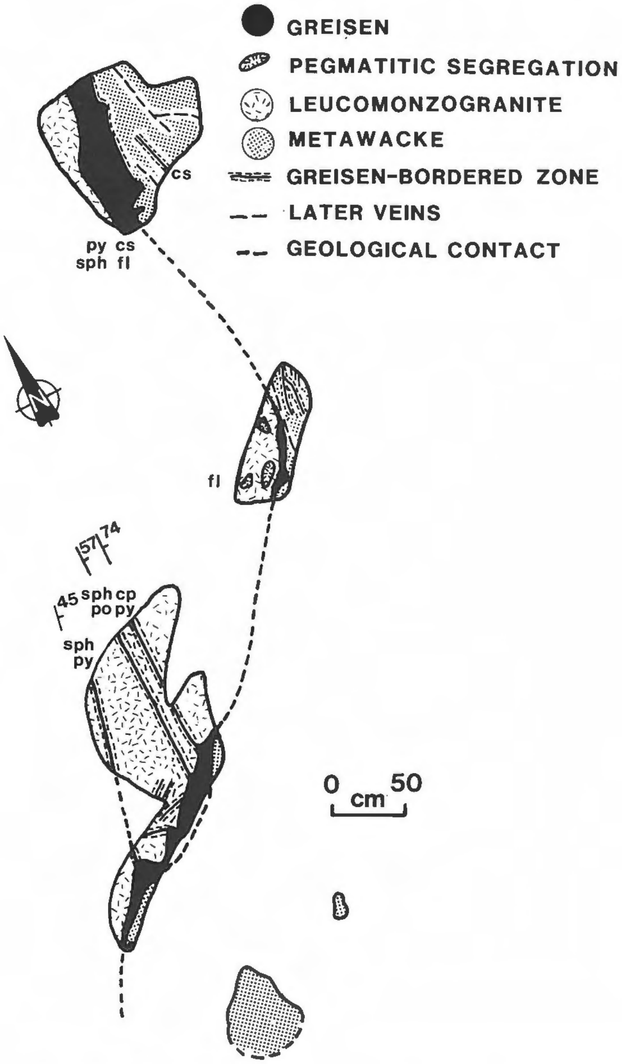

Figure 5. Field relationships at the granite-metawacke contact, Site 2. Note greisen- bordered zone emanating from small massive greisen zone and presence of pegmatitic segregations.

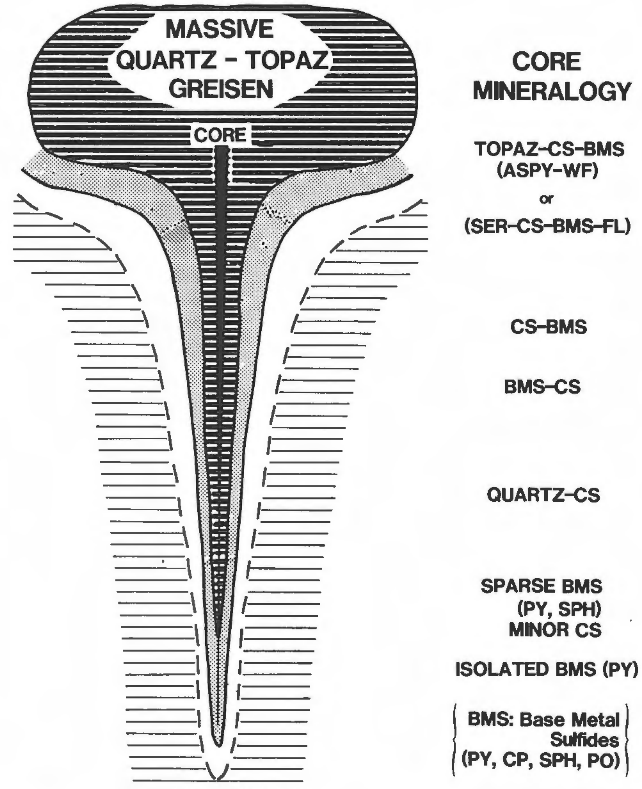

Figure 6. Schematic representation of relationship between massive greisen zones and greisen-bordered zones showing concentric alteration zones. The mineralogy of the central core varies along strike as indicated. From the outside, the alteration envelopes consist of incompletely greisenized leucomonzogranite, green muscovite-quartz, quartz-topaz-base metal sulphide minerals (BMS)cassiterite, topaz-quartz-BMS-cassiterite.

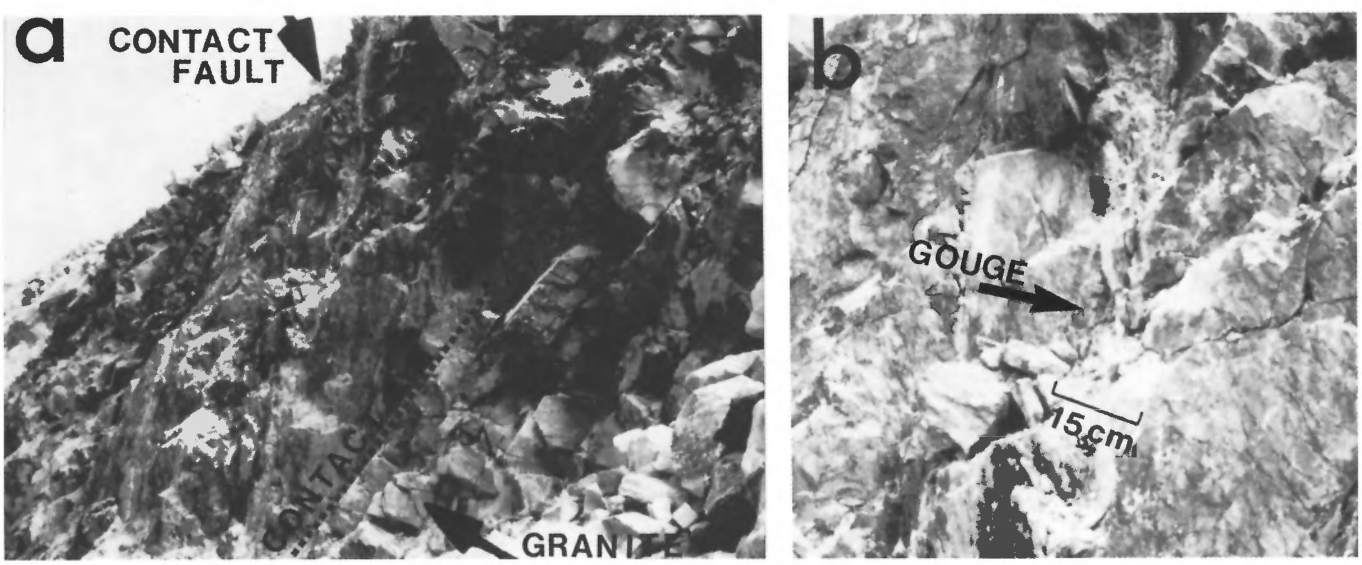

Figure 7. a) Gouge zone, Contact Fault. Located at 6230N, 9925E in deposit. b) Contact Fault crosscutting granite-metawacke contact.

Table 1. Whole-rock Rb-Sr data for East Kemptville greisen.

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$ (atomic)	$^{87}Sr/^{86}Sr$ (atomic)
EK25	53.6	11.4	13.8	0.8321
EK31	87.5	14.5	17.7	0.8529
EK32	1270	34.4	111.8	1.2048
EK33		21.0		1.1620
EK33	741	21.0	104.7	1.1767
EK34	219	22.5	28.3	0.7787
EK35	602	41.0	43.7	0.9978
EK38	126	8.76	43.0	1.0810
EK39	578	19.8	92.5	1.6880

Note: Rb and Sr concentrations by isotope dilution.
Analyst: Jean M. Richardson

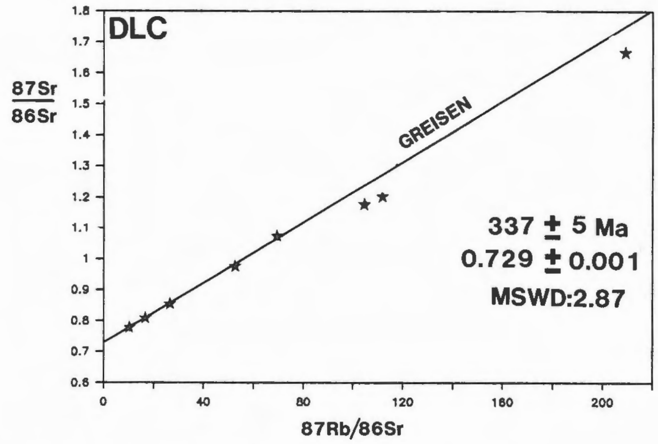

Figure 8. Isochron diagram for greisen.

ratios were normalized to an $^{88}Sr/^{86}Sr$ ratio of 0.1194. Over a three year period, the average $^{87}Sr/^{86}Sr$ ratio for 40 analyses of NBS 987 was 0.71023 + 0.00003 (2σ uncertainty). Isochrons were obtained using the ^{87}Rb decay constant of $1.42x10^{-11}a^{-1}$ and the methods of York (1969). Uncertainities were derived from the observed scatter of the data (York, 1966) for scatterchrons with an MSWD > 3, following the approach of Brooks et al. (1972). Errors are quoted at the 2σ level.

Interpretation

The Rb-Sr data for whole-rock samples of greisen are given in Table 1 and plotted in Figure 8. The Rb contents and Rb/Sr ratios are at least three times that normally encountered in peraluminous granitic rocks. These data yield an age of 317 ± 16 Ma (MSWD 25.3) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.733 ± 0.005. The large uncertainty associated with the initial ratio is magnified significantly beyond that of analytical precision because there are no data near the intercept.

Mineral separates from the Davis Lake monzogranite show open-system behavior (Harlow, 1981; Zentilli and Reynolds, 1985; Richardson, unpublished data) that can be correlated with regional post-300 Ma thermal events documented in the Meguma Terrane by Elias (1986). DLC biotite, like the East Kemptville greisen, has unusually high $^{87}Rb/^{87}Sr$ ratios. During post-Carboniferous thermal events, this mineral apparently lost a substantial amount of radiogenic Sr (Richardson, unpublished data). Samples with $^{87}Rb/^{86}Sr$ > 100 were excluded from the data reduction to assess the possibility of similar ^{87}Sr loss from greisen. The resultant isochron yields an age of 337 ± 5 Ma (MSWD=2.87) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.729 ± 0.001 for cassiterite-bearing greisen.

Such age and initial ratio for the greisen are within analytical uncertainity of recent Rb-Sr whole-rock determinations on the Davis Lake monzogranite. The monzogranite crystallized at 330 + 7 Ma (MSWD=2.82) with an initial $^{87}Sr/^{86}Sr$ ratio of 0.727 + 0.004 (Richardson, unpublished data). Both the biotite granite and the cassiterite-bearing greisen remained closed with respect to Rb and Sr at the whole-rock scale during the crystallization of the DLC magma and associated magmatic — hydrothermal events that occurred about 330 Ma. Samples with extreme $^{87}Rb/^{86}Sr$ ratios (> 100) likely reflect opensystem behavior that can be attributed to later thermal events.

DISCUSSION

Mineralogical variation in the pegmatitic segregations is due to separation and sequential chemical evolution of a magmatic aqueous fluid phase generated during the final stages of crystallization of the leucomonzogranite of the East Kemptville deposit. Types 1 and 2 segregations are not enclosed by alteration envelopes and thus were likely in equilibrium with the residual magma. Although hematitized perthitic microcline occurs only locally within the deposit, its presence in Type 1 segregations indicates that the residual magma from which these segregations crystallized was oxidized. This observation is supported by the mass transfer calculations of Candela (1986).

The inwardly terminating albite subhedra that characterize the margins of cassiterite- and molybdenite-bearing Type 2 segregations probably projected into cavities filled with magmatic aqueous fluid. Sn and Mo are thought to be transported in solution as chloro- and/or hydroxy- complexes (Jackson and Helgeson, 1985). The abundance of these elements in Type 2 segregations suggests that the aqueous fluid responsible for these segregations was also oxidizing and/or chlorine-rich.

The alteration zones that envelope Type 3 segregations are either small hematitic halos or greisenized leucomonzogranite that was altered to aggregates of quartz, green muscovite and minor fluorite. The presence of these minerals demonstrate that the fluids responsible for the Type 3 metal-, sulphur-, and fluorine-rich segregations were not in equilibrium with the enclosing leucomonzogranite. Wallrock microcline and albite apparently buffered the composition of the small segregations. Enough Ca was released during alteration to permit precipitation of fluorite. The presence of base-metal sulphides and pyrrhotite suggests that the fluids responsible for Type 3 segregations may have had a lower sulfur fugacity and/or were less oxidizing. MGZ and GBZ appear to be similar to Type 3 segregations, but MGZ and the central portions of GBZ are buffered only by wallrock quartz. These fluids were not buffered at a high Ca activity, thus topaz precipitated as opposed to fluorite.

Type 2 segregations both crosscut and are crosscut by GBZ. This suggests that fluid composition varied appreciably over a short time interval because these segregations do not have alteration envelopes and GBZ do. The fluid responsible for Type 2 segregations was oxdizing, in equilibrium with the magma, and saturated with respect to cassiterite and molybdenite. In contrast, the fluid responsible for Type 3 segregations was more widely distributed and obviously out of equilibrium with the leucomonzogranite. This second fluid was probably a precursor to the fluorine-, sulphur-, metal-rich greisen-forming fluid due to the mineralogical and chemical similarities between Type 3 segregations and the greisen.

USTs are used to document pressure and deformation regimes elsewhere (Moore and Lockwood, 1973; Kirkham and Sinclair, in press); however, at East Kemptville, USTs indicative of undercooling or pressure quenching, such as widespread comb quartz layers, aplitic bands and dendritic crystals, were not observed. Features indicative of plastic or brittle deformation such as crenulated comb layers, broken crystals and deformed orbicules, also were not observed. There are no USTs present that record multiple magmatic intrusions, such as at the Henderson mine (Shannon et al., 1982). Instead, only the documented pegmatitic segregations, remnant oriented stockscheider and stanniferous greisen zones are found beneath the granite-metawacke contact. As at the Sadisdorf tin deposit (Saxonian Erzgebirge district, GDR), the stockscheider is locally altered to quartz-muscovite greisen. These features reflect the passive concentration of magmatic aqueous fluids in the irregular apical region of the DLC, unaccompanied by pressure variations.

CONCLUSIONS

Late-magmatic features such as pegmatitic segregations and USTs link the granite and greisen of the East Kemptville deposit to a single magma that was passively emplaced at relatively low fluid pressures. Although the magma subsequently became fluid-saturated, fluid pressures were inadequate to result in extensive hydraulic fracturing and brecciation typical of many stockwork or porphyry-style deposits. Initially, the magma became saturated with an oxidized, stanniferous magmatic aqueous phase. The variable mineralogy and presence or absence of alteration envelopes that characterize the pegmatitic segregations illustrate the effect of changing fluid composition on leucomonzogranite of the East Kemptville deposit. As textural evidence indicates, the pressure did not change significantly; this compositional variation is tentatively attributed to decreasing temperature.

The foregoing textural evidence and the preservation of an isochron relationship in greisen with $^{87}Rb/^{86}Sr$ ratios < 100, appear to reflect closedsystem conditions during magma emplacement, crystallization, and subsequent greisen-style tin mineralization. The magmatic aqueous fluids generated by the Davis Lake leucomonzogranite and the subsequent derivatives were retained within the pluton, possibly due to the presence of an impermeable cap rock such as Meguma Group metawacke or possibly a now-eroded stockscheider.

ACKNOWLEDGMENTS

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Geochemical differentiation of Precambrian metacarbonate assemblages, Cape Breton Island, Nova Scotia[1]

Johannes R. Hill[2]

Hill, J.R., Geochemical differentiation of Precambrian metacarbonate assemblages, Cape Breton Island, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 173-185, 1988.

Abstract

Statistical evaluation of major and trace element geochemical data for marbles from Precambrian carbonate-detrital metasedimentary sequences of Cape Breton Island indicates that compositional variations between carbonate populations can distinguish specific assemblages. Three assemblages of metacarbonate rock are separated on metamorphic, structural, lithological and age criteria. Metacarbonate units underlying Craignish Hills/North Mountain are calcitic, of low metamorphic grade, and associated with detrital metasedimentary and metavolcanic members. The Cape Dauphin/Kellys Cove sequence, although similar to the Craignish units, has very calcitic members that contain relatively high base metal contents, possibly related to hydrothermal metasomatism. The Boisdale Hills assemblage, in contrast, is impure, siliceous metamorphosed marble with elevated major and trace element values; only locally is there evidence of contact metasomatic mineralization. Carbonates underlying the Cape Breton Highlands are a third type. These metalliferous marble units are unrelated to carbonates of southern Cape Breton Island.

Résumé

Une évaluation statistique de données géochimiques sur les éléments principaux et en traces pour certains marbres provenant de séquences métasédimentaires de nature carbonatée et détritique du Précambrien de l'île du Cap Breton indique que certaines variations de composition entre des populations de roches carbonatées peuvent permettre de distinguer des associations spécifiques. Trois associations de roches métacarbonatées sont séparées selon des critères de métamorphisme, de structure, de lithologie et d'âge. Les unités métacarbonatées sous-jacentes à celles de Craignish Hills et North Mountain sont calcitiques et de faible degré métamorphique, tout en étant associées à des membres métavolcaniques et métasédimentaires de nature détritique. La séquence de Cape Dauphin et Kellys Cove, bien qu'analogue aux unités de Craignish, possède des membres très calcitiques à teneurs en métaux de base relativement élevées, probablement dues à un métasomatisme hydrothermal. Par contre, l'association de Boisdale Hills est du marbre métamorphisé, siliceux et impur à fortes concentrations d'éléments principaux et en traces; ce n'est que par endroits que l'on détecte une minéralisation métasomatique de contact. Les roches carbonatées sous-jacentes aux hautes terres du Cap-Breton constituent le troisième type. Ces unités de marbre métalifère ne sont pas apparentées aux roches carbonatées du sud de l'île du Cap-Breton.

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

[2] Northwood Geoscience Ltd., Ottawa.

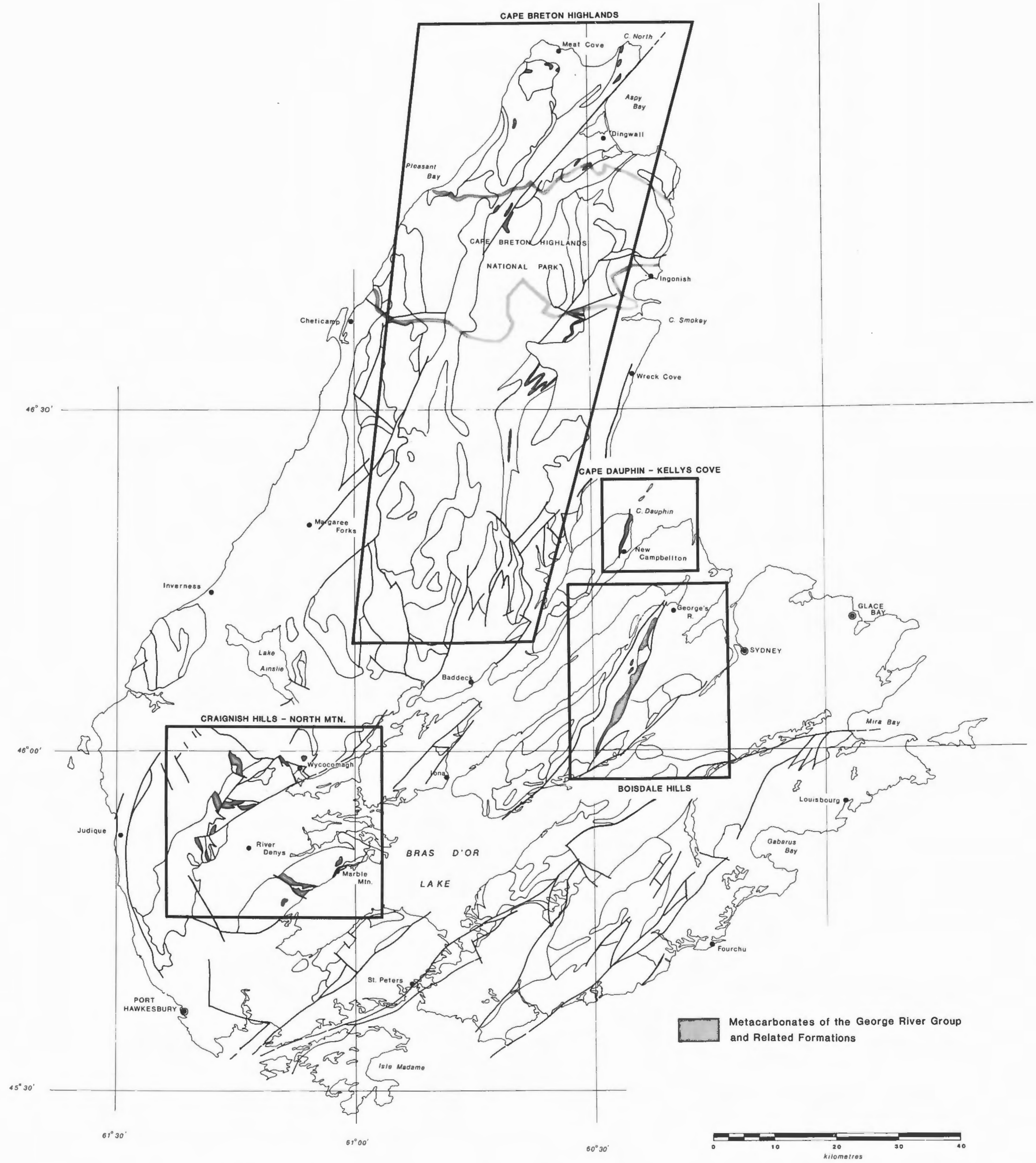

Figure 1. Location of study areas in Cape Breton Island underlain by Precambrian metacarbonate units of the George River Group and associated Late Proterozoic sequences. Simplified geology taken from Keppie (1979) and Barr et al. in press.

INTRODUCTION

A project initiated in 1986 to study the geology, geochemistry and potential of Precambrian carbonates to contain undiscovered mineral deposits has generated an extensive lithogeochemical database on marble formations for much of Cape Breton Island. The project was designed to increase the awareness of the mineral industry to the resource potential of the George River Group and related metacarbonate rocks of Cape Breton Island, and provide data for mineral potential evaluations.

The work entailed detailed lithogeochemical sampling and geological mapping, on a systematic regional basis, of the complex variety of carbonate units within the George River Group and related formations, as well as detailed examination and sampling of known areas of metallic and nonmetallic mineralization. Hill (1987a) summarized preliminary geological results from the 1986 survey showing that carbonate sequences examined throughout Cape Breton may be placed into one of at least three distinct metamorphic/stratigraphic assemblages. The geochemical data tend to support this grouping using selected major element oxides and trace metals to define compositional variations between major assemblages of metacarbonate rock. In addition, the database identifies marble formations or environments of specific chemical composition which may be considered economically significant. This paper presents the results of statistical evaluation of the geochemical data for Cape Breton marbles and discusses the compositional variations shown by marble assemblages interpreted according to geological and economic criteria.

STUDY AREAS AND PROCEDURES

To date, samples have been collected from four areas of Cape Breton Island (Fig. 1) each of which is underlain by a unique assemblage of mixed carbonate and detrital metasedimentary rock distinguished on the basis of lithology and stratigraphy. Three of the areas including Craignish Hills/North Mountain, Boisdale Hills and Cape Dauphin/Kellys Cove contain metasedimentary assemblages which have been correlated with the George River Group by Milligan (1970). Lithologies include: a) crystalline limestone and dolomite to marble, arenaceous and argillaceous metasedimentary rocks with subdominant metavolcanic rocks in the Craignish Hills/North Mountain area (with the exception of Lime Hill); b) approximately equivalent thicknesses of moderate metamorphic grade calcitic, dolomitic and siliceous marble and arenaceous sedimentary rock throughout the Boisdale Hills; and c) a narrow belt of dolomitic and siliceous calcitic marble in the Cape Dauphin-Kellys Cove area of the southeastern Cape Breton Highlands. Of more tenuous affinity are metacarbonate units which occur throughout the Cape Breton Highlands as discontinuous interlayers of calcareous, dolomitic and siliceous marble within argillaceous and arenaceous metasedimentary rocks of variable metamorphic grade. The study does not include carbonate samples from Zn-deposits hosted within inferred "pre-George River" basement complexes i.e. the Grenvillian marbles at Meat Cove and the inferred basement marbles underlying Lime Hill.

Table 1. Summary of analytical techniques component method detection limit

Component	Method	Detection Limit
major element oxides	XRF	0.01%
whole rock minor elements	XRF	10.00 ppm
Cu	DCP	0.05 ppm
Zn	DCP	0.05 ppm
Pb	DCP	2.00 ppm
As	FAA	0.01 ppm
Sb	NA	0.20 ppm
Sn	XRF	3.00 ppm
W	NA	1.00 ppm
Mo	DCP	1.00 ppm
Hg	CV	5.00 ppb
Ag	DCP	0.05 ppm
Au	FADCP	1.00 ppb

XRF = X-ray fluorescence; DCP = directly coupled plasma; FAA = flame atomic absorption; NA = neutron activation; CV = cold vapour; FADCP = fire assay, directly coupled plasma.

Within each area, an attempt was made to sample the complete suite of characteristic lithologies representative of each assemblage whereby mineralogy, colour and texture were used to distinguish discrete marble types. For each rock sample, lithologic characteristics including colour, grain size, texture, structure, alteration, type of mineralization, and dominant mineralogy were recorded as alphanumeric codes suitable for computer handling. A complete listing of all recorded data was presented in Hill (1987b).

A total of 328 samples (including duplicates and standards) were collected of which 247 are of carbonate composition. The samples were analyzed using a variety of geochemical techniques for the major element oxides plus L.O.I., Ba, Sr, Nb, Y, Zr, Cu, Zn, and Pb with optional analysis of selected samples for Sn, W, Mo, Hg, As, Sb, Ag, or Au. The following analytical techniques outlined in Table 1 were used by XRAY Assay Labs Ltd., Toronto.

Control reference standard samples were inserted at a frequency of one in ten samples to test laboratory precision. Two types of standards were used — a very pure dolomitic marble and a siliceous dolomitic marble containing base metal sulphides. In addition, blind laboratory duplicate samples consisting of a split of the previously crushed and ground rock sample were inserted at a frequency of one in every fifteen analyses. Results indicate a high degree of precision in reproducibility of geochemical values for all elements throughout the duration of analytical work.

GENERAL GEOLOGY

Historically, all marble formations mapped throughout Cape Breton were correlated with the George River Group regardless of lithologic composition, metamorphic grade or geological association. In reality, the group is a loosely- defined stratigraphic unit comprising interbedded carbonate and detrital sedimentary rocks with minor volcanic rocks of variable metamorphic grade and of probable Late Precambrian to pre-Middle Cambrian age (Milligan, 1970). Weeks (1954, p.8) stated that the name was probably first applied by Fletcher

(1877) in the George River area to describe metamorphosed calcareous rocks interbedded with a succession of detrital metasedimentary and intrusive units. However, Weeks alluded to problems in correlation between different areas underlain by George River rocks especially between the "type-section" at George River and the North Mountain and Eskasoni areas.

Metasedimentary rocks of the George River Group have been recognized throughout much of Cape Breton Island by two recurrent lithologies — arenite and carbonate. Within the areas examined, carbonate units range from relatively pure crystalline limestone and dolostone of very low metamorphic grade, as in the Craignish Hills and North Mountain areas, to more highly metamorphosed calcareous, dolomitic and siliceous (calc-silicate bearing) marble of the Boisdale Hills and Lime Hill areas. Associated sedimentary rocks include quartzite, impure quartzite and feldspathic sandstone which stratigraphically bracket carbonate members. Argillaceous and volcanic rocks are of more limited distribution.

Throughout the Cape Breton Highlands, more highly metamorphosed and deformed carbonate-clastic metasedimentary rocks are less obviously related to George River Group lithologies of southern Cape Breton Island. In the Cape North area, Macdonald and Smith (1980) tentatively correlated semipelitic and pelitic gneiss, marble and calc-silicate-bearing gneiss of the Cape North Group with the George River Group, even though arenaceous and calcareous facies represent a relatively minor component of the sequence. Elsewhere throughout northern Cape Breton Island, narrow metacarbonate units mapped by Neale (1963a,b; 1964a,b) and more recently by Barr et al. (in press) have been referred to by previous workers as George River equivalent (Murray, 1976; Keppie, 1979). In fact, carbonate lithologies here are greatly subordinate to pelitic, semipelitic and psammitic units. It is suggested that the equivalence of George River Group clastic-carbonate rocks (as identified in the southern half of Cape Breton Island) with higher grade metasedimentary sequences of the Cape Breton Highlands is tenuous at best (Barr, et al., 1985).

The presence of Grenvillian basement assemblages in Cape Breton Island has recently been confirmed by radiometric dating of lithologies in the extreme northeastern corner of Cape Breton Highlands (Barr, 1987). The assemblage includes isolated, high-grade metacarbonate roof pendants in a syenite intrusion. The roof pendants host large concentrations of low grade sphalerite mineralization (Meat Cove and adjacent Pine Brook deposits). Recent work has also shown that high grade carbonate and detrital metasedimentary rocks in the Lime Hill area reveal similarities in lithology and metamorphism (Justino and Sangster, 1987) and in metallogenic relationships (A.L. Sangster, personal communication, 1987) with the Grenvillian basement assemblages of northeastern Cape Breton Island.

Descriptions of the various lithologic members which form mappable units and a general dicussion of metamorphic and structural styles found in each of the main study areas can be found in Hill (1987a,b).

LITHOLOGIC CATEGORIZATION OF METACARBONATE ASSEMBLAGES

As a outlined by Hill (1987a), examination of selected Precambrian carbonate units covering a broad spectrum of sedimentary and metamorphic environments throughout Cape Breton Island has shown that, based on field criteria alone, carbonate sequences may be placed into discrete assemblages distinguished on the basis of metamorphic, structural, lithologic and age criteria.

One assemblage underlying the Craignish Hills, North Mountain (excluding the Lime Hill area) and the Cape Dauphin/Kellys Cove area is of low metamorphic grade and consists of thick, predominantly calcareous units associated with argillaceous and less commonly quartzitic metasedimentary and metavolcanic rocks — an assemblage representative of basinal sedimentary facies. Metavolcanic units, dominated by mafic to intermediate flows, are found interbedded with carbonate rocks throughout the Craignish Hills and Marble Mountain areas.

Another assemblage in the Boisdale Hills area is composed of siliceous calcitic and dolomitic carbonate units which have been subjected to much higher grade metamorphism, generating forsterite-diopside-tremolite or diopside- wollastonite mineral assemblages. Associated lithologies are dominated by arenaceous metasedimentary rocks, including quartzite and feldspathic sandstone.

Throughout the Highlands a third assemblage of narrow, discontinuous, calcitic to dolomitic and siliceous marble layers and lenses of the McMillan Flowage Formation, Middle River metamorphic suite and Cape North Group are associated with major thicknesses of moderate to high metamorphic grade (sillimanite zone) mica schist, quartz-feldspar-biotite gneiss and quartzitic units. Carbonate represents a very minor lithofacies within extensive, high-grade detrital metasedimentary sequences. Although the classic quartzite-carbonate association has locally been recognized, it becomes difficult to directly correlate carbonate units of Cape Breton Highlands assemblages with marbles of the George River Group to the south.

The highly deformed and metamorphosed basement complexes recognized in the northwestern corner of Cape Breton Island and in the Lime Hill area can be considered a fourth category of metacarbonate-bearing sequence. Formations such as the monticellite- and periclase-bearing, Zn-rich marbles at Meat Cove and Pine Brook of Grenvillian age and Zn-rich dolomitic marbles at Lime Hill of apparent pre-George River age are representative, although not included in the current study.

The geochemical data tend to confirm this field-based classification system. The relationships are examined in more detail in the following section.

GEOCHEMICAL CATEGORIZATION OF METACARBONATE ASSEMBLAGES

The first step in statistical evaluation of the geochemical data involved organization of the data into four subsets grouped according to study area. Each of the four areas is underlain by a unique assemblage of mixed carbonate-detrital metasedimentary rock of variable metamorphic grade and lithologic composition. Further subdivision of marble populations on the basis of CaO:MgO was possible only for the larger Craignish Hills and Boisdale Hills populations. Calcitic and dolomitic marbles were differentiated rather crudely in the field according to reaction with dilute HCl. While pure calcitic or pure dolomitic lithologies are easily identified by this method, marbles of intermediate composition (ie. 1.7 < CaO:MgO < 20.0) tend to be less accurately categorized. In addition, no attempt was made to classify the populations according to silica content although all skarn lithologies containing > 50 % calc-silicate and Fe-bearing minerals were excluded.

Histograms for calcitic and dolomitic marble from the Boisdale and Craignish areas illustrate the lithologic diversity of individual populations (Fig. 2, 3, 4, 5). For example, granitoid suite major element distributions (SiO_2, Al_2O_3, and TiO_2) from calcitic marbles are somewhat bimodal indicating the presence of an alumino-silicate-bearing calcitic marble population associated with cleaner marble units. This is especially obvious with SiO_2 in the Boisdale Hills suite but less so within the Craignish samples. Similar bimodal distributions are evident for MgO and CaO but, in this case, are a result of poor differentiation and classification of intermediate composition calcitic and dolomitic marbles.

Statistical measurements of the six populations have been summarized in Table 2. For each major element oxide and trace element, arithmetic mean, minimum, maximum and median values and standard deviation have been calculated.

The statistical data reveal a number of patterns in element distribution not obvious from field studies. For example, the alumino-silicate component, probably representing a detrital fraction in the carbonates, tends to be higher in calcitic than in dolomitic marbles. This is contrary to trends found in other Precambrian carbonate assemblages, for example those of the Grenville Province, where siliceous dolomitic marbles dominate the stratigraphic sequence. Base metals (Cu + Zn) tend to be concentrated in siliceous dolomitic marbles although anomalous concentrations can be found within all populations for example up to 1800 ppm Zn and

Table 2. Statistical summary of Precambrian marbles in Cape Breton Island, Nova Scotia.

	VALUES IN PERCENT											VALUES IN PPM									
PARAMETER	SIO2	AL2O3	CAO	MGO	NA2O	K2O	FE2O3	MNO	TIO2	P2O5	LOI	CR	SR	RB	Y	ZR	NB	BA	CU	ZN	PB
ARITHMETIC MEAN																					
1.	11.78	1.63	41.03	8.81	0.10	0.39	1.06	0.04	0.08	0.04	34.52	5.6	334.4	16.8	6.9	8.5	10.0	29.1	4.1	63.4	20.7
2.	10.79	1.41	34.23	15.55	0.16	0.24	0.77	0.06	0.07	0.08	36.51	5.4	204.7	15.3	6.6	9.0	10.4	37.9	11.3	152.8	35.3
3.	4.50	0.89	49.50	2.65	0.05	0.14	0.47	0.03	0.04	0.04	41.36	5.0	272.7	10.1	6.6	5.8	10.5	27.9	1.8	23.7	5.3
4.	3.57	0.41	31.20	19.58	0.06	0.05	0.66	0.06	0.03	0.04	44.43	5.0	117.9	10.4	6.2	5.0	10.7	18.6	0.6	11.0	2.2
5.	6.34	1.48	41.89	8.71	0.10	0.33	0.79	0.07	0.12	0.05	40.28	5.3	270.6	14.2	7.2	14.4	14.2	48.7	3.5	155.9	154.1
6.	15.37	1.86	36.95	9.88	0.13	0.34	1.63	0.15	0.11	0.13	33.61	7.9	367.9	21.0	9.5	12.9	11.6	80.8	8.8	192.4	13.9
MIN. VALUE																					
1.	1.29	0.21	20.70	1.24	0.02	0.02	0.12	0.005	0.01	0.02	10.39	5.0	120.0	5.0	5.0	5.0	5.0	5.0	0.25	4.0	1.0
2.	0.68	0.07	18.70	2.85	0.005	0.01	0.11	0.01	0.01	0.01	3.77	5.0	30.0	5.0	5.0	5.0	5.0	5.0	0.25	3.0	1.0
3.	0.65	0.05	30.70	0.20	0.005	0.01	0.10	0.005	0.005	0.01	24.39	5.0	40.0	5.0	5.0	5.0	5.0	5.0	0.25	1.5	1.0
4.	0.42	0.05	25.70	7.10	0.03	0.01	0.16	0.01	0.01	0.01	34.85	5.0	50.0	5.0	5.0	5.0	5.0	5.0	0.25	2.5	1.0
5.	0.64	0.09	30.10	0.63	0.03	0.01	0.005	0.005	0.01	0.01	25.85	5.0	20.0	5.0	5.0	5.0	5.0	5.0	0.25	4.5	1.0
6.	0.73	0.07	19.70	0.63	0.04	0.01	0.005	0.01	0.02	0.02	13.16	5.0	20.0	5.0	5.0	5.0	5.0	5.0	0.25	3.5	1.0
MAX. VALUE																					
1.	38.70	8.55	51.40	18.80	0.58	3.80	5.71	0.20	0.45	0.19	44.46	20.0	730.0	60.0	20.0	80.0	30.0	250.0	29.0	600.0	460.0
2.	49.60	12.10	51.50	23.70	1.20	1.86	4.25	0.19	0.56	0.31	46.16	20.0	890.0	80.0	20.0	100.0	30.0	410.0	170.0	1800.0	610.0
3.	28.50	5.12	55.10	12.90	0.62	2.07	2.93	0.28	0.26	0.41	44.54	5.0	1110.0	50.0	20.0	40.0	20.0	170.0	15.0	580.0	240.0
4.	15.80	1.47	45.00	23.00	0.10	0.21	2.44	0.16	0.09	0.20	46.50	5.0	300.0	30.0	20.0	5.0	20.0	190.0	5.0	46.0	20.0
5.	22.70	6.76	54.70	22.50	0.32	1.59	3.72	0.21	0.79	0.19	46.16	10.0	1000.0	40.0	20.0	100.0	30.0	300.0	40.0	1900.0	2600.0
6.	44.30	8.23	53.30	23.60	0.43	1.81	6.47	0.63	0.45	0.46	45.62	30.0	1110.0	90.0	30.0	120.0	20.0	580.0	41.0	2400.0	160.0
MEDIAN																					
1.	12.35	0.76	42.20	8.38	0.06	0.15	0.53	0.03	0.03	0.04	34.62	5.0	330.0	10.0	5.0	5.0	10.0	20.0	1.5	15.5	1.0
2.	4.43	0.44	31.10	19.10	0.07	0.08	0.58	0.04	0.03	0.04	40.62	5.0	150.0	10.0	5.0	5.0	10.0	20.0	1.0	30.0	2.0
3.	2.78	0.60	51.00	1.44	0.04	0.05	0.27	0.01	0.03	0.03	42.46	5.0	220.0	10.0	5.0	5.0	10.0	20.0	1.0	9.5	1.0
4.	1.68	0.33	30.45	20.50	0.06	0.02	0.48	0.06	0.02	0.03	45.50	5.0	100.0	10.0	5.0	5.0	10.0	7.5	0.25	8.2	1.0
5.	4.90	0.70	46.00	2.87	0.06	0.04	0.49	0.06	0.05	0.03	41.23	5.0	160.0	10.0	5.0	5.0	10.0	20.0	0.75	19.5	1.0
6.	12.00	0.88	33.20	5.89	0.08	0.11	0.79	0.14	0.07	0.13	36.23	5.0	200.0	10.0	5.0	5.0	10.0	40.0	2.0	42.0	1.0
STANDARD DEVIATION																					
1.	9.58	2.02	7.37	4.71	0.11	0.68	1.31	0.04	0.10	0.03	8.48	2.7	146.5	11.8	3.9	13.1	5.4	42.8	6.3	135.9	80.1
2.	13.02	2.41	7.97	7.32	0.27	0.42	0.81	0.04	0.10	0.08	11.07	2.3	164.0	14.8	3.5	16.9	5.4	72.2	28.2	350.9	109.8
3.	4.65	0.86	4.88	3.00	0.08	0.26	0.55	0.04	0.04	0.04	3.34	0.0	185.5	7.3	3.4	4.1	4.4	26.9	2.3	68.8	24.7
4.	4.65	0.33	3.49	3.37	0.02	0.06	0.54	0.04	0.02	0.04	2.74	0.0	63.3	6.2	3.2	0.0	4.8	35.7	1.1	9.4	3.7
5.	6.39	1.85	9.62	9.68	0.09	0.53	1.00	0.06	0.19	0.05	5.68	1.2	285.9	9.9	3.9	25.1	7.3	79.1	9.2	443.5	610.9
6.	14.44	2.09	11.16	8.33	0.11	0.53	1.77	0.15	0.11	0.10	10.51	7.1	357.0	22.0	7.6	26.6	4.7	137.6	13.4	540.6	37.6

1 = Calcitic Marbles of the Boisdale Hills (n = 34)
2 = Dolomitic Marbles of the Boisdale Hills (n = 45)
3 = Calcitic Marbles of the Craignish Hill (n = 103)
4 = Dolomitic Marbles of the Craignish Hills (n = 28)
5 = Marbles of the Kellys Cove Area (n = 18)
6 = Marbles of the Cape Breton Highlands (n = 19)

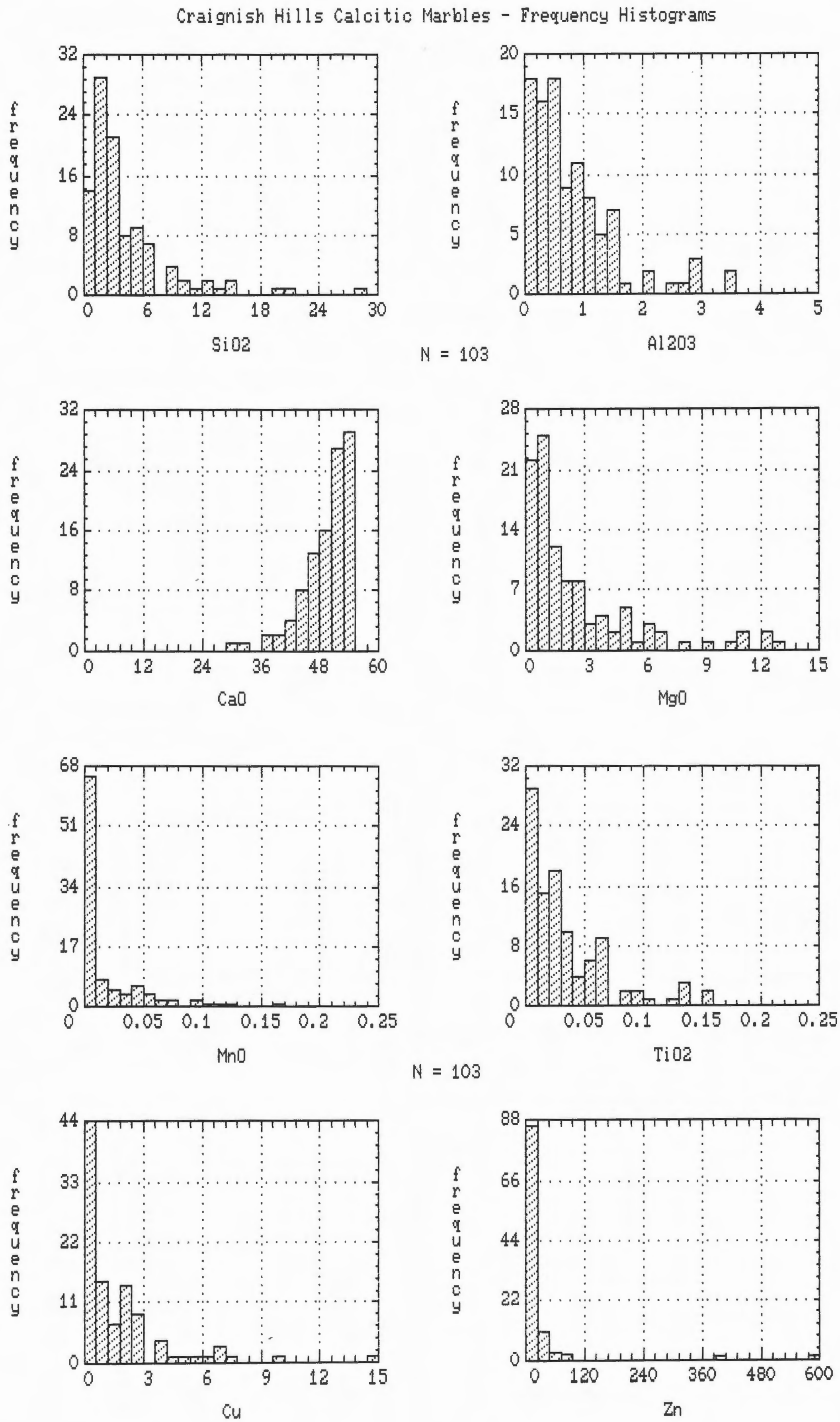

Figure 2. Frequency distributions for selected elements of the Craignish Hills/North Mtn. calcitic marble population (n = 103)(Cu and Zn in ppm, all other elements in %).

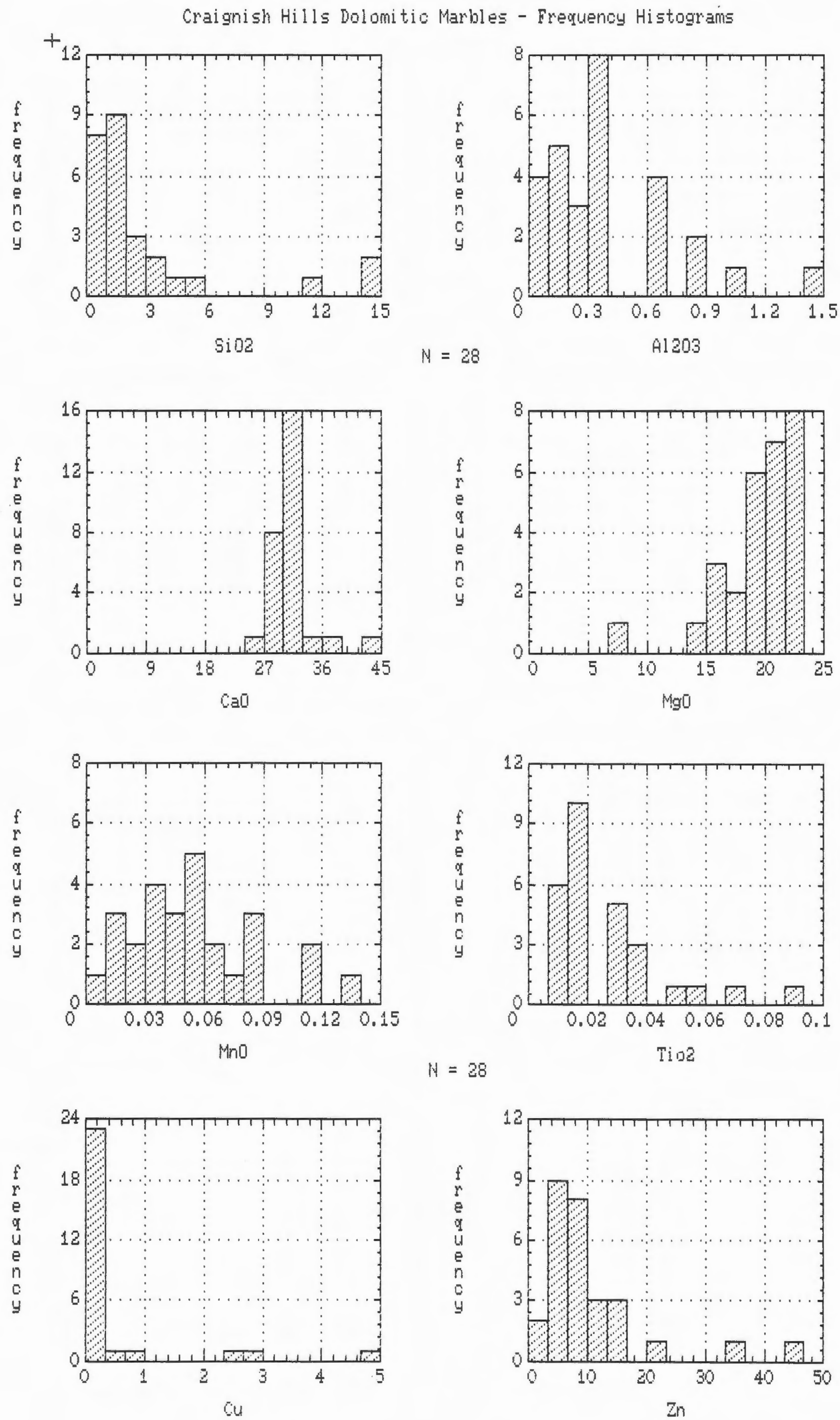

Figure 3. Frequency distributions for selected elements of the Craignish Hills/North Mountain dolomitic marble population (n = 28)(Cu and Zn in ppm, all other elements in %).

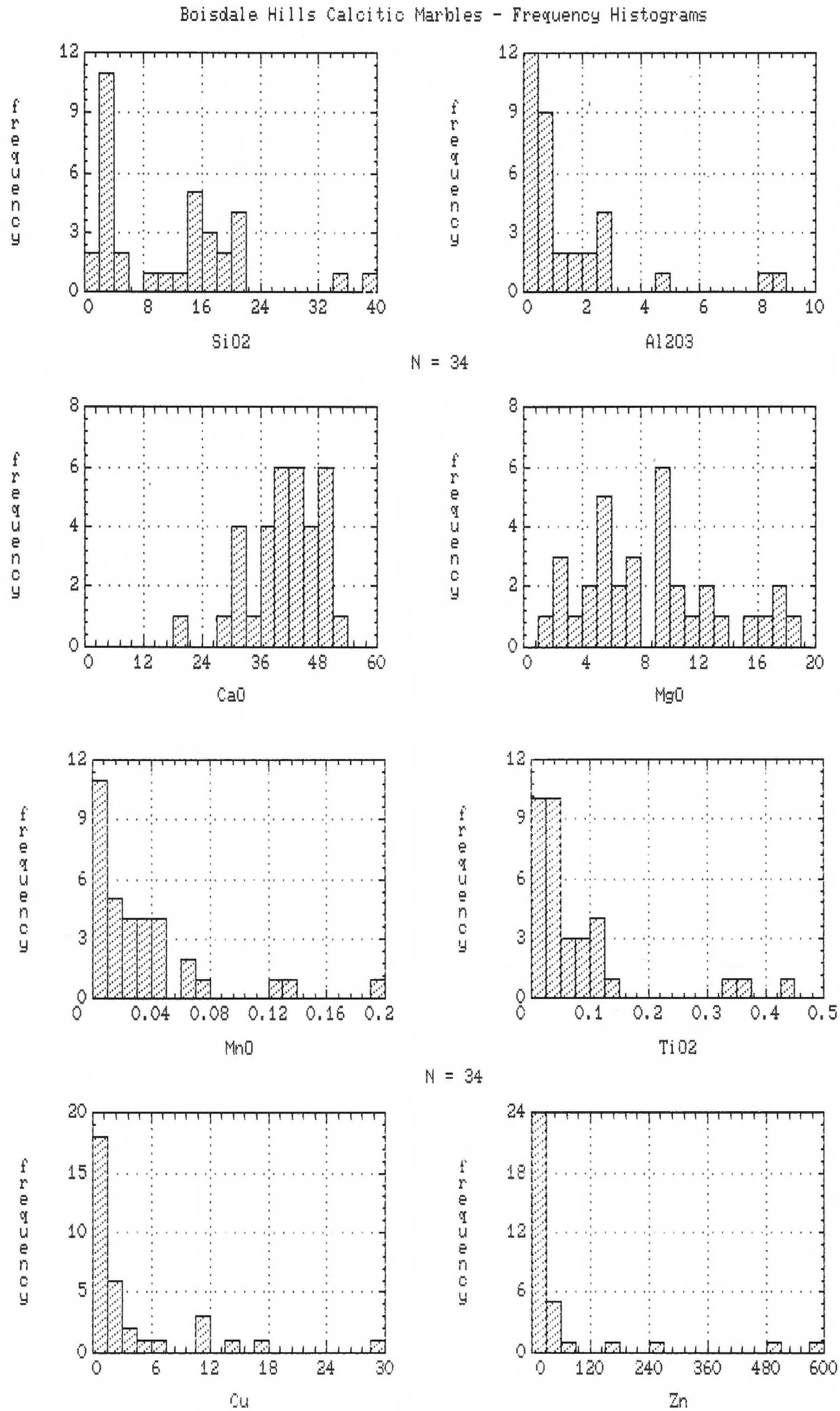

Figure 4. Frequency distributions for selected elements of the Boisdale Hills calcitic marble population (n = 34)(Cu and Zn in ppm, all other elements in %).

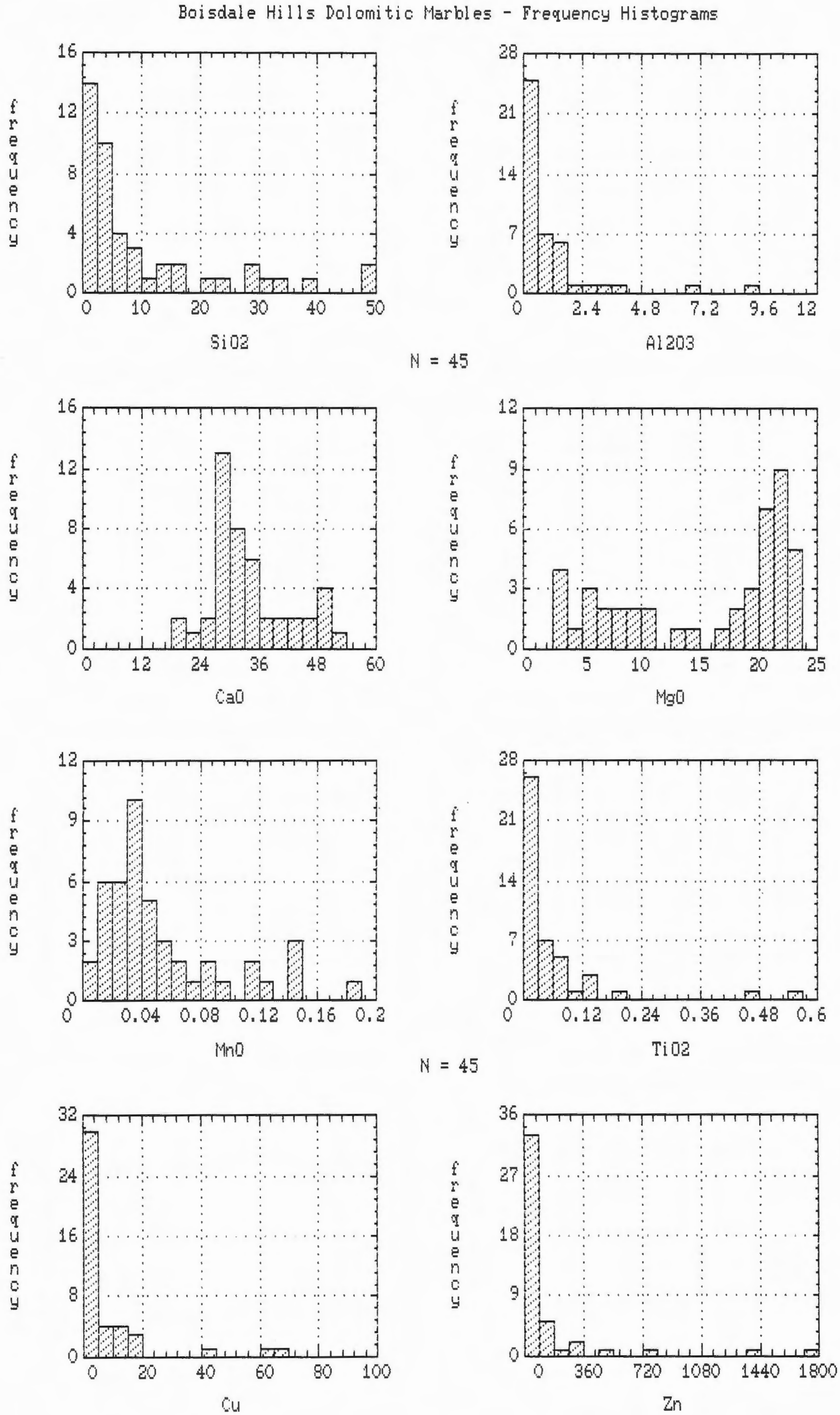

Figure 5. Frequency distributions for selected elements of the Boisdale Hills dolomitic marble population (n = 45)(Cu and Zn in ppm, all other elements in %).

170 ppm Cu in "pure" dolomitic marbles of the Boisdale Hills. In general, background Zn and Cu levels are higher in calcitic than in dolomitic marbles of the Craignish Hills while the opposite is true of the Boisdale population.

Based on arithmetic mean and median values for selected components, distinct chemical signatures define each of the carbonate populations underlying the four study areas. The relationship is best illustrated by a graphic representation obtained through the construction of notched box and whisker diagrams (Fig. 6a to 6f). The plots are defined by the following statistical parameters for each population (McGill et al., 1978):

1. upper and lower quartiles of the population define the top and bottom edges, respectively, of the box horizontally bisected by the median value (50th percentile);

2. "whiskers" cover sample points 1.5 times the interquartile range while extreme points beyond are plotted as individual adjacent values;

3. box widths are proportional to the square root of the number of observations in a sample population; and

4. notches correspond to the width of the 95th confidence interval and are a measure of the population's standard deviation.

Distribution patterns for SiO_2, Al_2O_3, MnO, TiO_2, Cu and Zn show the greatest degree of variability between populations. Internally, however, sample groups display unusual chemical uniformity, especially for SiO_2, MnO and Zn (Fig. 2a,d,f), in spite of variable lithologic composition. In fact, detailed comparisons of chemical and field data show that such lithologic features as colour and texture create only minor chemical diversity within any population.

The diagrams illustrate differentiation in geochemical signatures between carbonate assemblages of the Craignish Hills, Boisdale Hills and Cape Breton Highlands. The Kellys Cove assemblage, contrary to geological criteria, is closer in chemical composition to the Boisdale Hills carbonates than the Craignish marbles. The greatest contrast in chemistry is evident between the Boisdale and Craignish populations, showing significant variation in the median values of Si, Mn, Ti, Cu and Zn. Unfortunately, the small size of both the Cape Breton Highlands and Kellys Cove populations has generated a high degree of variance in their calculated median values. It is therefore difficult to distinguish Cape Breton Highlands marbles statistically from either Boisdale or Kellys Cove suites. It is expected that data from additional sampling completed in 1987 will create a more statistically valid database.

INTERPRETATION AND DISCUSSION

In addition to generating a purely quantitative categorization system for metacarbonate assemblages in Cape Breton Island, the geochemical database reveals more empirical relationships relevant to the development of a post-depositional model for carbonate evolution. For example, a large component of lithophile elements as shown by the Boisdale, Highlands and, to a lesser extent, the Kellys Cove carbonate assemblages can be considered evidence of either a large detrital sedimentary component in the original depositional environment or a later, hydrothermal, metasomatic event. The relative significance of depositional versus post-depositional processes in the development of observed chemical signatures can be approximated by examining dominant element associations in each carbonate assemblage using discriminant analysis. A correlation matrix is a simple but effective tool which targets those elements showing a high degree of interdependence. Element associations are summarized in Table 3.

Both the Craignish and Boisdale sequences are characterized by similar strong granitoid-element associations, unrelated to the chalcophile suite of Cu + Zn + Pb. This suggests that the present chemical composition of Boisdale and Craignish carbonate rocks is predominantly a reflection of the original sedimentary environment. The slight enrichment of Cr in the Boisdale suite may, however, be a function of metasomatism associated with the regular emplacement of mafic dykes characteristic of the area. In contrast, Kellys Cove marbles show strong correlation between granitoid and chalcophile elements, with a felsic potassic granitoid association similar in composition to the adjacent Kellys Mountain pluton. It is apparent that the assemblage possesses a chemical signature representative of a strong metasomatic overprint. Cape Breton Highlands carbonates show similar, although less well developed, correlation between Cu and granitoid elements. Contact metasomatic mineralization has been described at isolated localities throughout the Highlands.

METALLOGENIC SIGNIFICANCE OF GEOCHEMICAL DATA

The Craignish Hills assemblage, which includes carbonate members of the George River Group underlying North Mountain but not those underlying Lime Hill, represents a relatively clean, compositionally uniform, predominantly calcareous carbonate sequence associated with a basinal sedimentary environment. Background base metal values are low compared with marble formations found elsewhere and compared

Table 3. Element associations in carbonate populations based on correlation coefficients

Assemblage Correlation	Strong Pos. Correlation (avg. coeff. >0.65)	Weak Pos. (avg. coeff. 0.45 — 0.60)
Craignish	Si-Al-Na Al-Na-K-Ti-Zr Zn-Pb	Al-Ti-Ba
Boisdale	Si-Al-Na Al-Na-Ti-Zr-Cr Cr-Zr-Ti Al-K-Ti-Rb	Cu-Pb-Zn
Kellys Cove	Si-Al-K-Fe-Ti-P-Cr-Zr-Ba-Cu-Zn-Pb	
Cape Breton	Al-K-Ti-Cr-Rb-Y-Zr	Si-Na-Ti-Zr-Cu
Highlands	Al-K-Cu Zn-Pb	

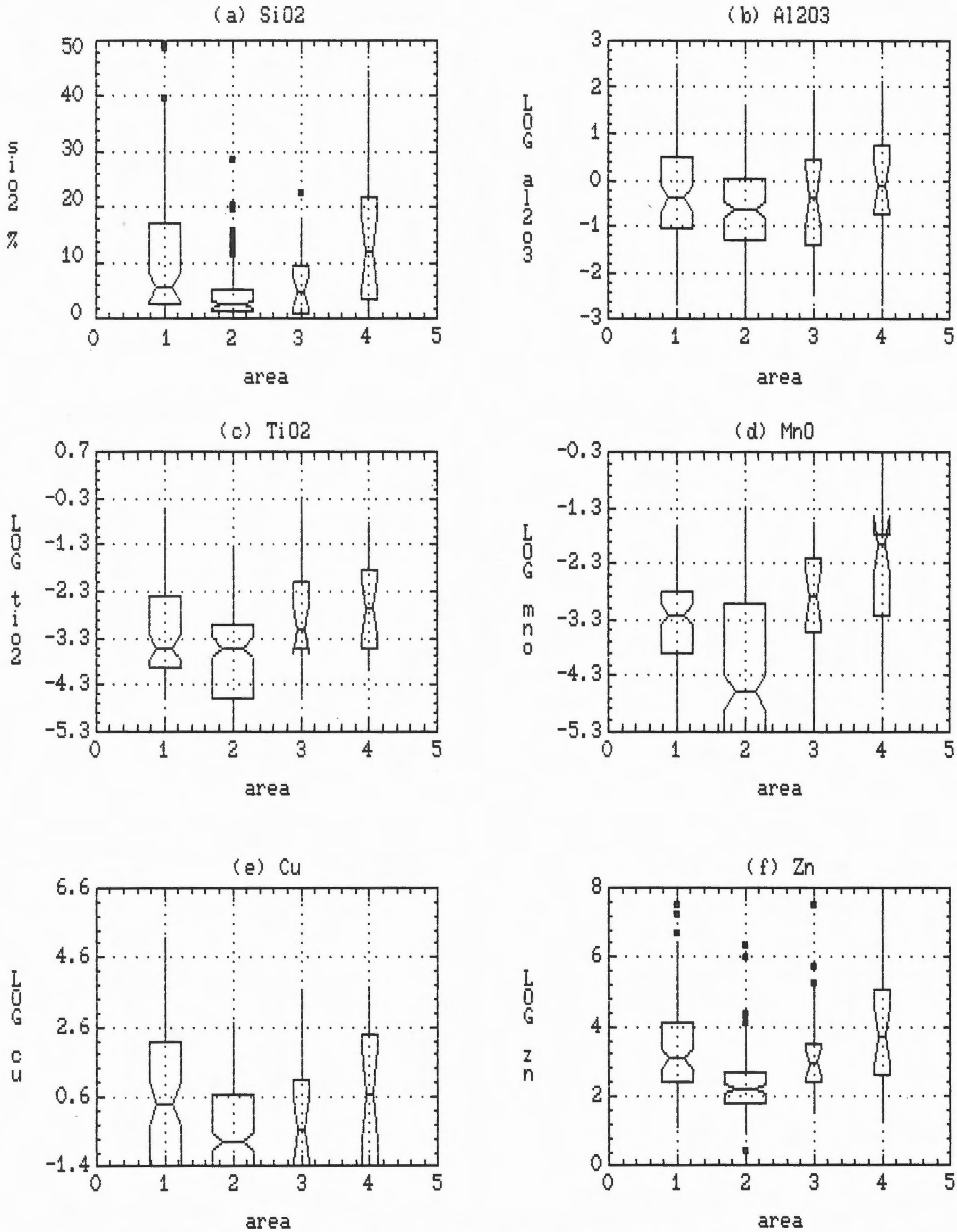

Figure 6. Notched box and whisker diagrams showing comparative stasistics for marble populations of the four study areas (1 = Boisdale Hills, 2 = Craignish Hills/North Mtn., 3 = Cape Dauphin/Kellys Cove, 4 = Cape Breton Highlands).

with average base metal values for limestones from Taylor (1964). Threshold values defined by Hill (1987b) for the classification of anomalous samples can also be considered low at 5 ppm Cu, 44 ppm Zn and 7 ppm Pb. It is apparent that no large-scale metasomatic event was associated with regional low grade metamorphism. Rare concentrations of base or lithophile elements examined throughout the area are associated within either indisputable contact metamorphic environments (Whycocomagh Cu-W-Mo and Glencoe skarns) or occur as isolated concentrations of sphalerite or galena of probable diagenetic origin.

The Boisdale Hills assemblage represents a "dirtier" carbonate sequence containing a greater detrital sedimentary component than the Craignish units. It can be considered more closely related to a nearshore, deltaic sedimentary environment. All carbonates of the Boisdale sequence have undergone a higher grade of regional metamorphism and only locally is there evidence of hydrothermal (contact) metasomatism. The higher background base metal content of these marbles, in comparison with the Craignish carbonates, is probably related to the original depositional environment or to diagenetic processes and does not appear to be evidence of a major regional metasomatic event associated with large-scale post-diagenetic plutonism. Background base metal values for carbonate units are 0.5 times (Cu, Pb) and 3 times (Zn) greater than the Craignish rocks while threshold values were calculated at 40 ppm Cu, 660 and 92 ppm Pb. The majority of metallic mineral showings are in fact of questionable contact metasomatic origin. Skarn lithologies have been reported in the Rear Boisdale Zn-Pb-Ag deposit (although not observed by the author) while a new massive sulphide showing examined in 1987 near Upper Leitches Creek is associated with a tremolite skarn. The remaining occurrences, most of which were examined in 1986, consisted of disseminated sphalerite $\pm$ chalcopyrite $\pm$ bornite $\pm$ galena or fracture-controlled magnetite $\pm$ chalcopyrite associated most commonly with the emplacement of mafic and/or felsic porphyry dykes.

The Kellys Cove assemblage is closely related by chemical composition to the Boisdale sequence but lithologically shows closer affinities to the Craignish Hills assemblage by the low grade of regional metamorphism plus associated argillaceous and quartzitic metasedimentary rocks. The high degree of correlation between granitoid and chalcophile elements suggests that extensive, post-depositional contact metasomatism, probably related to intrusion of the Kellys Mountain quartz monzonite, has affected at least the narrow band of calcitic marbles adjacent to the intrusion. Recent examination of an isolated Pb-Zn anomaly in siliceous calcitic marble identified poorly developed skarn lithologies at what is now a fault-controlled intrusive contact. It therefore appears that emplacement of the granitoid intrusion during the Late Hadrynian to Cambrian (Barr, et al., 1982) resulted in localized skarn development and base metal mineralization at the contact subsequent to fault displacement. Background values for Cu, Zn and Pb in carbonate units of the Kellys Cove area are similar to those characterizing Boisdale Hills marbles. In addition, the same threshold values were applied to base metal data from the two areas.

Carbonate members associated with high-grade metasedimentary sequences underlying the Cape Breton Highlands are characterized by the highest granitoid suite element component of all the carbonate formations. The unique chemistry may in part be a function of detrital contamination of the original depositional environment. Evidence also suggests that, locally, intrusive-related contact metasomatism of the high-grade metasedimentary rocks has altered the original composition. In the McMillan Flowage area, Chatterjee (1977) recognized Cu + Zn + Sn + W + Mo mineralization associated with skarn lithologies adjacent to a quartz monzonite intrusion. However, the majority of metallic showings that occur throughout calcareous members of Highlands assemblages consist of isolated base metal anomalies generated by disseminated chalcopyrite, sphalerite and galena in calc-silicate marbles. The high grade of regional metamorphism tends to complicate lithological and mineralogical relationships to the point that metallization patterns are difficult to interpret. For example, A.L. Sangster (personal communication, 1987) has suggested that the Meat Cove Zn deposit, hosted in marbles of a Grenvillian basement assemblage, is more closely related to the Balmat-Edwards-type of sedimentary exhalative Zn deposit than to a classic contact metasomatic (skarn) deposit-type.

CONCLUSIONS

Geological and geochemical data have shown that four unique marble-bearing metasedimentary assemblages dominate Late Proterozoic to very early Paleozoic carbonate-detrital shelf sequences of Cape Breton Island. Thick carbonate units interbedded with detrital metasedimentary rocks and mafic to intermediate volcanic rocks of low metamorphic grade characterize the Craignish Hills/North Mountain area. The sequence here most closely fits the lithologic description of the George River series summarized by Milligan (1970). Similarly, the narrow, low metamorphic grade, marble-bearing metasedimentary belt extending from Cape Dauphin to Kellys Cove is lithologically comparable to the Craignish Hills suite of George River rocks. Chemically, thick sections of high purity, predominantly calcitic marble characterize Craignish Hills/North Mountain carbonate formations. Metallic mineral potential in the area is restricted to isolated skarn occurrences containing variable amounts of Fe, Cu, Mo and W. The same is true of isolated sections of the Kellys Cove assemblage where high purity dolomitic marbles can be found. Calcitic marbles of the Kellys Cove area are, however, highly siliceous and display evidence of a contact metasomatic overprint. Only trace base metal mineralization has been observed in marble formations along the western faulted contact with the Kellys Mountain quartz monzonite. Lithologic and metallogenic, but not necessarily chemical, evidence has therefore led to the conclusion that metacarbonate rocks of the Kellys Cove area are more closely related to the George River series underlying Craignish Hills than to the higher grade assemblages of the Boisdale Hills.

The Boisdale Hills sequence is composed of massive, medium to high metamorphic grade, calc-silicate-bearing marbles interbedded with arenaceous metasedimentary rocks.

It is less obviously related to the George River series underlying Craignish Hills even though the type-section for the group is located here. The high alumino- and calc-silicate mineral content of the marbles appears to be a product of regional metamorphism of impure limestones and dolostones more than a product of regional metasomatism. Retrograde alteration leading to the development of a talc-serpentine-calcite mineral assemblage appears to be most widespread throughout areas pervasively cut by mafic dykes. The relationship between serpentinization and mafic dykes is unclear but it is most likely that hypabyssal emplacement promoted ground preparation for the later transportation of hydrothermal fluids. A number of massive and fracture-controlled magnetite showings are found in this environment. More commonly, metallic mineralization is restricted to disseminated base metal sulphide concentrations hosted in calcitic marbles. Although marble formations of the Boisdale assemblage contain higher than average background base metal values, only rarely has evidence been found to directly relate metallic mineralization to associated intrusive activity or to any structural mechanism controlling the transport of epigenetic, hydrothermal fluids. Elevated trace and major element content of the Boisdale marbles has also decreased their suitability as a source of industrial grade carbonate material.

Lastly, throughout the Cape Breton Highlands, isolated and discontinuous carbonate units < 100m in thickness, are interbedded with detrital metasedimentary sequences of variable stratigraphic association, metamorphic grade and age. Both lithologic and compositional criteria suggest that the carbonate-clastic metasedimentary sequences mapped throughout the Highlands cannot be considered equivalent to the George River series of southern Cape Breton Island. The complexity of lithologic relationships makes it difficult to interpret the present chemical signature of carbonate units. The composition may be related to original depositional or diagenetic environments, to later igneous related metasomatic activity or to combinations of the two. Elevated background Cu and Zn values, as with the Boisdale suite, are difficult to explain by observed metallogenic relationships. The only major carbonate-hosted, metallic showing examined in a Highlands assemblage was the Meat Cove Zn deposit of questionable contact metasomatic origin. It is concluded, based on element relationships, that only limited potential for skarn associated metallic mineralization exists within marble formations of the high grade Highlands assemblages.

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Technical discussions with numerous individuals on recent developments in Cape Breton geology were appreciated. They include: A.L. Sangster, Geological Survey of Canada; S.M. Barr and R.P. Raeside, Acadia University; D.A. Murray and A.K. Chatterjee, Nova Scotia Department of Mines and Energy; G.C. Milligan, Dartmouth, N.S.; and H.R. Oldale, College of Cape Breton, Sydney, N.S.

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Stratigraphy and geochemistry of the McGerrigle granite trains of Gaspésie, Quebec[1]

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David, P.P., Bédard, P., and Charbonneau, R., Stratigraphy and geochemistry of the McGerrigle granite trains of Gaspésie, Quebec; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 187-199, 1988.

Abstract

Excavations made across the granite trains of the McGerrigle Mountains reveal a number of glacial lithofacies interstratified with nonglacial sediments, which differ from one another in sedimentology, composition, and stratigraphy. The most common nonglacial deposit, locally derived colluvium, occurs either under glacial sediments over weathered bedrock, or between two glacial sediments, and is omnipresent at the surface of the region. The thickness of the nonconsolidated deposits decreases exponentially with distance from the McGerrigle Mountains. Concentration of elements in the clay fraction of the various sedimentary units may either increase or decrease with depth, or show no changes in vertical sections. In addition, the chemical compositions of the various sedimentary units are different from one another and can be related to differences in bedrock. While both the lithological and the geochemical composition of the individual glacial sediment facies vary, these attributes are useful in tracing their bedrock sources only if the exact nature and origin of a particular sediments facies are correctly identified.

Résumé

Les excavations, réalisées à travers les traînées de granite des monts McGerrigle, ont mis en évidence une séquence stratigraphique composée de lithofaciès glaciaires et non glaciaires. Parmi les unités non-glaciaires, des colluvions d'origine locale et omniprésentes à la surface de la région, ont également été relevées entre la base des dépôts glaciaires et la roche de fond altérée ainsi qu'entre deux dépôts glaciaires. L'épaisseur des dépôts non consolidés décroît exponentiellement avec la distance à partir des monts McGerrigle. Les analyses effectuées sur la fraction argileuse des différents sédiments ont établi que l'abondance des éléments dans les profils verticaux peut croître, décroître ou demeurer constante avec la profondeur. En outre, la composition chimique des différentes unités sédimentaires est hétérogène mais peut être reliée à la composition de la roche de fond. L'étude des provenances de matériaux exige l'identification exacte de la nature et de l'origine des faciès glaciaires en raison du fait que la composition lithologique et géochimique varie d'un faciès à l'autre à l'intérieur d'une séquence de dépôts glaciaires.

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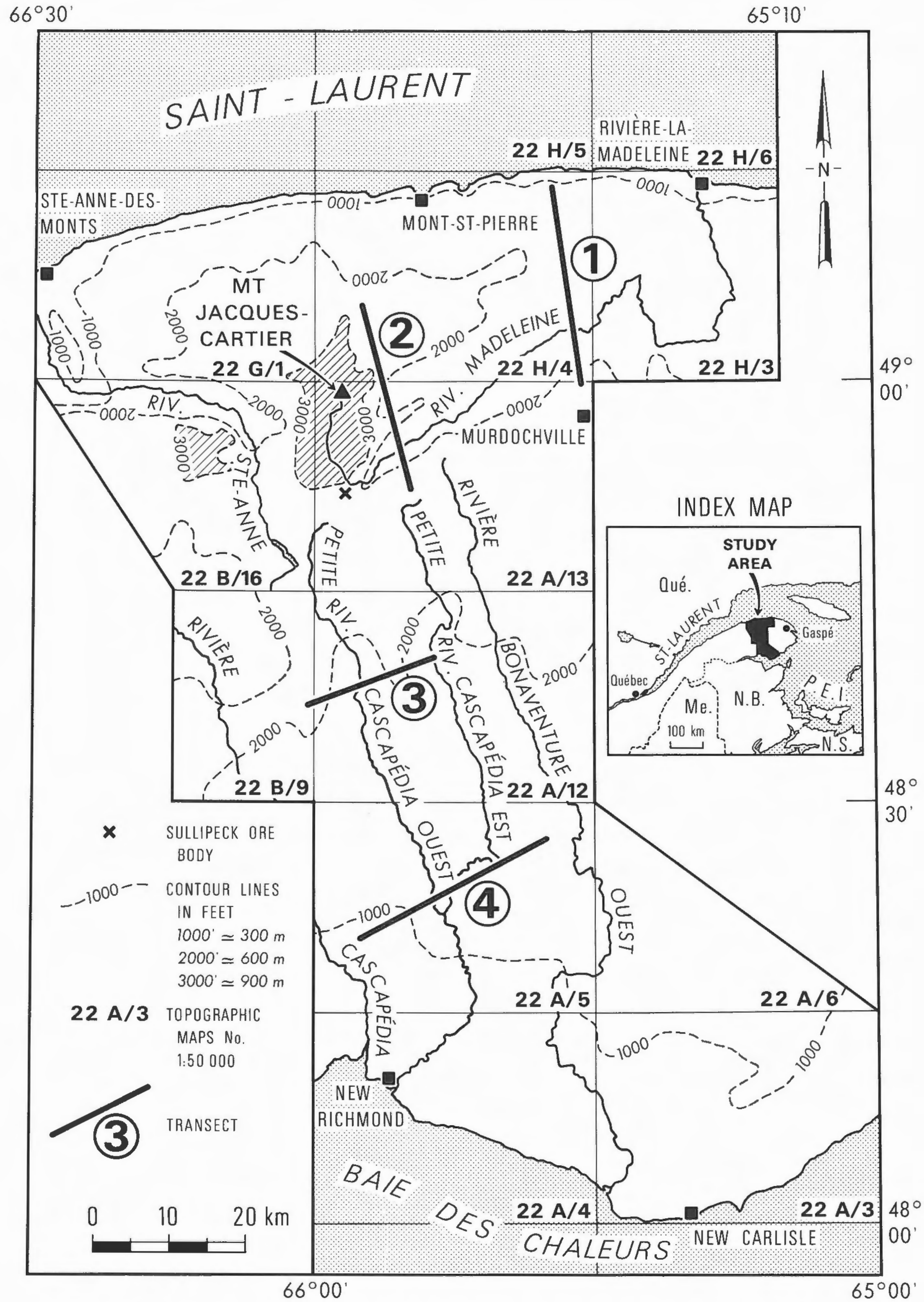

Figure 1. Location of study area. Contours approximately delimit the extent of plateau surfaces: hachured areas show land over 3000 feet (ca. 900 m) (the larger area with Mt. Jacques Cartier outlines the McGerrigle Mountains).

INTRODUCTION AND AIM OF STUDY

A project involving excavation and till sampling began in central Gaspésie in 1985 as part of the Canada Economic Plan for Gaspésie and the Lower St. Lawrence region. The principal objectives of this project are (1) to identify the type, extent and mode of origin of glacial and non glacial deposits, including the weathered bedrock, that underlie the granite dispersal trains of the McGerrigle Mountains in the central part of Gaspésie Peninsula; (2) to determine the nature of trace element distribution in unconsolidated formations overlying unweathered bedrock; (3) to determine the glacial influence on the geochemistry of the fine-grained fluvial sediments that have been analyzed for trace element concentrations (Choinière, 1982); and, finally, (4) to identify those element concentrations that are related to local mineralization and, consequently, to contribute to mineral exploration in Gaspésie. This report summarizes the preliminary results of geochemical analyses of samples collected during the first phase of field work in 1985 (David and Bédard, 1986) and the results of stratigraphic exploration done during the second phase in 1986.

LOCATION AND BEDROCK GEOLOGY

The study area forms a north-south corridor across Gaspésie Peninsula between St. Lawrence Estuary in the north and Baie des Chaleurs in the south; it is about 140 km long and from 40 to 80 km wide (Fig. 1).

Gaspésie forms part of the Appalachian Orogen and comprises a series of folded and faulted Paleozoic sediments intruded by Devonian granites (McGerrigle granite). The lithological units of the bedrock, which run more or less parallel with the elongation of the peninsula, trend perpendicularly across the study area (Fig. 2). Since the area traverses the whole peninsula, the lithological units within it include a large variety of sedimentary, metamorphic and igneous rocks and represent practically the whole lithological sequence of Gaspésie Peninsula.

GLACIAL GEOLOGY

Sedimentary facies

The study area includes two well defined glacial dispersal trains derived from the McGerrigle Mountains granite intrusion (Fig. 2, unit 7a, Devonian granite) (Chauvin, 1984; Chauvin and David, 1987; LaSalle, 1984). The larger train is oriented southeastward at about 160° and extends from the intrusive body to Baie des Chaleurs (Fig. 3a); the other one is oriented northeastward at about 55° to 60° and extends for about 40 km from the mountains toward the St. Lawrence Estuary. In addition to the trains, glacially dispersed granite blocks are found elsewhere around the mountains in all directions either in low concentration as part of a boulder blanket or as race occurrences of granite erratics (David and Bédard, 1986).

Within the limits of the 1985 field area, several distinct sedimentary facies have been identified in the excavation pits based on differences in sedimentary properties, structures, and lithologies (Table 35.1 in David and Bédard, 1986). They are, saprolite; a number of till facies which include deformation till, basal till of distinct local origin, basal till of more distal origin, basal melt-out till, ablation till, the train deposit, as well as a supraglacial avalanche deposit; glaciofluvial deposits; talus deposit; and colluvium. The stratigraphic position of the units corresponds to the order presented above from oldest to youngest, except for two units, the glaciofluvial deposits and colluvium. Glaciofluvial deposits may occur anywhere in the glacial portion of the sequence, while colluvium has been observed at three different stratigraphic positions. They are: (1) a preglacial monolithologic colluvium that has been recognized at a number of localities where it overlies saprolite and is overlain by glacial deposits. (2) A pre-last till colluvium that contains rare pebble or cobble-size erratic and overlies either a glacially eroded bedrock surface or a glacial diamicton and is overlain by younger glacial sediments. Finally, (3) there is the widespread post-glacial polylithologic colluvium that seems to dominate the surface in the whole region.

Ice flow directions

The principal glacial flow directions that have contributed to the dispersal of the granite blocks and to the deposition of glacial sediments have been summarized by David et al. (1984). Ice flow directions have been determined on the basis of the trends of striations and of dispersal trains throughout the region (David and Lebuis 1985, Chauvin, 1984; LaSalle and David, 1984). Glacial flow includes (1) early radial flow from the north-central highland regions, (2) a subsequent period of local easterly flow in the western part of the area, followed by (3) a regional southeasterly flow across the whole peninsula, (4) easterly and northeasterly flows affecting the northeastern part of Gaspésie and, finally, (5) a northerly flow in the northern part. David and LaSalle (1987) suggest late Sangamonian age for phase 1, early Wisconsinan for phases 2 and 3, mid Wisconsinan for phase 4 and late Wisconsinan for phase 5. Earlier interpretations related both phases 4 and 5 to the late glacial reversal of ice flow in the northern part of Gaspésie (Chauvin, 1984; David and Lebuis, 1985). The southeast-trending (160°) granite dispersal train developed during the early Wisconsinan regional southeasterly phase of ice flow that extended beyond the peninsula (LaSalle et al., 1985). Evidence for this direction of ice flow is recognized at more and more places on the central plateau areas (Cloutier and Corbeil 1986). The granite dispersal train oriented towards the northeast (55-60°) has originally been related to the short-lived late glacial reversal of ice flow to the northeast that occurred near the end of the Late Wisconsinan (Chauvin and David, 1987). It is presently thought to have developed during the more important and longer lasting phase of ice flow which affected a large part of the eastern plateau areas during the Mid Wisconsinan (David and LaSalle, 1987) and may have formed much of the sediments there (David et al., 1985).

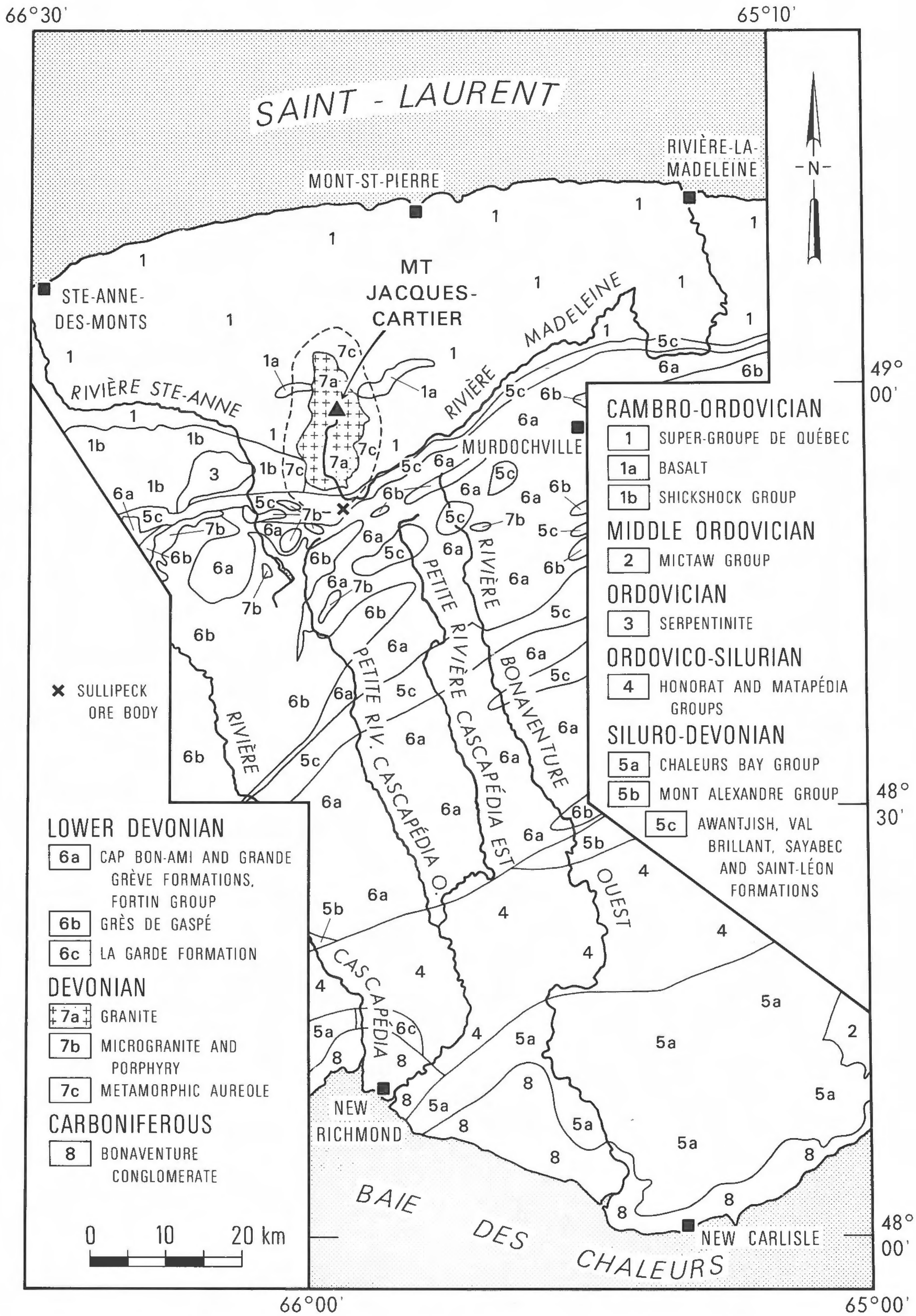

Figure 2. Bedrock geology of the study area; the McGerrigle Mountains are shown by units 7a and 7c (geology adapted from McGerrigle and Skidmore, 1967).

Figure 3. Photos of McGerrigle Mountains. (a) Above: View from top of Mont Jacques Cartier (highest peak) southeastward along direction of principal granite train (160°). Foreground: uneroded granite felsenmeer; background; glacially eroded surface of granite batholith. (b) Below: View of the mountains form the east. Rubble-covered area with thick glacial deposits in foreground.

METHODS

Field methods

As in the previous field season (David and Bédard, 1986), a hydraulic retro-excavator (back-hoe) was used to excavate 95 pits along four transects, two across each of the trains (Fig. 1). The northeasterly oriented train is crossed by two north-south trending transects one of which is located at the east side of McGerrigle Mountains and the other some 28 km to the east. The southeasterly oriented train is crossed by five east-west trending transects, three of which were excavated in 1985 near the south side of the mountains between the headwaters of Ste-Anne and Bonaventure rivers (see Fig. 35.3 in David and Bédard, 1986). Two additional transects excavated in 1986 are located about 26 and 61 km south of the mountains, respectively.

Individual pits along the 1986 transects are 1 to 6 m deep, most of them extending to bedrock. At places, the base of the glacial deposits was not reached owing to their great thickness. In addition to the 95 new excavation sites, a few of the pits that were excavated in 1985 been enlarged either for more detailed verification of their sedimentary record or for further sampling, or both. Samples were collected for geochemical analysis following the method described by David and Bédard (1986), for clay and heavy mineral analysis, grain-size and pebble lithological analysis, and age determination of the saprolites.

Laboratory methods

A total of 313 geochemical samples, collected in 1986, were prepared for analysis in the Quaternary Laboratory of the Université de Montréal, following the method of sample preparation and separation of clay-sized (<0.002 mm) material used by the Sedimentology Laboratory of Terrain Sciences Division. Theses samples have been analyzed for Cr, Mn, Fe, Co, Ni, Cu, Zn, Pb, As and U. The choice of these elements has been influenced, in large part, by the list of those elements for which the fluvial sediments from Gaspésie have been analyzed (Choinière, 1982). Grain-size analysis has been done on a number of samples using a combined hydrometer-sieve technique. The results will be used to characterize the various glacial and non glacial sediments.

STRATIGRAPHY

Sedimentary facies

Sedimentary facies described by David and Bédard (1986) are recognized in the newly excavated pits (Fig. 4, 5), with the following differences: (1) the thickness of the units vary more abruptly from one transect to another; (2) colluvium that underlies till at several locations is found to occupy an intertill position at one of the sites; (3) a previously reported basal till (David and Bédard, 1986) that was described as weathered and overlain by the youngest till, has been found to form part of an undertill colluvial sequence (the pit in which it was originally identified in 1985 was enlarged and deepened in 1986 for detailed examination).

The total thickness of sediments (excluding saprolite) fluctuates irregularly from excavation to excavation along each of the transect, except for the easternmost transect (transect 100 in Fig. 5) along which sediments are thick in the south and thin in the north. This is because that transect crosses over two different zones of glacial activity similar to those described from the west part of Gaspésie by David and Lebuis (1985). In fact, the southern part of the transect lies within a zone of intense glacial activity (Zone II of Chauvin, 1984) where thick glacial sediments were deposited, while its northern part lies within a zone of reduced glacial activity characterized by thin sediments. In the southern part, the thickness of sediments is in the order of 5 to 6 m and more (in 3 out of 5 pits bedrock was not reached), while in the north, it rarely exceeds one metre.

The thickness of sediments varies also with distance away form the base of the McGerrigle Mountains; it decreases exponentially both to the south and east. In order to demonstrate this relationship, the average value of thickness, calculated for each transect, has been plotted against the shortest distance measured from the base of the mountains to the transects (Fig. 6). The best fit curve, with a correlation coefficient of $R=0.93$, is obtained for the points only when the rate of decrease of average thickness is treated as an exponen-

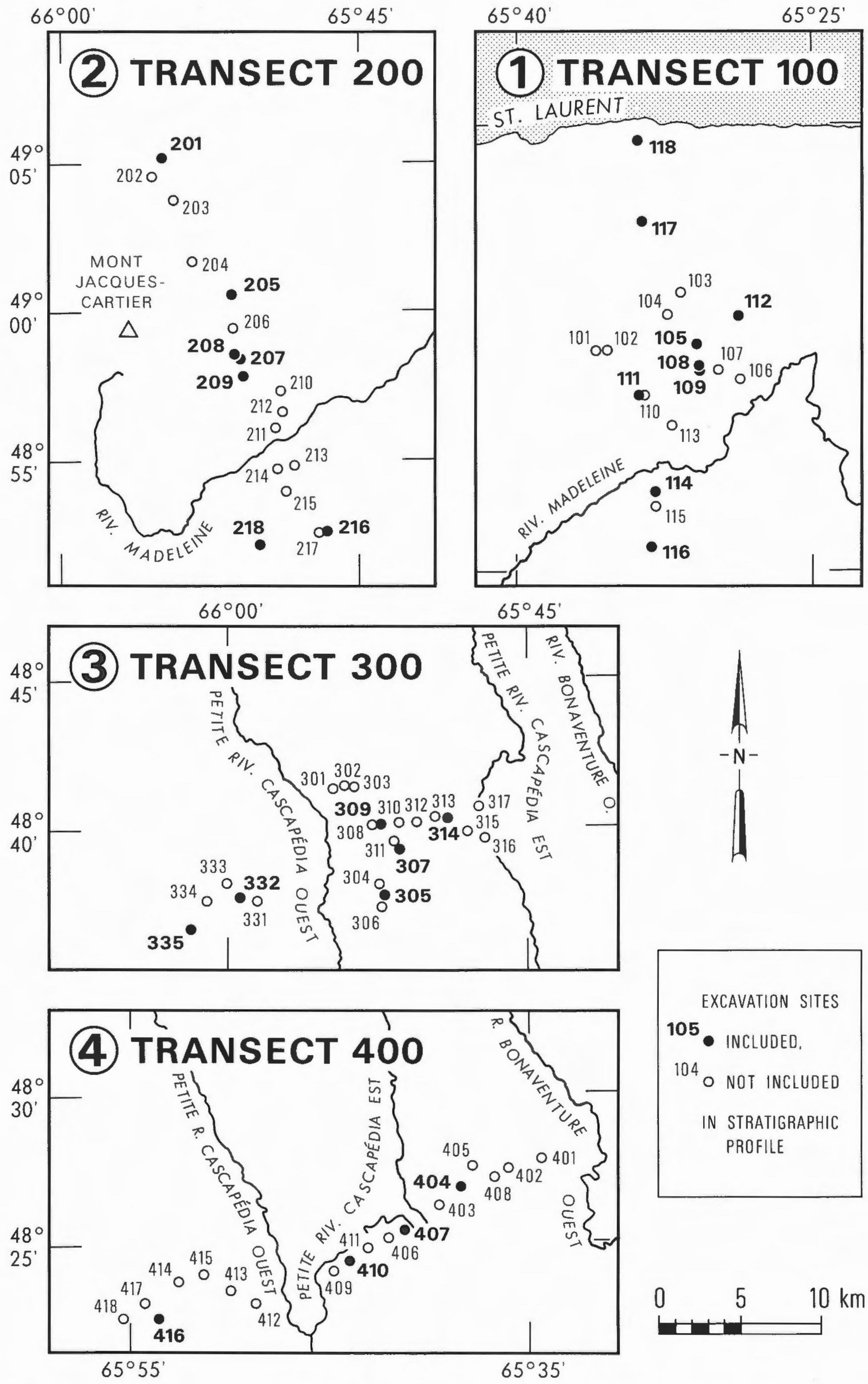

Figure 4. Location of excavation sites along the four transects. See Figure 1 for location of transects.

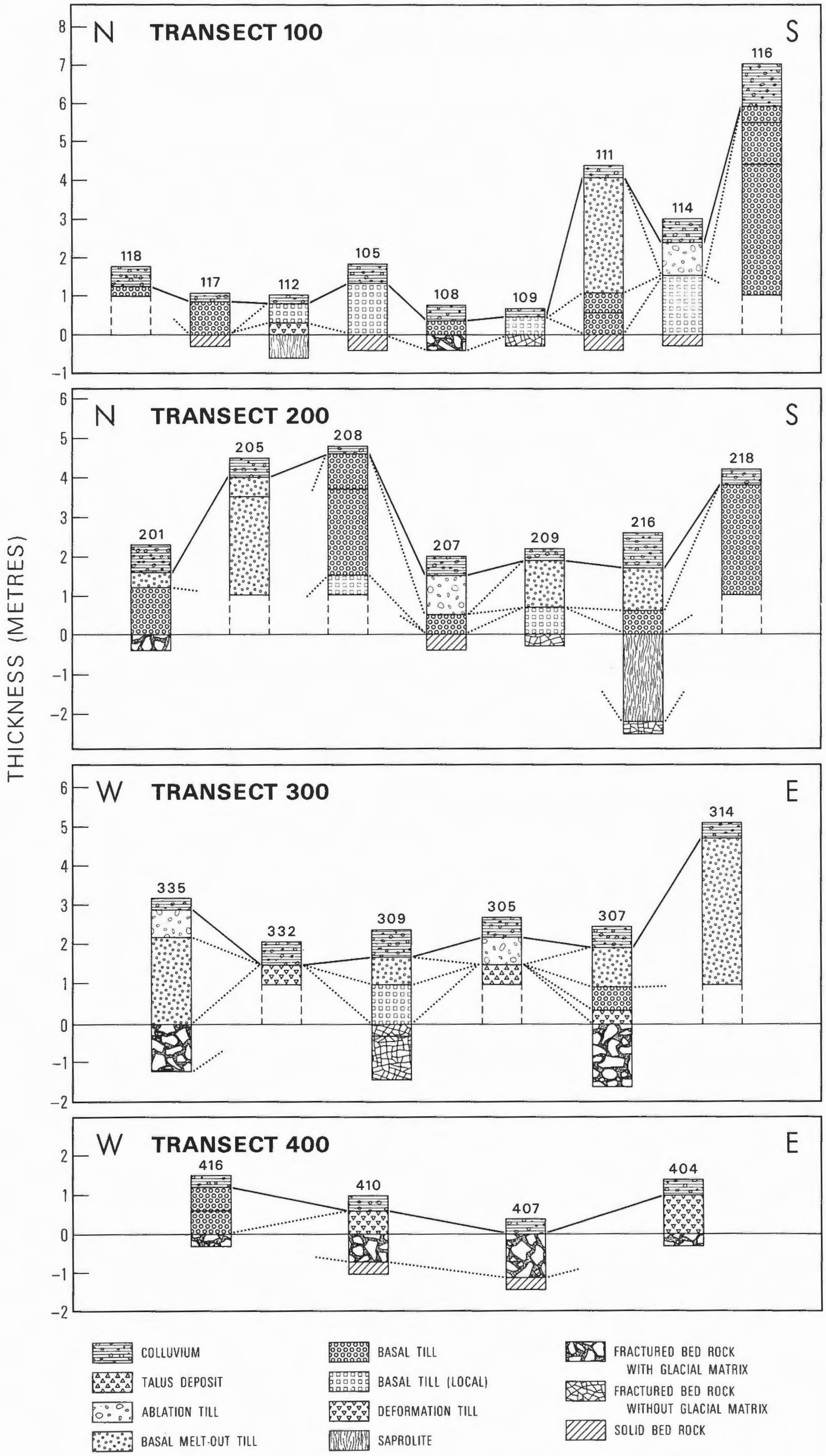

Figure 5. Stratigraphic profiles showing the correlation of sedimentary units in selected excavations along the four transects.

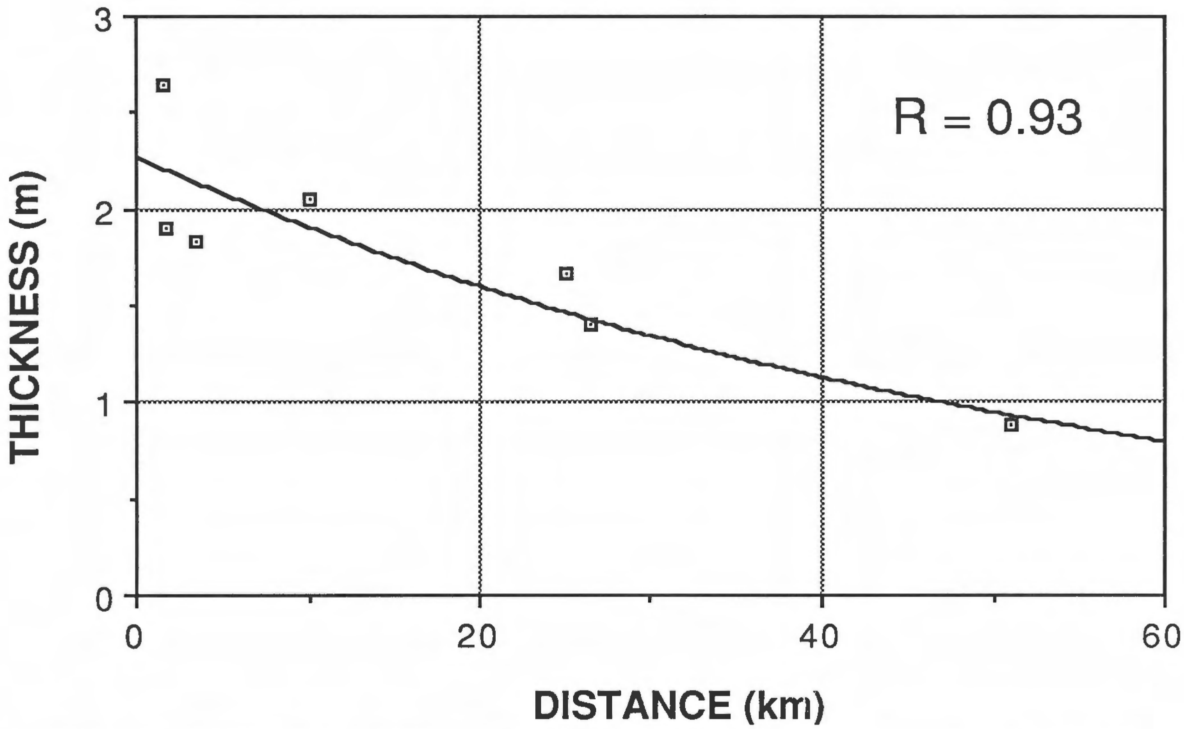

Figure 6. Diagram showing the exponential decrease of average thickness of deposits with distance from the McGerrigle Mountains. One average value has been calculated for each transect.

tial function of distance. Linear correlation gives a slightly lower correlation coefficient of R=0.88. The gradual decrease in the thickness of sediments away from the mountains could be the result of a decrease in the quantity of debris load in the glacier relfecting decreased activity at its base.

Regional changes in the thickness of glacial sediments form north to south across the peninsula have been observed both in the western half (David and Lebuis, 1985) and the north central part of Gaspésie Peninsula (Chauvin, 1984). In those areas the changes have been attributed to differences in glacial erosional activity produced by changes in the basal thermal regime of the glacier in the direction of ice flow. The same explanation may apply within the study area to both the southeasterly and northeasterly ice flow directions that produced the two granite trains.

Lithology

At any site: (1) the principal lithological component of the lowermost glacial sedimentary facies corresponds to the underlying bedrock (up to and over 90 %). The exact percent value however depends entirely on the type of the sedimentary facies; (2) there is a slight upward increase in the variety of lithologies present in the various facies within the same excavation; (3) there are abrupt lateral changes in the lithological composition of individual units from pit to pit and, more so, form transect to transect. These observations are best explained by the large variety of bedrock lithologies crossed by the ice and the relatively short distance of transport of the basal sediment facies as compared with those facies situated higher up in the stratigraphic coloumn. Since the lithological contacts trend perpendicularly to the principal, southeasterly ice flow direction, the glacier flowed across the largest number of possible bedrock types along its path.

Colluvium

The sediment facies that becomes increasingly prominent with distance from the mountains is colluvium. A large part of the study area has been mapped both in the north (Chauvin, 1984) and in the south (LaSalle, 1984) as having either colluvium or colluviated glacial sediments as the prominent deposit over bedrock. Along the southernmost transect and in the northern part of the easternmost transect, where unconsolidated sediments are the thinnest, colluvium, dominated by fractured rock fragments, is the dominant surficial deposit and may overlie either a thin glacial diamicton or occur directly over bedrock. Because of the areal importance of this type of colluvium, it is discussed in detail and a mode of origin is suggested.

The widespread occurrence of a surficial diamicton dominated by fractured rock fragments, has been noted in a large part of Gaspésie. This sediment has variably been called colluvium (David and Lebuis, 1985; Chauvin, 1984) rubble (LaSalle, 1984; LaSalle and David, 1984) and recently, "regolith" ["régolithe" in French] (Cloutier and Corbeil, 1986). We strongly recommend that the use of this last term be discontinued before it becomes widespread in the literature because it leads to confusion. The term "regolith" is an unfortunate choice of expression to describe the surficial blanket of fragmented rocks since, by definition, regolith includes "the entire layer or mantle of fragmentary and loose, incoherent, or unconsolidated rock material, of whatever origin [...] that nearly everywhere [...] covers the more coherent bed rock [SIC]" (Gary et al., 1972, p. 598). Regolith in that sense comprises all the unconsolidated material of diverse origins, including the rubble itself, the saprolite and the glacial deposits, that overlie the coherent, solid bedrock.

During mapping of the surficial sediments, a project with which the senior author has been involved, colluvium was always recognized in gravel pits and road cuts or shallow had-dug pits as an omnipresent deposit of variable thickness. Nonetheless, the fact that relatively thick accumulations of it may mask the underlying glacial deposits on a large scale became evident only through the excavation work carried out within this project. It was noted that in those areas where shallow pots, 40 to 50 cm deep, made during mapping, revealed only rubble or colluvium as the only surficial deposit, excavation exposed, notably along the southern half of the transect located on the east side of McGerrigle Mountains (Transect 200 in Fig. 5), thick glacial sediments the base of which could not be even reached. In several of the exposures in that area, colluvium is form 60 to 90 cm thick (for example, section #216 in Fig. 5). This explains why colluvium has been so widely mapped in areas where glacial sediments are now known to be quite thick (Fig. 3b). Nervertheless, in a large central part of the study area, colluvium is thinner and it overlies either a thin, 1 m thick layer of glacial deposit, or a glacially eroded bedrock surface, or the fractured bedrock (Fig. 7).

The surficial rubble forming the colluvial material varies in appearance from place to place. While generally it may appear to be uniform, locally it shows small patches or concentrations of a particular lithology. Its thickness may also vary within short distances or be uniform over a large area. The deposit occurs equally on gently or moderately sloping surfaces or on perfectly horizontal areas.

Origin

Several hypotheses have been put forward to explain the origin of the surficial rubble. Most of the hypotheses call upon frost shattering of the bedrock. While we do not dismiss the importance of this process, we suggest that some of the rock types ivolved in rubble formation do produce small angular fragments through a simple process of dehydration. Probably, forst shattering and drying are the two most important processes of rock disintegration that produce rubble.

Source material

It is proposed here that rubble is derived partly from the local bedrock and partly from the disintegration of glacially transported and deposited rock material that is either dominated by or composed entirely of the local bedrock lithologies. A glacial source of material comprising the rubble has so far been ignored. The lithologies that are most apt to disintegrate are indurated shale or silstone, fine-grained sandstone, and some of the porphyric rock types. We have noted earlier that the lithological composition of the glacial deposits strongly reflects the underlying bedrock and, therefore, could produce a surface rubble compositionally similar to the underlying parent material. For example, the so-called monolithologic till, a deformation till, described form the western half of Gaspésie (David, 1982; David and Lebuis, 1985), which has been formed by the process of local, minor mobilization of the bedrock by the glacier, would be an ideal glacial source for the formation of a monolithologic rubble. In those areas where more than one variety of easily fragmented rock type was present at the ground surface, when they shattered, they left behind patches of rock fragments lithologically distinct the underlying rubble.

Evidently, in areas where the bedrock outcrops, it may have provided material for the rubble. But in areas where bedrock is masked by either thin or thick glacial deposits and where the ground surface is almost horizontal, the glacially transported sediments, chiefly those that had little matrix, may also have contributed large quantities of rock fragments to the rubble. In those areas where matrix-rich glacial

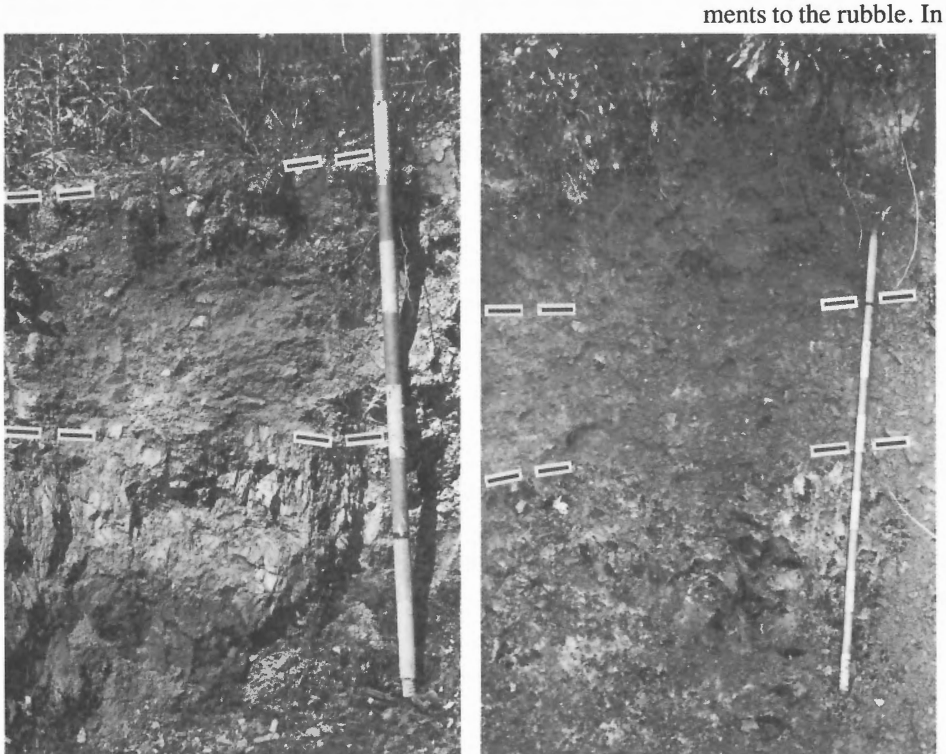

Figure 7. Photographs of sediment facies. Left: Section GP-86-210, showing polylithological colluvium (with granite) over fractured schist. Right: Section GP-86-303, showing polylithological colluvium (without granite) over glacial diamicton, overlying more or less weathered volcanic rock.

deposits occur, a simple removal of the matrix form the uppermost parts of the sediments by slope wash, could expose the easily fractured lithologies to the elements and would provoke the production of rubble. More importantly in Gaspésie, glacial blocks could have been continuously moved through the matrix to the surface of the deposits by frost heave (Washburn, 1973, p. 66-80) providing ample quantities of material for the formation of the rubble. Through these processes of slope wash, frost heave, frost shattering and desiccation, the easily fragmented rocks could have formed part of the rubble, while the other, more resistant, glacially transported rock types would have become concentrated at the surface. Once formed, a thick layer of rubble would have protected underlying glacial deposits from further disintegration.

The above model of rubble formation would explain the presence of rubble over glacially striated bedrock surfaces in areas of low bedrock relief. There, the rubble could not have formed from the disintegration of the solid bedrock since its glacially eroded surface is still intact. It could have only formed, therefore, from the glacially displaced blocks. David et al. (1985) reported such rubble-covered glacially striated surfaces from the north central plateau regions.

GEOCHEMICAL DATA

Precision of results:

In 1985, 241 samples were submitted for geochemical analysis for the ten elements indicated in Table 1. In addition, repeat test samples were also submitted by splitting one of very 25 samples into two equal sub-samples which were then independently numbered. The results of the doubly analyzed samples were compared so that we would be able to quantify the precision of the results. The range of values that were obtained would eventually be used to separate possible analytical errors from differences in geochemical composition of the various sedimentary facies. Furthermore, we also used, but separately, the analyzing laboratory's routine duplicate analyses. Table 1 shows the per cent deviation of the two series of test samples accompanied by statistical parameters. While most elements show relatively good comparable results, some do not. Among them, the worst results were found to be associated with As in those samples that were analyzed by a hybrid generator method. Analysis for As by a colourimetric method gave satisfactory results.

Table 1. Comparison of doubly analyzed samples.

Vertical distribution of elements

Within pits, the vertical distribution of the various elements was examined. We report only the results of analyses in those esposures where at least five sediment samples have been taken for geochemical analysis (Fig. 8, 9, 10). The vertical variations in the concentration of the elements differ from pit to pit. In some of the excavations (1) the concentration of almost all the elements increases vertically (Fig. 8), while in others (2) it decreases (Fig. 9). There are also pits in which (3) the concentration of only some of the elements decreases while, in comparison, that of the others varies irregularly with depth. Finally, (4) in a few sections, some of the elements show rather aberrant values, such as Zn and Pb in Figure 10. These aberrant values require further explanation. Nevertheless, it has not been statistically established which of the elements show the greatest similarities or dissimilarities with the others or which of them deviates most. As a general rule, however, U has been found to be one element that shows the least conformity with the others.

Table 1. Comparison of doubly analyzed samples.

LAB. DUPLICATES	Cr#	Mn#	Fe†	Co#	Ni#	Cu#	Zn#	Pb#	As clr#	As hyb#	U#
% minimum	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%		0.0%
% maximum	10.0%	12.5%	6.5%	20.0%	22.2%	7.4%	7.4%	26.7%	20.4%		34.3%
Mean (%)	3.7%	2.3%	2.0%	6.1%	6.3%	3.3%	2.5%	8.0%	8.1%		9.0%
Number of pairs	(23)	(23)	(23)	(23)	(23)	(23)	(23)	(23)	(17)	(0)	(14)
UdeM DUPLICATES											
% minimum	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	7.1%	0.0%
% maximum	6.7%	11.1%	8.4%	15.4%	24.2%	6.1%	7.3%	31.1%	34.5%	93.1%	27.9%
Mean (%)	2.9%	3.7%	3.7%	8.3%	7.4%	2.8%	2.2%	10.1%	14.6%	37.7%	6.5%
Number of pairs	(11)	(11)	(11)	(11)	(11)	(11)	(11)	(11)	(10)	(11)	(11)
COMBINED VALUES											
% minimum	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	0.0%	7.1%	0.0%
% maximum	10.0%	12.5%	8.4%	20.0%	24.2%	7.4%	7.4%	31.1%	34.5%	93.1%	34.3%
Weighted mean (%)	3.4%	2.8%	2.6%	6.8%	6.6%	3.2%	2.4%	8.7%	10.5%	37.7%	7.9%
Number of pairs	(34)	(34)	(34)	(34)	(34)	(34)	(34)	(34)	(27)	(11)	(25)

Per cent differences for values in ppm
† Per cent differences for values in %

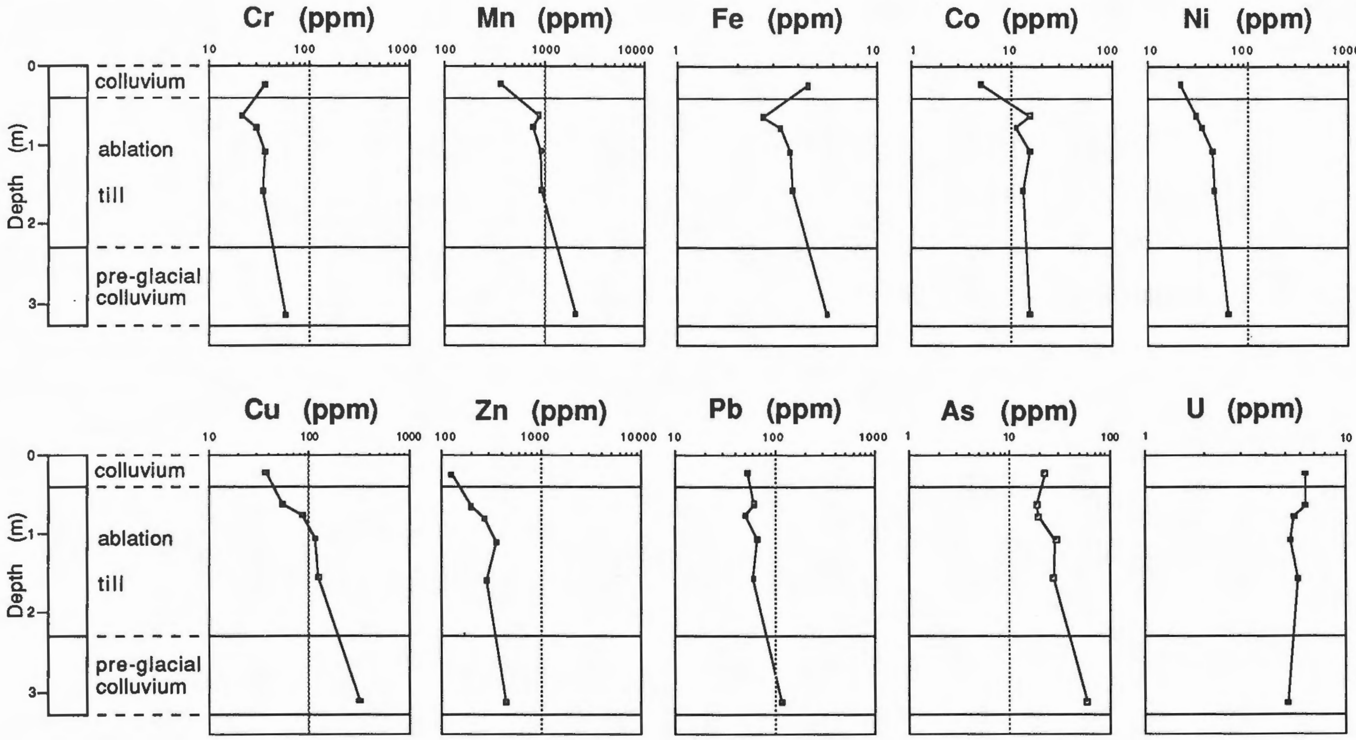

Figure 8. Trace element abundance profile and sediment facies at section GP-619; general increase of trace element concentration with depth.

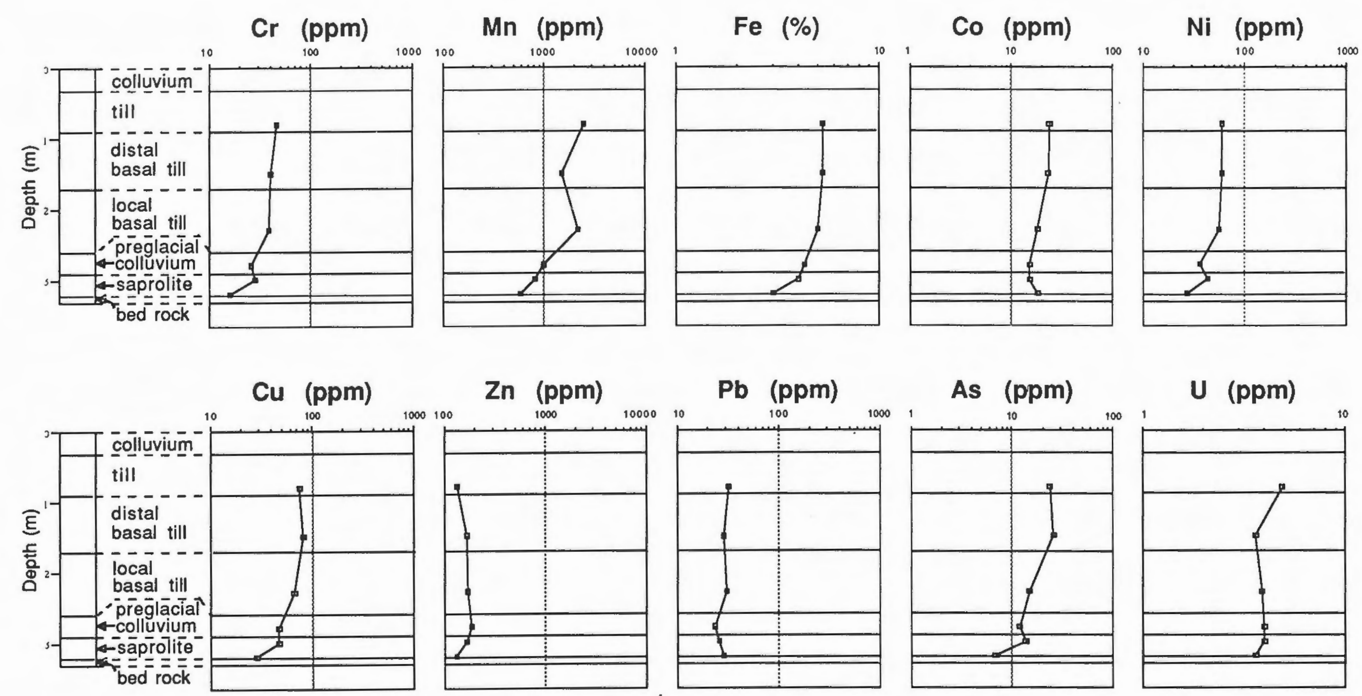

Figure 9. Trace element abundance profile and sediment facies at section GP-723: general decrease of trace element concentration with depth.

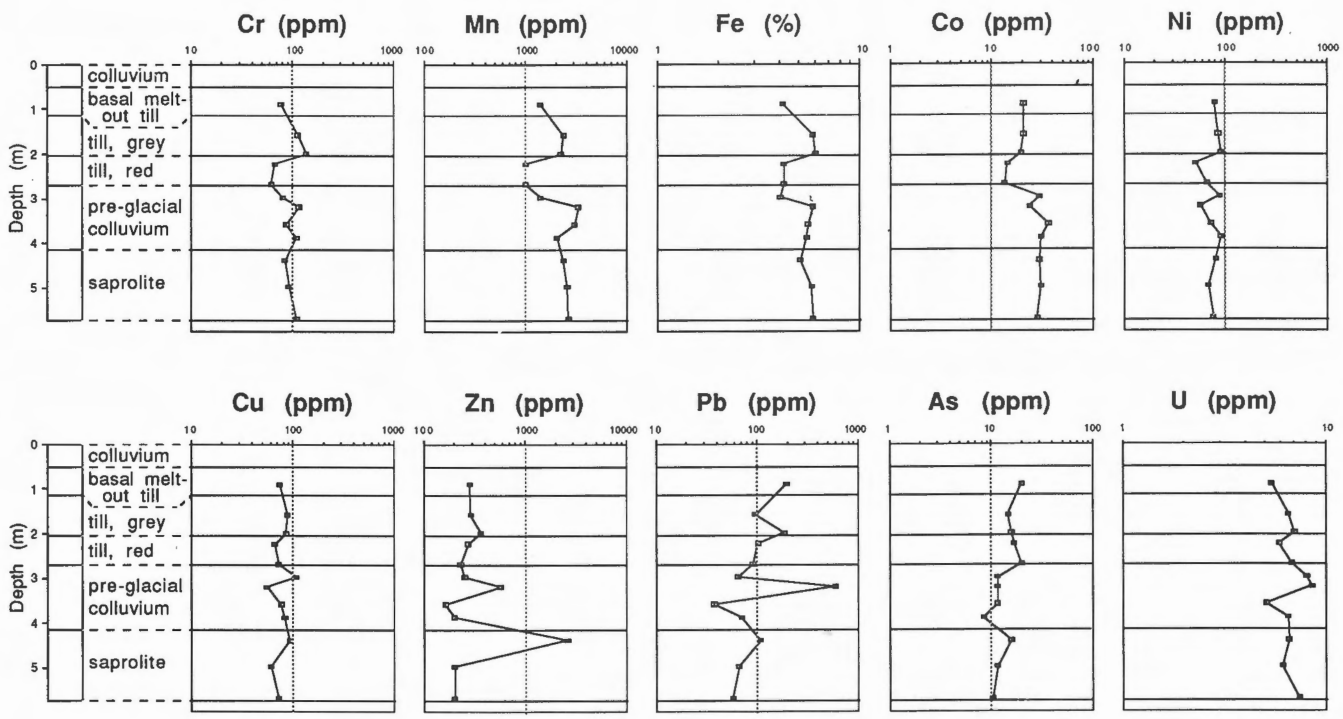

Figure 10. Trace element abundance profile and sediment facies at section GP-517: varied trace element concentration with depth.

The aberrant values associated with Zn and Pb in Figure 10 may be explained either as problable analysis errors, or as chance inclusions of dustsized mineral particles, such as galena or spharelite in the analyzed sediments fractions. It should be noted that in the case of the high Zn value, some of the other elements, such as Cu, Pb and As, also show a slight peak in the same direction as Zn. In the case of Pb, Zn shows a peak while Cr, Mn and Fe show an abrupt increase at the same level. It is noted that, at this time, the source of these high values is not certain.

A good example to see how certain elements behave at depth is Ni. Ni has been found to decrease with depth in weathered glacial deposits (Rencz and Shilts, 1980). In our area, at some of the sites, Ni decreases with depth, similar to some of the other elements, including Fe and Mn, (Fig. 9). In contrast, at other sites, it increases with depth the same way as Fe and Mn. The explanation for such opposing changes cannot be attributed to differences in the degree of weathering of the glacial materials, as weathering seems to have affected in much the same way all the glacial sediments which are all relatively thin and permeable. The explanation, therefore, may lie in the differences in the quantity of element concentration that was initially present in the various sediment facies. In order to demonstrate the relationship of sediment facies and element concentration, the various facies have been identified in Figures 8, 9 and 10. It appears evident that changes in the vertical distribution of elements occur between lithofacies. Since the composition of the sediment facies is primarily source dependent, the variations in element concentration appear to reflect compositional differences.

CONCLUSIONS

A number of preglacial, glacial, intertill non-glacial, and postglacial stratigraphic units have been identified from excavations made across the southeasterly and northeasterly oriented trains of the McGerrigle Mountains, in central Gaspésie. The sedimentary units differ from one another in their sedimentary, compositional, and stratigraphic attributes within one excavation, and the same unit occurring in adjacent pits may greatly differ from one another in composition. Colluvium is present at three different stratigraphic positions, namely, it overlies weathered bedrock under glacial sediments; it occurs between two glacial sediments; and, finally, it is omnipresent at the surface of the whole region masking totally the underlying glacial deposits. The results of chemical analyses for ten elements in the clay fraction of the various sedimentary units show that the concentration of elements may either increase or decrease with depth, or show no changes in vertical sections. In addition, the chemical compositions of the various sedimentary units are different from one another and can be related to differences in their bedrock source. The most important conclusion that may be drawn is as follows; while both the lithological and the geochemical composition of the individual sediment facies vary from unit to unit, these attributes are useful in tracing the bedrock source only if the exact nature and origin of the particular facies is correctly idientified.

Future research involves the completion of the evaluation of the distribution of the ten elements in the different sedimentary units; the determination of the regional variations in the concentration of the elements; the detection of any problable showing or anomaly within one transect and across adjacent transects.

ACKNOWLEDGMENTS

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Geological setting of granites and related tin deposits in the North Zone, Mount Pleasant, New Brunswick

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Abstract

Granitic intrusions and associated mineral deposits at Mount Pleasant were emplaced in a subvocanic environment near the southwestern margin of the Mount Pleasant caldera. Mineral deposits in the North Zone are associated with three phases of granite. Tungsten-molybdenum deposits apear to be related to the oldest phase (Granite I). Most tin deposits are within, or adjacent to, an intermediate phase (Granite II). The youngest phase (Granite III) has only minor associated tin zones. Textural features of Granite II, such as miarolitic cavities and comb quartz layers, are characteristic of fluid saturation in the parent magma. Other features such as micrographic intergrowths of quartz and K-feldspar reflect undercooled conditions that may have been related to rapid release of a fluid phase (i.e. pressure quenching). The association of the tin deposits with granite containing these features indicates that a significant portion of the mineralizing fluids were magmatic.

Résumé

Les intrusions granitiques et les gisements de minéraux associés que l'on trouve sur les lieux du mont Pleasant ont été mis en place dans un milieu subvolcanique près de la marge sud-ouest de la caldeira de Pleasant Mount. Les gisements de minéraux de la zone nord sont associés à trois phases de granite. Des gisements de tungstène et molybdène semblent être apparentés à la phase la plus ancienne (Granite I). La plupart des gisements d'étain se trouvent à l'intérieur ou contigus à une phase intermédiaire (Granite II). La phase la plus récente (Granite III) ne possède que de petites zones d'étain connexes. Les éléments texturaux de Granite II, comme les cavités miarolithiques et les couches de quartz à épontes fortement minéralisées et presque jointives, sont caractéristiques d'une phase de saturation fluide au sein du magma originel. D'autres particularités comme les intercroissances micrographiques de quartz et de feldspath potassique reflètent des conditions de sous-refroidissement qui peuvent avoir été liées à la libération rapide d'une phase fluide (c'est-à-dire la trempe sous pression). L'association des gisements d'étain au granite renfermant ces éléments révèle qu'une partie appréciable des fluides minéralisateurs sont d'origine magmatique.

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INTRODUCTION

Recent exploration and underground development of tin deposits in Billiton Metals Canada Inc.'s North Zone at Mount Pleasant has been carried out by Lac Minerals Ltd. ;(who are earning a 50% interest in the project). This work has outlined at least three distinct phases of granite designated Granite I, Granite II and Granite III. Some of these phases may be correlative with phases in the Fire Tower Zone previously defined by Kooiman et al. (1986). The nature of the various granitic phases and their relationship to tungsten-molybdenum and tin deposits is the subject of continuing research by the first author. This paper summarizes the geological setting of the Mount Pleasant intrusions and associated tin deposits in the North Zone, incorporating the results of recent exploration and re-examination of core drilled in the 1960s and 1970s by previous operators.

GENERAL GEOLOGY

Granitic intrusions and associated deposits at Mount Pleasant were emplaced in Late Devonian or Early Carboniferous time along the southwestern margin of the Mount Pleasant caldera (Fig. 1). They have intruded intracaldera volcanic and sedimentary rocks of the Piskahegan Group, recently dated as Late Devonian (McCutcheon, 1986). At Mount Pleasant, the Piskahegan Group consists of at least three felsic ash-flow tuff units, a felsic to intermediate flow and two conformable interlayers of sedimentary breccia. In the Mount Pleasant area, the Piskahegan Group rocks generally dip about 10 to 15° to the northwest except along the margin of the caldera where they dip at 15 to 50° toward the centre of the caldera. They are bounded to the west by slate and greywacke of the Silurian Waweig Formation which form the southwest boundary of the Mount Pleasant caldera. At least

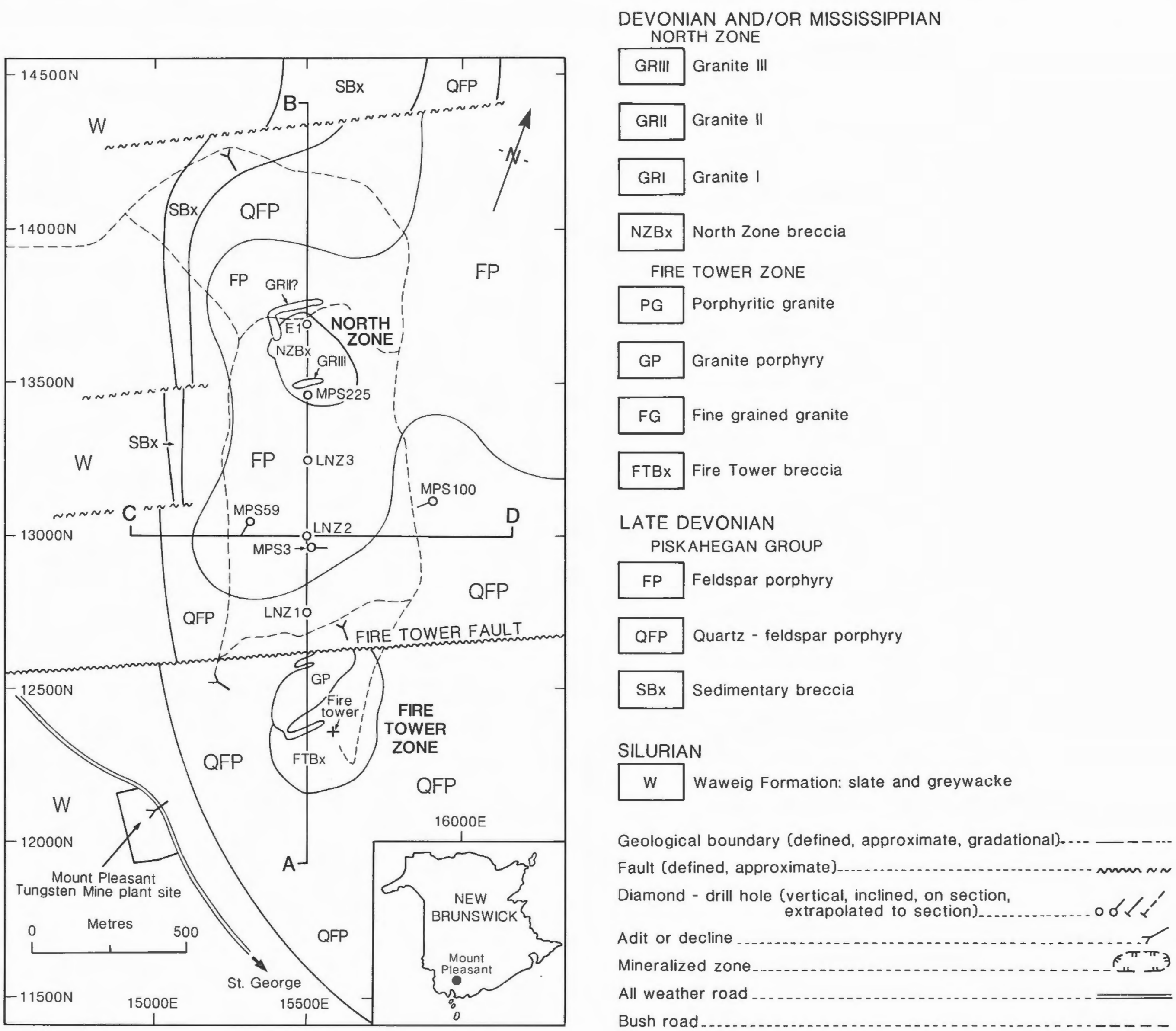

Figure 1. Surface geological map of the Mount Pleasant area (modified from Ruitenberg, 1967 and McCutcheon, 1983).

two sets of faults are present. One set trends 070° and includes the Fire Tower fault (Fig. 1). Another set trending approximately northwest is prominent in the North Zone but is not well documented.

Quartz-feldspar porphyry

The three ash-flow units now recognized at Mount Pleasant have been referred to collectively in the past as "quartz-feldspar porphyry" (e.g. Ruitenberg, 1967; Parrish and Tully, 1978; Kooiman et al., 1986). The three units are all similar in appearance and consist of rhyolitic crystal tuff characterized by medium-grained phenocrysts of quartz (15 to 25 % by volume) and pink and white feldspar (10 to 30 %) in a light-grey, microcrystalline matrix. The central parts of the flows are typically massive and structureless; eutaxitic foliation is present in places near the upper and loweer contacts of the flows. Fragments of slate or argillite up to several centimetres across occur locally in the flows near their lower contacts. Where no eutaxitic foliation is present, quartz-feldspar porphyry resembles an intrusive rock and intrusive phases may be present, particularly in the Fire Tower Zone (Kooiman et al., 1986). The quartz-feldspar porphyry exposed at surface is equivalent to McCutcheon's (1983) Little Mount Pleasant Tuff unit which occurs to the east of Mount Pleasant.

Feldspar porphyry

"Feldspar porphyry" is a rhyodacitic unit with abundant mdium- to coarse-grained pink feldspar phenocrysts (20 to 30 % by volume) and minor fine- to medium-grained quartz phenocrysts (less than 5 %) in a brown to grey microcrystalline matrix. The phenocrysts are subhedral to euhedral and unbroken; they do noto exhibit fragmentation characteristic of a tuffaceous origin. Feldspar porphyry is also relatively massive with no obvious primary layering. It appears to overlie quartz-feldspar porphyry conformably but the contact in the vicinity of the North Zone has been obscured by alteration and appears gradational. This unit is thus extrusive and is probably a lava flow rather than a crystal tuff. It is equivalent to the intrusive Feldspar Porphyry unit of McCutcheon (1983).

Sedimentary breccia

The two sedimentary breccia units consist mainly of angular to subrounded, clast-supported fragments of slate or argillite in a fine-grained matrix of similar material. Quartz fragments make up less than 10 % of the breccia and fragments of quartz-feldspar porphyry are rare. All the fragments are unsorted and typically range from 1 to 5 cm across. The sedimentary breccias are relatively massive with little internal structure; thin beds of fine-grained material that show sedimentary layering are rare. However, textural features are generally indistinct because the sedimentary breccias have been metamorphosed to biotite hornfels by the underlying granitic intrusions.

The distribution of sedimentary breccias at Mount Pleasant is shown in Figures 2 and 3. They appear to be conformable layers that separate the three ash flow units of quartz-feldspar porphyry. The uppermost breccia unit is at least 50 m thick along the margin of the caldera and pinches out to the east (Fig. 3). It has been traced for a short distance south of the Fire Tower fault where it either pinches out or is truncated by granitic dykes or croscutting breccias (Fig. 2). The lowermost unit is as much as 25 m thick near the caldera margin and pinches out to the south. It likely pinches out to the east also but its distribution in this direction is not well known.

The sedimentary breccia is lithologically similar to much of the sedimentary breccia or "sharpstone conglomerate" in McCutcheon's (1983) Scoullar Mountain Formation, which occurs at various localities around the margin of the Mount Pleasant caldera. The sharpstone conglomerate consists

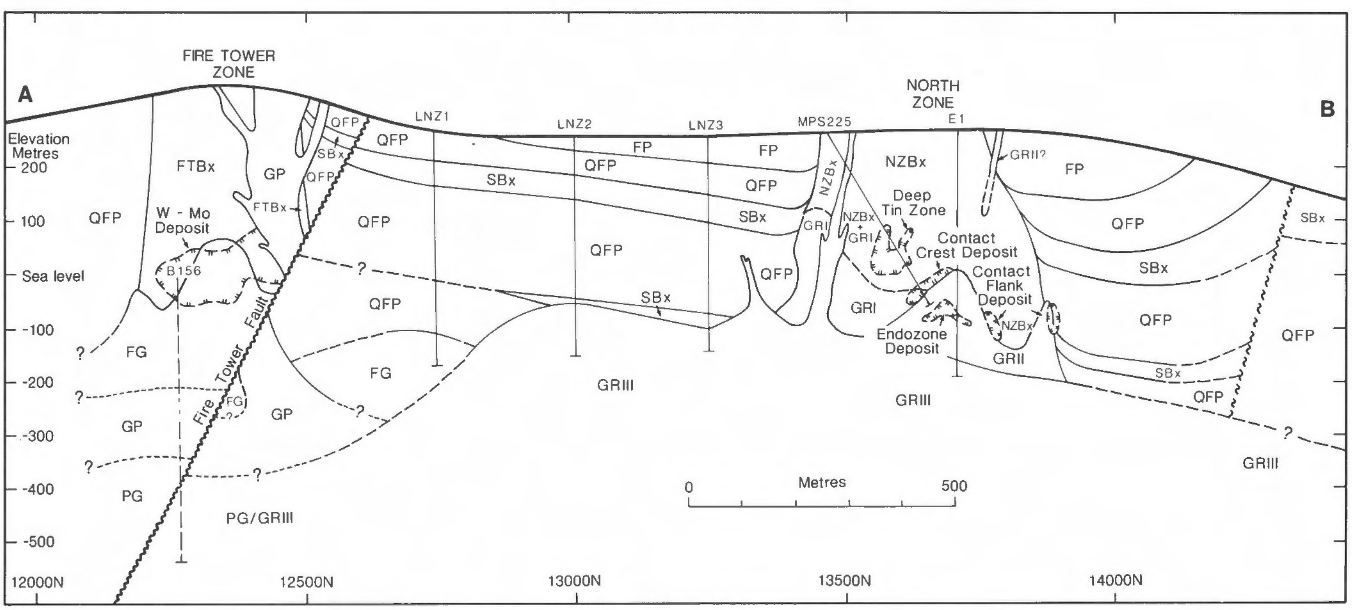

Figure 2. Cross-section A-B (Fig. 1) through the Mount Pleasant area (see Figure 1 for legend).

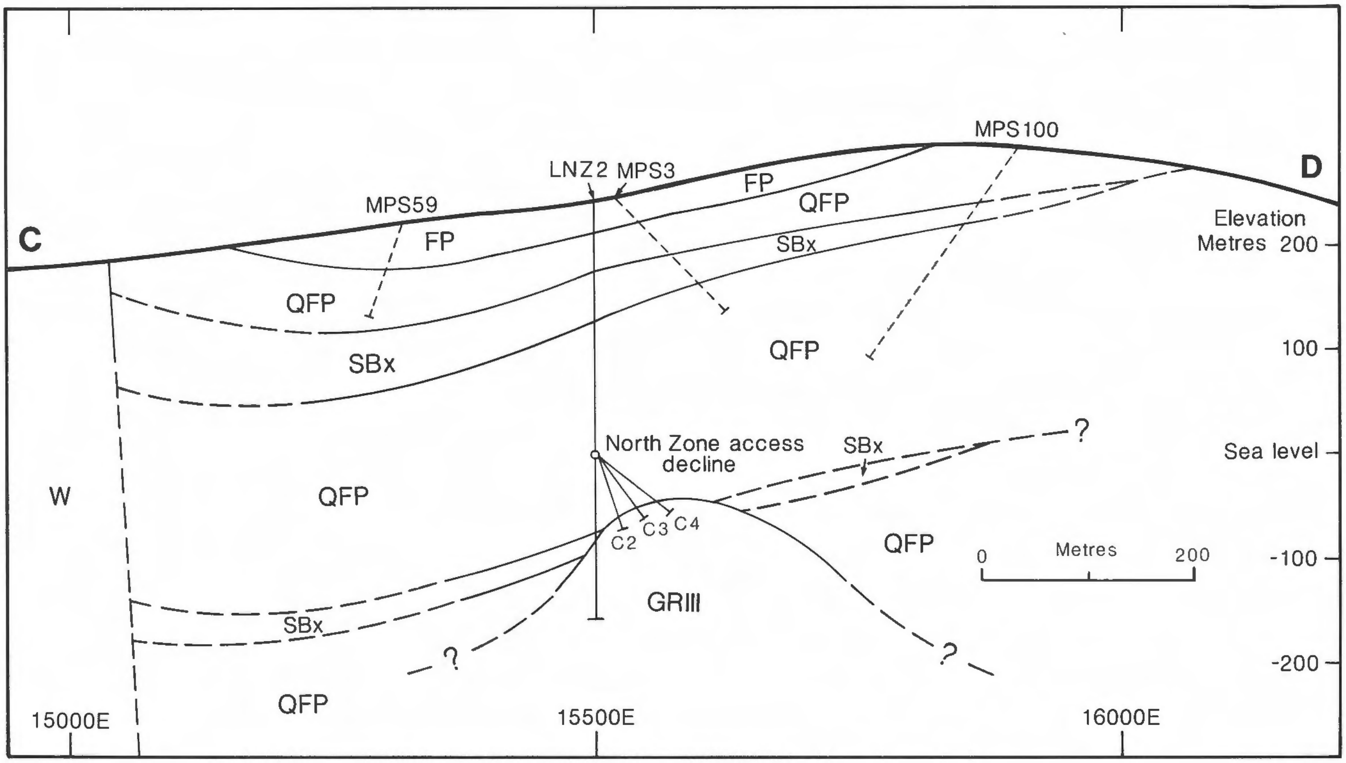

Figure 3. Cross-section C-D (Fig. 1) through the Mount Pleasant area (see Figure 1 for legend).

predominantly of angular, unsorted, clast-supported fragments of metasedimentary rock in a comminuted matrix of the same material. Fragments of felsic crystal tuff and, in places, granite are lss abundant. Internal layering with a water-worked matrix is rare. The conglomerate is interbedded with mafic to intermediate lavas and pyroclastic rocks, and felsic tuffs. Which of the Scoullar Mountain sedimentary breccia units correlate with those at Mount Pleasant is uncertain. However, all the sedimentary beccia units represent talus and/or other types of colluvial deposits derived mainly from nearby metasedimentary and granitic rocks that formed the caldera walls during the initial infilling of the caldera. The volcanic fragments probably were contributed by a precaldera eruptive unit which has been eroded (McCutcheon, 1985).

North Zone breccia

The ash-flow tuffs and interbedded sedimentary breccias have been highly brecciated and intruded by granitic rocks in two areas designated the Fire Tower Zone and the North Zone (Fig. 1). In both these areas, crosscutting magmatic-hydrothermal breccias and intrusive rocks form irregular, pipelike complexes that were centres of subvolcanic intrusive and hydrothermal activity. Similar to the Fire Tower breccia as described by Kooiman et al. (1986), the North Zone breccia is composed of multiple phases which range from matrix-supported breccias with rounded fragments to clast-supported crackle breccias with predominantly angular fragments. Both fragments and matrix material in the breccias have been altered extensively. Siliceous alteration consisting of quartz and topaz is most prevalent; however chlorite and/or biotite alteration is also present. Because of the alteration, textures and lithologies of the fragments are difficult to recognize. Relationships between the breccia and associated granitic rocks have also been obscured by alteration in many places.

The North Zone breccia probably was formed by repeated episodes of hydraulic fracturing related to the release of volatiles during crystallization of Granite I. Forceful emplacement of the breccia is indicated by the deformation of older sedimentary breccia units which have steep to vertical dips adjacent to their contacts with North Zone breccia.

GRANITIC INTRUSIONS

Based on field observations, granitic rocks in the North Zone appear to have formed three distinct intrusions which, from oldest to youngest, are designated Granite I, Granite II and Granite III. The three granites, where relatively unaltered, are similar mineralogically and consist of quartz, K-feldspar, plagioclase and biotite (typically chloritized). Texturally, they range from aplitic to porphyritic and, in Granite III, to medium-grained equigranular. As they can all appear similar in hand specimen, distinction between the different granites is based to a large extent on mutual contact relationships. However, Granite I and Granite II are commonly fractured and highly altered and in many places contact relationships between these two intrusions are not obvious.

Granite I

Granit I occurs as irregular bodies closely associated with the North Zone breccias. Its contacts with the breccias are commonly gradational and fragments of Granite I are abundant locally within the breccias. In this regard it is analogous to fine-grained granite of the Fire Tower Zone. Granite I is typically fine grained and equigranular in relatively unaltered specimens. However, in most areas textural features of Granite I have been obscured by pervasive chloritic and/or silicic alteration. Porphyry tungsten-molybdenum deopsits in the North Zone appear to be related to Granite I.

Granite II

The main body of Granite II occurs at the base of the North Zone complex (Fig. 2). Dyke-like extensions of Granite II have intruded the overlying breccias and associated host rocks in places. Porphyry dykes that outcrop at surface may be related to Granite II.

Granite II varies from aplitic to porphyritic in texture. Parts of Granite II also contain features that reflect fluid saturation and/or undercooling of the parent magma, including miarolitic cavities and comb quartz layers. The combquartz layers consist of parallel to subparallel layers in which quartz crystals are oriented approximately perpendicular to the planes of layering (Fig. 4,5) (cf. Kirkham and Sinclair, in press). They are one of a family of unidirectional solidification textures (USTs) that are associated with fluid-saturated and/or undercooled magmas (cf. Shannon et al., 1982). Granite associated with the comb quartz layers commonly contains micrographic intergrowths of quartz and K-feldspar (Fig. 6), a feature that is also characteristic of undercooled granitic melts (Walker, 1976; Fenn, 1979).

Granite II consists of multiple phases which have internal contacts marked by chilled zones and USTs such as comb quartz layers. The distribution of USTs shown in Figure 7 appears to outline a small intrusive phase within Granite II. Figure 7 also shows an internal contact in diamond-drill hole C83 that is marked by aplitic layering. In general, these internal phases are either local or obscured by alteration and cannot be traced from one drillhole to another.

Tin deposits in the North Zone are associated mainly with Granite II. The Endozone, Contact Crest and Contact Flank deposits, for example, occur either within or at the contact of Granite II (Fig. 2). The Deep Tin Zone is also likely related to Granite II although not as closely associated spatially. Tin zones near surface in the North Zone (not shown in Fig. 2) are associated with porphyry dykes that may be offshoots of Granite II. Because of its relationship to tin deposits, Granite II is comparable to granite porphyry of the Fire Tower Zone, which is also associated with tin zones (Kooiman et al., 1986).

Granite III

Granite III forms a large body that underlies both the North and Fire Tower zones. In the North Zone, contact relationships show clearly that Granite III has intruded granites I and II. The contacts between Granite III and the other granites are commonly sharp and in many places are marked by thin (0.5 to 2 cm wide) layers of USTs, mainly K-feldspar, in Granite III. The upper contact of Granite III appears to have its apex west of the North Zone. Dykes of Granite III extend upward from the main body. Preliminary chemical data indicate that Granite III is identical to porphyritic granite of the Fire Tower Zone.

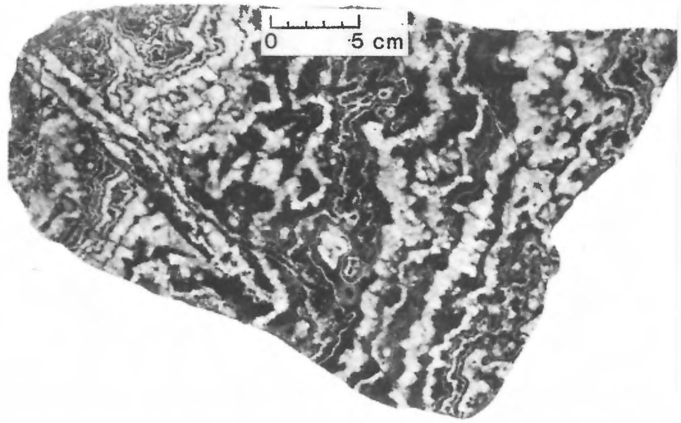

Figure 4. Comb quartz layers in granite porphyry from the Fire Tower Zone; quartz crystals in the layers terminate towards the right. GSC 204152-H

Figure 5. Photomicrograph (uncrossed nicols) of comb quartz layer in Granite II showing euhedral termination of quartz crystals approximately perpendicular to the layers; the very fine grained texture of the granite adjacent to the crystal faces probably resulted from pressure quenching related to the sudden release of the fluid phase in which the quartz crystals were growing. GSC 204152-T

Granite III is typically fine- to medium-grained equigranular although porphyritic and pegmatitic varieties are also present. Miarolitic cavities ranging from 1mm to 2 cm in size are locally abundant; in places they form 1 to 5 % by volume of the granite over several metres of drill core. The cavities are typically lined by quartz and feldspar crystals and filled by light green, very fine grained sericite (Fig. 8). Muscovite, chlorite, fluorite and carbonate are present in many cavities; arsenopyrite, molybdenite and other metallic minerals (unidentified) are rare.

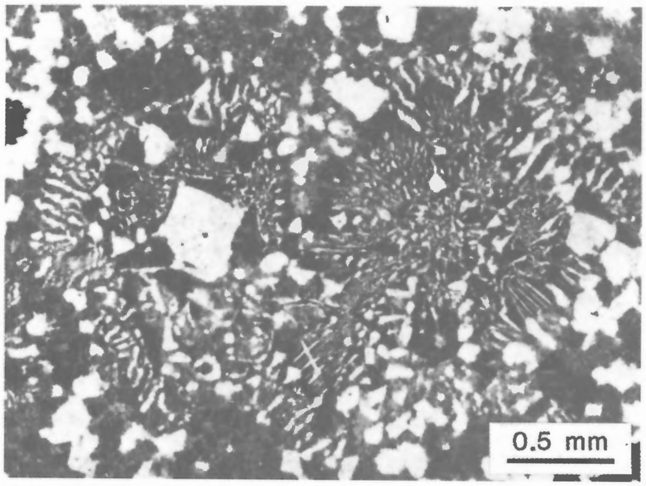

Figure 6. Photomicrograph (uncrossed nicols) of well-developed micrographic intergrowth of quartz (light) and K-feldspar (dark) in Granite II. GSC 204152-U

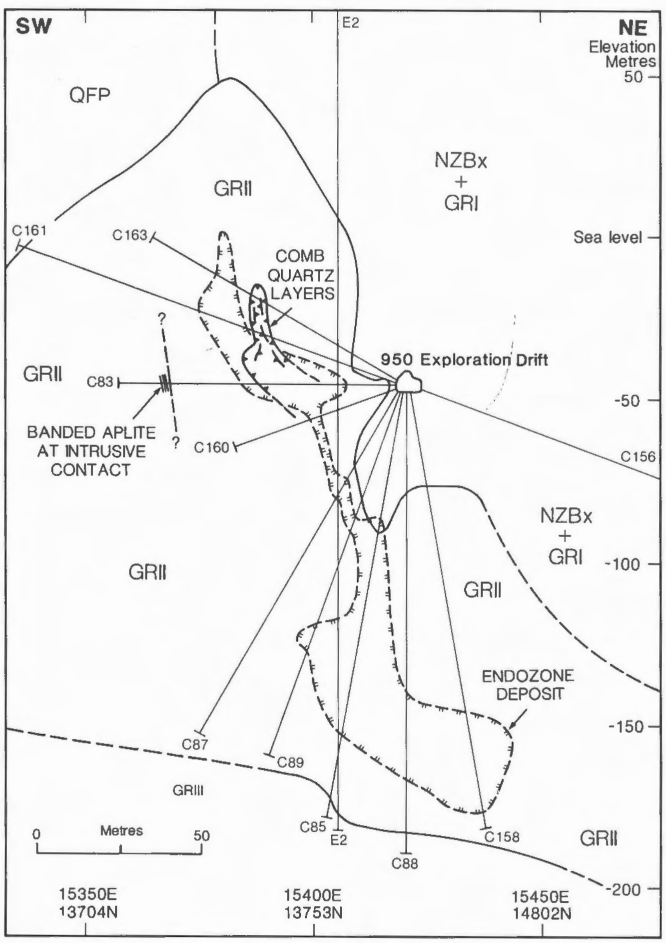

Figure 7. Cross-section (3A) through the Endozone deposit, North Zone (see Figure 1 for legend).

Granite III is generally less fractured and altered than Granite I and Granite II although some areas have been pervasively sericitized. Despite this sericitic alteration and fluid saturation textures such as the miarolitic cavities described above which are evidence of hydrothermal activity related to Granite III, only a few isolated tin zones have been found associated with it. However, exploration of Granite III to date has been too limited to assess fully its potential for associated deposits.

TIN DEPOSITS

Tin deposits and porphyry tungsten-molybdenum deposits in the North Zone represent two separate episodes of mineralization even though the deposits overlap spatially. Porphyry tungsten-molybdenum deposits resemble in many respects those in the Fire Tower Zone described by Kooiman et al. (1986). They consist of large zones of mineralized factures, quartz veinlets and disseminations in North Zone breccia and, to a lesser extent, Granite I. The host rocks have been altered to a very fine grained assemblage of quartz, topaz, fluorite and sericite. Wolframite, molybdenite, bismuth and bismuthinite are the principal ore minerals; abundant arsenopyrite and loellingite are also present. Total reserves in the porphyry tungsten-molybdenum deposits of the North Zone are approximately 11.5 million tonnes grading 0.20 % ;W, 0.06 % Mo and 0.09 % Bi.

Tin is distributed in the North Zone in a variety of deposits which span a wide vertical range. Near surface, numerous small, poorly-defined sulphide-rich replacement deposits occur in altered feldspar porphyry associated with aphanitic dykes referred to as "banded quartz-feldspar porphyry" by Ruitenberg (1967). These deposits have been

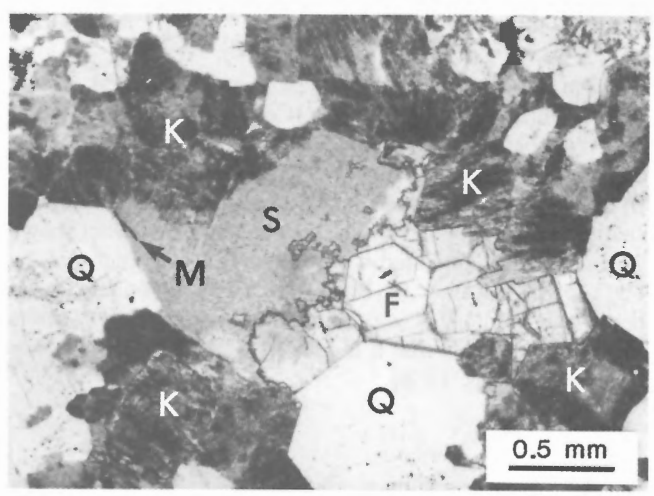

Figure 8. Photomicrograph (uncrossed nicols) of miarolitic cavity in Granite III lined with crystals of quartz (Q), K-feldspar (K), fluorite (F) and molybdenite (M) and filled with sericite (S). GSC 204152-V.

described by Ruitenberg (1963, 1967) and others (e.g. Hosking, 1963; Petruk, 1964). In these deposits, very fine grained cassiterite and lesser amounts of stannite are associated with arsenopyrite, loellingite, sphalerite and chalcopyrite. Other less abundants sulphides and sulpharsenides include pyrite, marcasite, galena, molybdenite, tennantite, bornite, bismuthinite, wittichenite and roquesite. Quartz, fluorite, topaz and chlorite are the principal alteration minerals. Proven reserves in these deposits are approximately 250,000 tonnes with an average grade of 0.6 % Sn, 2.3 % Zn, 0.30 % Cu and 0.36 % Pb (Mulligan, 1975).

Most of the potentially economic tin deposits in the North Zone occur at depths of 200 to 400 m below surfac. They include the Deep Tin Zone, Contact Crest, Contact Flank and Endozone deposits (Fig. 2). The Deep Tin Zone is a relatively large, irregular deposit that consists of fracture--controlled and disseminated cassiterite in silicified and chloritized North Zone breccia and Granite I. The cassiterite is fine grained but visible in hand specimens. Other associated minerals include arsenopyrite, sphalerite, chalcopyrite and galena. Some wolframite and molybdenite are also present, which may be due to the superposition of the Deep Tin Zone on one of the porphyry tungsten-molybdenum zones (not shown on Fig. 2). Reserves in the Deep Tin Zone are approximately 2.14 million tonnes grading 0.45 % Sn, 0.06 % W, 0.03 % ;Mo and 0.80 % Zn (0.45 % Sn represents a "cut" grade; intervals grading more than 2 % Sn have been cut to 2 %).

The Contact Crest and Contact Flank deposits occur mainly in North Zone breccia or other associated host rocks at the upper contact or along the sides of Granite II, respectively. The Endozone deposit, on the other hand, occurs mainly within Granite II. In many other respects, these deposits are similar. Cassiterite is the principal tin mineral and occurs as finely disseminated grains and as fine to medium-sized grains in veins or veinlets and along fractures. Minerals associated with the cassiterite include arsenopyrite, sphalerite, chalcopyrite, pyrite and pyrrhotite. Chlorite, fluorite, quartz and topaz are the main alteration minerals. Crosscutting relationships indicate as many as 6 stages of alteration and mineralization; details of the paragenesis of the Endozone, the largest of these deposits, is curently the subject of a Master's thesis study by C. Inverno (Colorado School of Mines). Total reserves in the Deep Tin Zone, Contact Crest, Contact Flank and Endozone deposits have been estimated at 5.9 million tonnes averaging ;0.79 % Sn (The Northern Miner, 14 October 1985, p. 11).

DISCUSSION

The Mount Pleasant intrusions were emplaced at the margin of the Mount Pleasant caldera in a subvolcanic environment. The present depth of the intrusions from surface (300 m or less) and the likelihood that no more than 600 to 700 m of overlying volcanic and sedimentary rocks have been removed by erosion suggest that the intrusions were emplaced at a depth of about 1 km or less.

The subvolcanic ;emplacement of the granitic intrusions has implications for the generation of the hydrothermal fluids that formed the ore deposits. In such a shallow environment of intrusion, involvements of both meteoric and magmatic fluids in the formation of the tin deposits might be expected. Preliminary data from a study of fluid inclusions (Samson, 1987) indicate that meteoric water may have been involved in at least one stage of tin mineralization. On the other hand, Granite II, which is most closely associated with the tin deposits, contains a number of features that suggest a significant portion of the ore-forming fluids were magmatic. The presence of miarolitic cavities and comb quartz layers in some phases of Granite II indicate that they were fluid-saturated. The same phases commonly contain micrographic intergrowths of quartz and K-fldspar. These textures reflect undercooled conditions in the magma which could have been due to pressure quenching caused by rapid separation of ore-bearing fluids. A genetic relationship between tin deposits and associated phases containing these features, such as the one outlined by the comb quartz layers on Figure 7, thus seems likely, Research currently in progress with B.E. ;Taylor (Geological Survey of Canada) to examine stable isotopes (O,H,C,S) of the granitic rocks and associated mineralized zones will provide insights to the nature of the ore-forming processes, including the relative importance of magmatic versus meteoric fluids.

Whether or not the Mount Pleasant intrusions are closely related to volcanic rocks within the Mount Pleasant caldera is uncertain as the age of the intrusions has not been completely resolved. K-Ar and Rb-Sr isotopic studies (Kooiman et al., 1986) indicated a Late Mississippian age of 340 to 330 Ma. On the other hand, $^{40}Ar/^{39}Ar$ data suggested a Late Devonian-Early Mississippian age of about 360 Ma (D.A. Archibald, pers. comm., 1985), essentially the same age as the Piskahegan Group rocks which the granites have intruded. More accurate and precise dating of the granitic rocks by U-Pb isotopic analysis of manozite is being considered to resolve this discrepancy. Rb-Sr isotopic studies of fluorite associated with the various granitic intrusions and associated deposits, currently underway with R.P. Taylor (Carleton University), may also contribute to a resolution of the age of the Mount Pleasant intrusions.

The Mount Pleasant intrusions have been classified as A-type granite (Taylor et al., 1985; Kooiman et al., 1986). Geochemical and petrographic work currently underway will help to characterize the intrusions at Mount Pleasant within a broader context of other granites in New Brunswick (e.g. Fyffe. et al., 1981; Ruitenberg and Fyffe, 1982; McLeod, 1986; Whalen, 1986).

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Observations sur la géologie glaciaire du nord-est de la Gaspésie, Québec[1]

J.J. Veillette[2]

Veillette, J.J., Observations sur la géologie glaciaire du nord-est de la Gaspésie, Québec; dans Recherche en cours, partie B, Commission géologique du Canada, Étude 88-1B, p. 209-220, 1988.

Résumé

On n'a trouvé aucune trace de la présence d'une glace laurentidienne active à l'intérieur de l'extrémité nord-est de la péninsule gaspésienne. Un seul bloc erratique d'âge précambrien a été observé au-dessus de la limite marine, à 170 m, et aucune marque d'érosion glaciaire sur le substrat rocheux n'a pu être relié à des glaces venant du nord ou du nord-ouest. Les stries glaciaires indiquent plutôt des écoulements dominants vers le NNE, dans le nord-ouest de la région, et vers l'ESE au sud et au sud-est de la région. Les grès de Gaspé, utilisés comme traceurs lithologiques, indiquent aussi une direction de transport vers le NNE au nord-ouest de la région. Des coquillages, à 40 m, dans la péninsule de Forillon, ont donné un âge de 13 100 ans ± 120 (CGC-4497), haussant ainsi de 10 m le niveau marin établi pour la transgression de la mer de Goldthwait. La présence d'un diamicton glaciomarin fossilifère, observé dans de nombreuses baies du littoral nord, est probablement due à l'action de glaciers de vallée survenue après que la mer de Goldthwait eut atteint son niveau maximal.

Abstract

No evidence was found to suggest the presence of active Laurentide ice in the interior of the northeastern part of the Gaspé Peninsula. Only one Precambrian erratic was observed, above the marine limit, at an altitude of 170 m, and no ice-flow indicators, associated with northward to northwestward flow of ice, were found. Instead, glacial striae indicate predominant ice flows to the NNE, in the northwest part of the area, and to the ESE in the southern and southeastern parts of the area. The use of Gaspé sandstone erratics as lithological indicators of glacial transport indicates a NNE ice flow direction in the northwestern part of the area. A date of 13 100 years ± 120 (GSC-4497) on shells, at 40 m, on Forillon Peninsula raises by 10 m the dated maximum level of the Goldthwait Sea. The fossiliferous glaciomarine diamicton present in several bays of the northern coast was probably deposited by valley glaciers from the interior postdating the maximum level reached by the sea.

[1] Contribution au Plan de développement économique Canada/Gaspésie et Bas Saint-Laurent, volet Mines 1983-1988.
[2] Division de la science des terrains.

INTRODUCTION

La Commission géologique du Canada, dans le cadre du Plan de développement économique Canada/Gaspésie et Bas Saint-Laurent, volet mines 1983-1989, a entrepris en 1985 de dresser la carte du Quaternaire de la péninsule gaspésienne à l'échelle de 1/250 000. Les cartes des formations en surface à 1/50 000, produites par le Ministère des Richesses naturelles du Québec au début des années 70, constituent la source première de renseignements pour cette compilation. Toutefois, certaines parties de la péninsule ont dû être étudiées de plus près avant de procéder à la compilation finale. Tel est le cas pour l'extrémité nord-est de la péninsule (fig. 1), soit les feuillets de Cloridorme 22 H2, Petit-Cap 22 H1W, Gaspé 22 A15 et Sunny Bank 22 A16, où les données étaient incomplètes. En plus d'une meilleure connaissance de la répartition des formations en surface, les principaux nouveaux apports sont: (1) la confirmation d'une absence de traceurs lithologiques d'origine précambrienne à l'intérieur de la péninsule, au-dessus du niveau maximal de transgression marine, (2) la présence d'écoulements glaciaires vers le NNE et l'ESE à partir de la marge ouest de la zone étudiée et, (3) une hausse d'au moins 10 m du dernier niveau daté de transgression marine par rapport au niveau maximal connu.

LE CADRE PHYSIQUE

La région (fig. 1) fait partie de la division physiographique des Appalaches, plus précisément de la sous-division des monts Notre-Dame (Bostock, 1970). La couverture de sédiments meubles étant généralement mince, le relief réflète fidèlement l'alignement NO-SE de la roche de fond (fig. 2). Il s'agit d'un ensemble de roches sédimentaires allant du Cambrien jusqu'au Dévonien inférieur et composé majoritairement de schistes ardoisiers et de grès terrigènes, mais aussi de calcaire et autres lithologies (McGerrigle et Skidmore, 1967; Brisebois, 1981). La disposition transversale des strates par rapport au sens dominant de l'écoulement des glaces a facilité l'étude du transport glaciaire.

Afin de circonscrire l'influence possible du relief sur l'écoulement glaciaire, une classification du territoire en tranches d'altitude a été utilisée (voir fig. 8). L'influence de la physiographie sera ainsi réexaminée plus loin de pair avec l'écoulement glaciaire.

TRAVAUX ANTÉRIEURS

Le premier travail d'envergure traitant de la géologie glaciaire de la Gaspésie remonte à Coleman (1922) bien que d'autres avaient déjà, avant cette date, fourni d'importantes observations (Bell, 1863; Chalmers, 1886, 1906; Goldthwait, 1911). McGerrigle (1952) a dressé les grandes lignes de la géologie du Quaternaire pour l'ensemble de la péninsule ajoutant aux travaux de ses prédécesseurs par des mesures, observations et hypothèses, qui encore aujourd'hui, forment l'ossature des connaissances actuelles. Plus récemment, suite aux travaux d'inventaire des formations en surface menés par le Ministère des Richesses naturelles du Québec dans

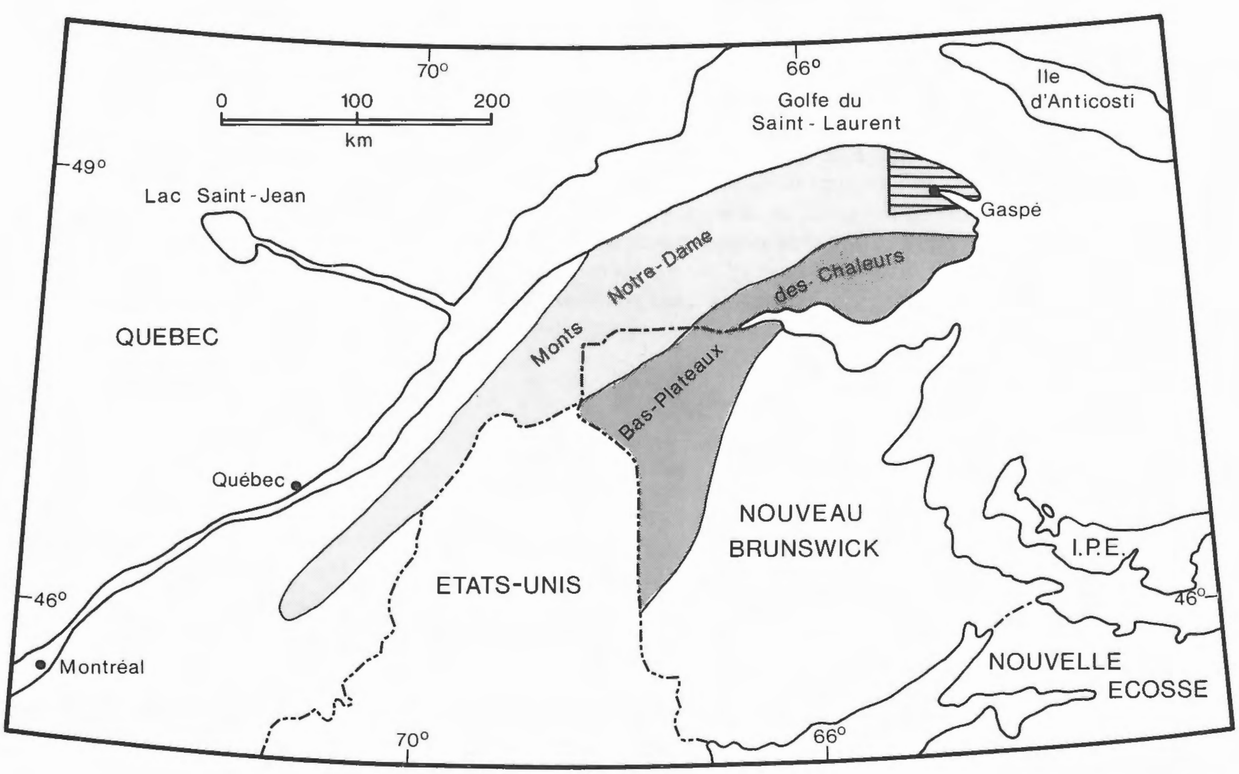

Figure 1. Localisation et physiographie.

l'ouest de la péninsule, Lebuis et David (1977) ont précisé les séquences glaciaires de la demie ouest. Ces travaux forment les assises d'un modèle de l'histoire glaciaire de l'ouest de la Gaspésie (David et Lebuis, 1985) qui comporte des implications importantes pour la demie est de la péninsule. Cette dernière partie est encore mal connue, les travaux de Chauvin (1984), Chauvin et coll. (1985) et Chauvin et David (1987) juste à l'ouest de la région décrite ici, ont établi les grands axes d'écoulement et de transport glaciaire à l'est des monts McGerrigle. Allard et Tremblay (1981) dans le Parc de Forillon, Bédard (1985) dans la région de Gaspé et Veillette (1987) dans la région de Cloridorme ont cartographié la répartition des formations en surface et fourni les données de base pour l'interprétation de l'histoire glaciaire. Enfin LaSalle (1984), Bail (1983, 1985) et Cloutier et Corbeil (1986a, 1986b) ont fourni d'importants renseignements sur la nature des sédiments de surface et sur les écoulements et le transport glaciaire des parties centrale et sud-est de la péninsule. Gray (1987) a fait la synthèse de la plupart de ces travaux, de même que ceux d'auteurs qui ont poursuivi des recherches plus thématiques.

LES FORMATIONS EN SURFACE

Une description détaillée des formations en surface ne constitue pas l'objectif du présent travail. Les unités décrites, ne le sont, que pour permettre une meilleure compréhension de l'histoire glaciaire. Le dépôt le plus répandu de la région (fig. 3), comme c'est le cas pour une grande partie de la péninsule, consiste en un diamicton lâche, fortement oxidé, à matrice sablonneuse et avec une abondance de fragments angulaires (fig. 4a). Les indices les plus communs de l'activité glaciaire, soit les cailloux polis et striés, sont rarement présents dans ce dépôt. Sur le sommet de la plupart des interfluves, au-dessus de 300 m, on ne trouve souvent qu'une mince couverture de fragments angulaires (fig. 4b) provenant de la désagrégation in situ du substrat rocheux. Par endroits, le roc est fortement saprolitisé (fig. 4c) et ceci sur des étendues cartographiables (Veillette, 1987). Dans le feuillet de Cloridorme (22 H2), il semble exister une certaine affinité entre le substrat calcareux et la formation d'une saprolite possédant une forte fraction argileuse (fig. 4d). L'origine et l'âge de ces dépôts demeurent problématiques et ces points

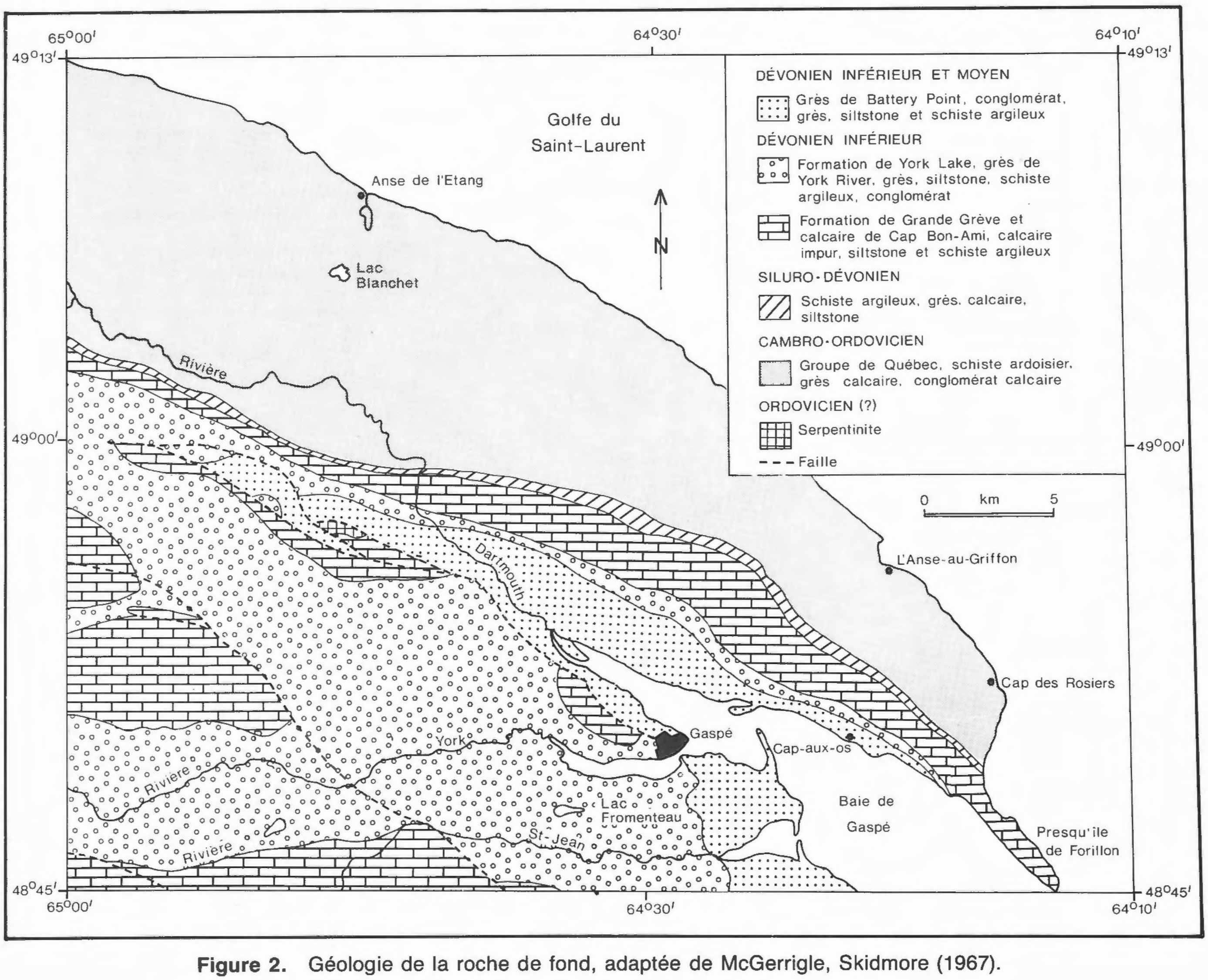

Figure 2. Géologie de la roche de fond, adaptée de McGerrigle, Skidmore (1967).

de discussion sortent des cadres de ce travail. Le lecteur trouvera dans Chauvin (1984), LaSalle (1984), David et Lebuis (1985) et Gray (1987), diverses descriptions de ce type de dépôts pour l'intérieur de la Gaspésie.

Les colluvions, étant donné le relief accidenté de la région, couvrent une grande superficie. L'ampleur des mouvements de pente a été clairement démontrée par Hétu et Gray (1980) dans la région d'Anse-Pleureuse. Ces dépôts, alliés à ceux décrits plus haut, donnent parfois lieu à une structure complexe (fig. 4e). Ils sont tous ici inclus dans la même unité, avec les barres rocheuses sans couverture de sédiments meubles, mais seront cartographiés séparément sur la carte de compilation finale.

Le till

En superficie, le till occupe le deuxième rang (fig. 3). Il s'agit d'un diamicton de plus d'un mètre d'épaisseur et dont l'origine glaciaire ne laisse aucun doute; galets et blocs striés abondants, matrice à dominance silto-argileuse et caractère polylithologique. Ce dépôt tapisse les flancs inférieurs et les fonds de vallée mais est absent du sommet des interfluves. Bien qu'on le trouve à différentes altitudes, sa plus grande superficie se manifeste sous le niveau de 150 m (fig. 3 et 8). Des coupes, fortes de plusieurs mètres, mais de faible étendue, ont été relevées dans le cours supérieur de la rivière Dartmouth (où la présence d'un till plus ancien demeure à vérifier), de même que dans plusieurs vallées fortement incisées à l'intérieur de la péninsule. Dans ces vallées, des lambeaux de till ont été conservés et sont par endroits exposés par les chemins forestiers. Enfin, il est à noter que la grande étendue de till sous la côte de 150 m repose aussi sur les grès du Dévonien inférieur et moyen (fig. 2, 3 et 8). Il est probable que le till accumulé dans ce secteur soit le produit de l'action conjuguée d'un système de glaciers alpins issus des vallées des rivières Saint-Jean, York et Dartmouth, qui se fondaient en un seul glacier dans la baie de Gaspé, lors du Wisconsinien supérieur.

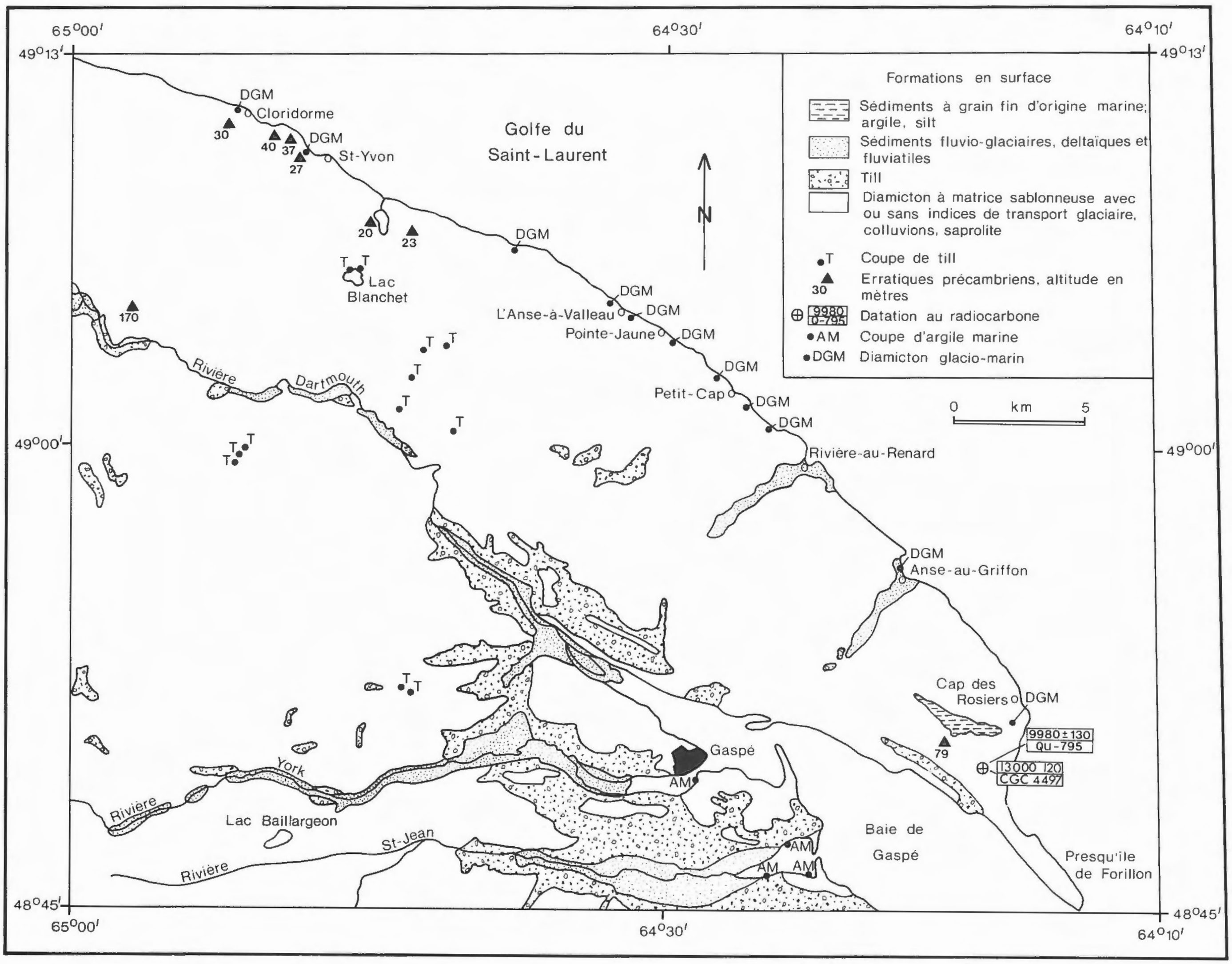

Figure 3. Les formations en surface, répartition et altitude des erratiques précambriens et distribution des âges radiocarbones.





Figure 4.
a. Diamicton lâche, à matrice sablonneuse (CGC 203990-W)
b. Couverture de fragments angulaires, roc altéré in situ. (CGC 203990-R)
c. Saprolite, des colluvions recouvrent la saprolite (schiste). (CGC 203990-S)
d. Saprolite argileuse formant l'assise du chemin. (CGC 203990-T)
e. Colluvions sur saprolite sur roc (schiste). (CGC 203990-U)

Étant donné la complexité des sédiments glaciaires de la péninsule gaspésienne, il est important de tenter de préciser les propriétés physiques et géochimiques de chaque type de sédiment afin de les comparer pour ainsi arriver à déterminer les processus responsables de leur mise en place. Des analyses ont été effectuées sur des échantillons de till (tab. 1) provenant majoritairement des roches cambro-ordoviciennes du groupe de Québec (fig. 5). Ce till présente une homogénéité un peu surprenante, tant granulométrique que géochimique, si l'on considère les lithologies variées du substrat dont il est majoritairement issu. Il est de couleur (humide) brun gris (2.5Y 5/2, charte de Munsell) lorsque peu ou pas altéré et brun jaunâtre (10 YR 5/4) lorsque fortement oxidé. Sa teneur en carbonate (fraction de moins de 63 μ), bien que très variable, phénomène d'ailleurs conforme à la grande variabilité de la roche de fond, dépasse rarement 20 %, avec une valeur médiane (n = 13) de 7,5 %. L'échantillon 85-VJ-19, pris à faible profondeur, illustre l'effet du lessivage sur la teneur en carbonate. L'échantillon 85-VJ-30, provenant du grès, marque aussi la plus faible teneur en carbonate de ce substrat. Le triage est pauvre (Tr = 4,5 phi) et le diamètre moyen des grains est de 5,5 phi (silt moyen à grossier); la figure 6 illustre graphiquement ces données.

Les sédiments fluvioglaciaires

Les sédiments fluvioglaciaires sont d'étendue restreinte et se limitent aux fonds de vallées (fig. 3). Ils sont regroupés ici avec les sédiments deltaïques et fluviatiles. Les baies profondes comme celles de Rivière-au-Renard et de l'Anse-au-Griffon contiennent de fortes accumulations de ces sédiments recouverts par des sédiments marins du littoral, ou interdigités avec ces derniers.

Les sédiments marins et glaciomarins

Ces sédiments n'occupent qu'une mince bande de la frange côtière. Il s'agit surtout de sables et graviers littoraux (fig. 3). Au sud-ouest de Cap-des-Rosiers, des argiles ont été signalées par Allard et Tremblay (1981) jusqu'au niveau de 76 m. Lors de la présente étude, des argiles fossilifères ont été échantillonnées à 40 m dans ce même secteur. Enfin, de petites plaques d'argile marine (fig. 3) ont été observées sur le pourtour des estuaires des rivières Saint-Jean et York, toutes deux situées à moins de 15 m d'altitude.

Les sédiments glaciomarins se présentent principalement sous la forme d'un diamicton argileux, massif, fossilifère et mal trié (DGM ; fig. 3 ; tab. 1) mais dont la courbe granulométrique témoigne de l'action de l'eau dans sa mise en place (fig. 7). Il est d'une couleur caractéristique gris foncé à bleuâtre (5 Y 4/1) et se trouve dans des endroits protégés dans de nombreuses baies de la rive nord (fig. 3). Cette dernière particularité est connue depuis les travaux pionniers (Bell, 1863; Chalmers, 1904; Coleman, 1922). La présence de galets précambriens dans ce dépôt a engendré diverses hypothèses sur son origine. Coleman (1922) a suggéré un contact

Tableau 1. Granulométrie et géochimie du till et du diamicton glacio-marin reposant sur les roches cambro-ordoviciennes (voir fig. 5).

Échantillon	Prof. (m)	Granulométrie % Sa	Si	Ar	Mz ϕ	Tr ϕ	Teneur en carbonates %	Géochimie (ppm) Cu	Zn	Pb	Cr	Mn	Fe % Fe	Co	Ni	As	U
85-VJ-15	1.6	39	33	28	5.8	4.9	7.50	66	148	17	68	500	4.7	19	100	8	<0.1
85-VJ-16	1.5	42	38	20	5.2	4.6	21.67	65	172	13	62	530	4.9	18	73	13	0.3
85-VJ-19	0.6	--	--	--	--	--	0.0	99	188	25	39	780	4.5	2	86	12	<0.1
85-VJ-25	1.2	40	50	20	5.4	4.0	10.83	30	100	15	76	1400	4.6	24	100	10	<0.1
85-VJ-26	4.0	30	38	32	6.6	4.5	11.67	26	104	10	74	1450	4.7	25	91	10	<0.1
85-VJ-27	4.0	41	35	24	5.7	4.4	5.00	37	112	17	130	1300	5.4	33	148	11	<0.1
85-VJ-28	4.0	54	34	12	3.8	3.8	9.17	44	122	12	84	840	4.9	29	103	9	<0.1
85-VJ-29-1	12.0	42	37	21	5.6	4.2	3.33	27	99	9	104	1100	5.1	29	134	8	<0.1
85-VJ-30	0.8	--	--	--	--	--	0.83	36	90	14	108	920	4.4	23	112	23	<0.1
85-VJ-31	1.0	30	40	30	6.2	4.7	11.67	50	133	18	75	400	5.2	19	81	7	0.5
85-VJ-32	2.5	51	29	20	4.6	4.8	23.33	56	140	14	69	630	5.1	28	71	7	<0.1
85-VJ-34	0.8	42	34	24	5.4	4.9	4.17	67	112	22	56	760	4.3	19	81	6	<0.1
85-VJ-35	1.0	27	45	28	6.0	4.6	1.67	51	150	23	50	900	4.6	22	66	9	<0.1
moyenne		40	37	24	5.5	4.5		50	128	16	77	885	4.8	24	96	10	<0.1
Glacio-Marin																	
85-VJ-23	1.5	--	--	--	--	--	13.33	48	118	22	77	880	4.5	24	109	9	0.3
85-VJ-24	1.5	47	37	16	5.0	3.5	9.17	29	117	20	56	520	3.7	15	65	5	<0.1

M_3 = moyenne graphique 1/3 (ϕ16+ ϕ50+ ϕ84)
Tr = indice de triage ½(ϕ84- ϕ16)

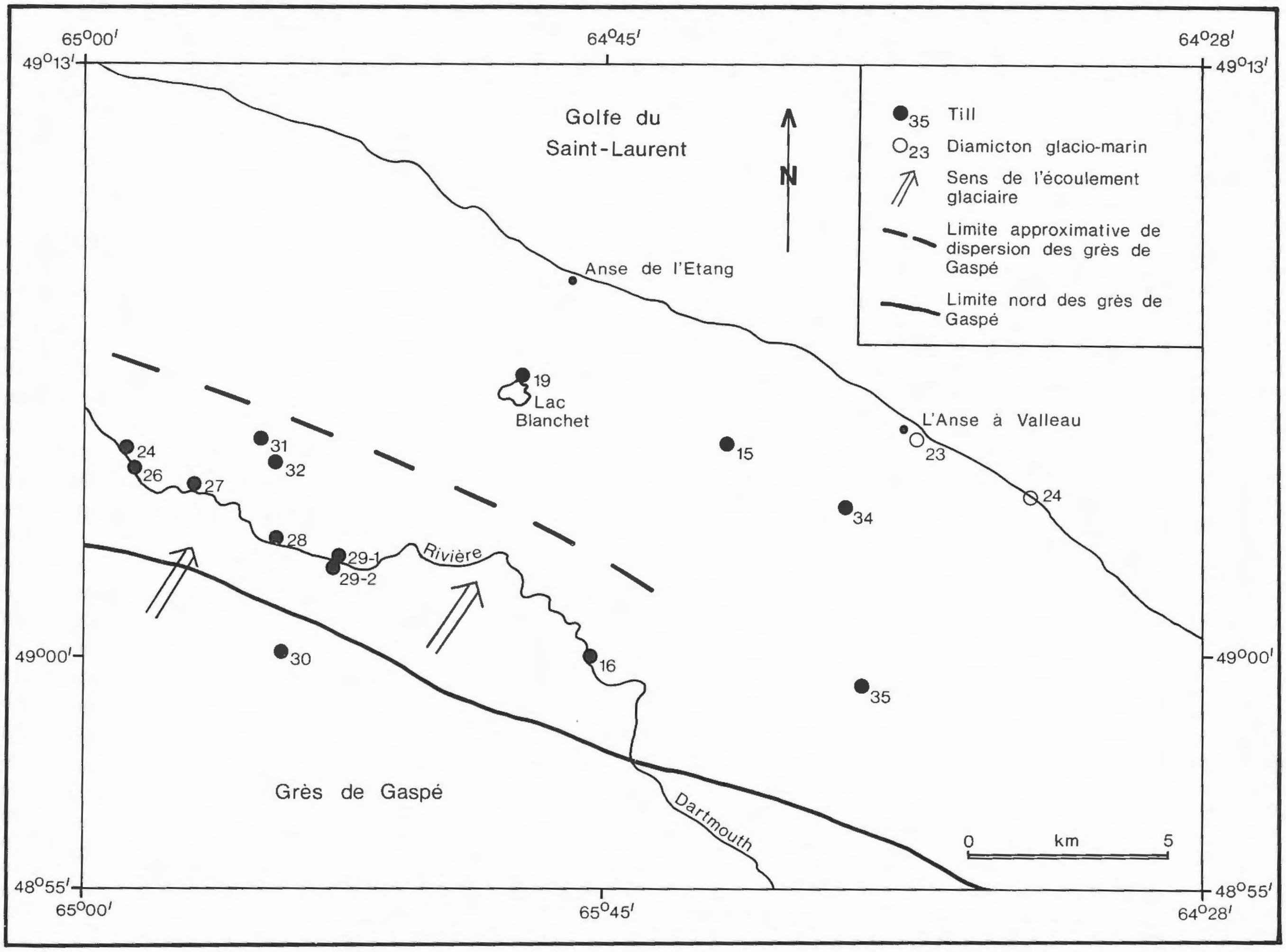

Figure 5. Localisation des échantillons du till (tab. 1) reposant sur les roches cambro-ordoviciennes et le transport glaciaire des grès de Gaspé (les grès du Dévonien inférieur et moyen, fig. 2).

de la calotte laurentidienne sur la rive nord de la péninsule mais non à l'intérieur des terres, ou encore la présence de glace de plate-forme, « ice-shelf », qui serait venue s'arrêter sur la rive nord de la péninsule. Allard et Tremblay (1981) ont revu ces hypothèses et inventorié des dépôts glaciomarins dans la presqu'île de Forillon semblables à ceux décrits ici. De même Chauvin (1984), dans la région juste à l'ouest, rapporte des âges de 10 660 ± 150 (QU-1433) et de 12 360 ± 170 ans (QU-1435) sur des fragments de coquilles dans des diamictons glaciomarins. Allard et Tremblay (1981) rapportent des âges de 11 810 ± 210 (QU-1115) et de 12 340 ± 170 ans (QU-1116) sur des diamictons glaciomarins à Rivière-au-Renard et à l'Anse-au-Griffon.

Il est donc probable que ce diamicton glaciomarin résulte simplement de l'avancée en milieu marin de glaciers de vallées provenant de l'intérieur de la péninsule. Un âge de 12 200 ± 200 BP (CGC-4321), à Grande-Vallée, juste à l'ouest de la région, s'ajoute aux âges donnés plus haut. Ces glaciers auraient ainsi incorporé les sédiments fins et les coquillages marins auxquels étaient mêlés les éléments précambriens apportés par les glaces flottantes. Cette hypothèse est conforme aux données suivantes : (1) l'âge de la mise en place du sédiment, soit environ 11 000 à 12 500 ans après que la mer de Goldthwait eut atteint son niveau maximal (Dionne, 1977) et, (2) la présence de stries sur le substrat rocheux, directement sous le diamicton. Ces dernières sont partout conformes aux grands axes des vallées où elles se trouvent. Dans les baies de l'Anse-à-Valleau, Pointe-Jaune, Saint-Maurice et l'Anse-au-Griffon (fig. 3), il a été possible de dégager le diamicton pour exposer le substrat. À ces endroits, le diamicton est en contact avec le substrat rocheux. Le glacier responsable de la mise en place du diamicton et de la formation des stries s'étendait donc bien en-dessous du niveau marin d'alors, incorporant ainsi les dépôts glaciels (précambriens) et les coquillages de la zone sub-littorale. Allard et Tremblay (1981) suggèrent que ces glaciers ont pu rester jusqu'à environ 1000 ans en contact avec les eaux marines.

ÉCOULEMENTS ET TRANSPORT GLACIAIRES

La figure 8 montre la localisation d'environ 125 surfaces striées, 36 sont de Bédard (1985) et quatre de Allard et Tremblay (1981). Les stries et autres microformes sur le substrat

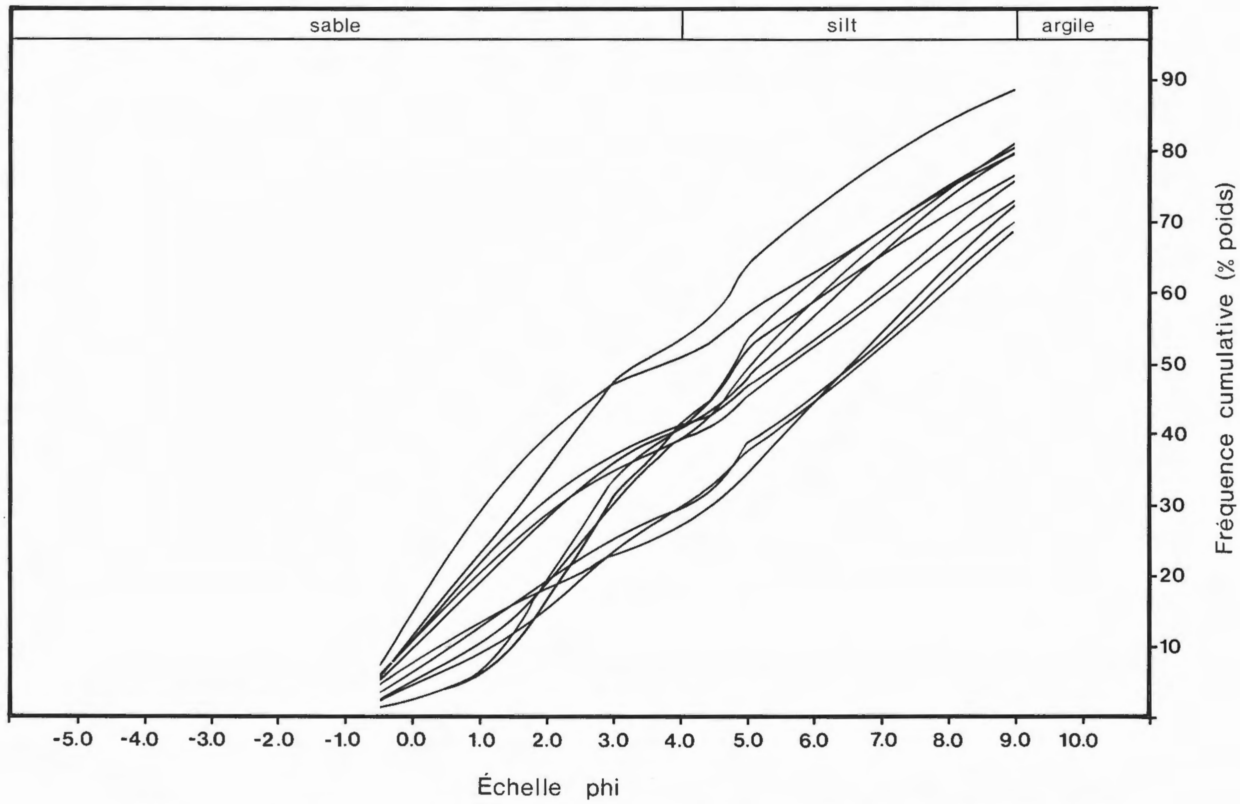

Figure 6. Enveloppe granulométrique du till reposant sur les roches cambro-ordoviciennes (tab. 1).

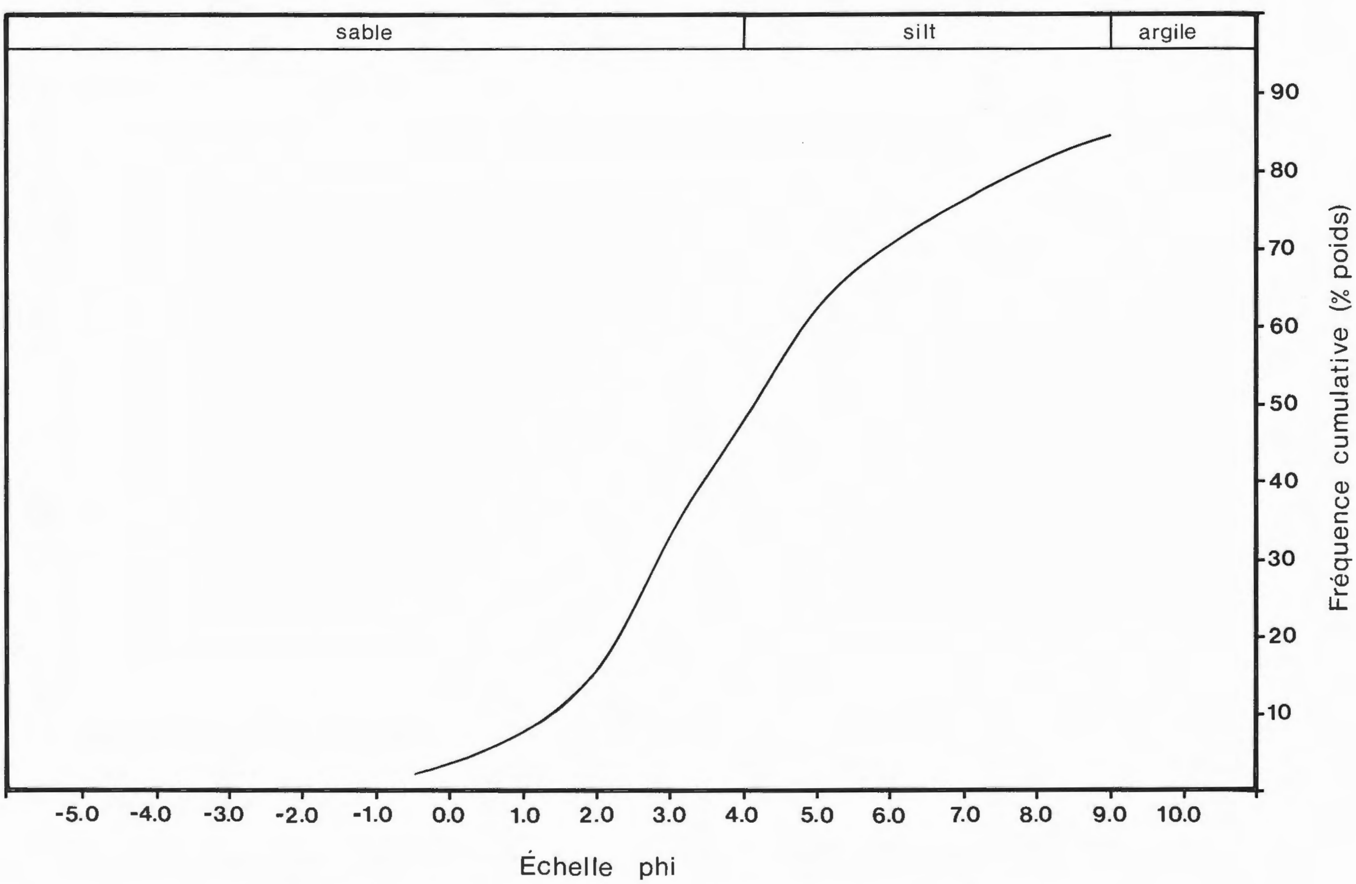

Figure 7. Courbe granulométrique du diamicton glaciomarin (VJ-85).

rocheux semblent être les seuls indices permettant une analyse satisfaisante des complexités de l'écoulement glaciaire. Les macroformes de terrain, soit les roches profilées et les dépôts de contact glaciaire, ne sont pas assez bien développées pour permettre cette analyse.

La région a été découpée en quatre classes d'altitude (fig. 8) afin d'évaluer l'influence possible de la topographie. Les grands axes suivants se dégagent: (a) du nord-ouest vers le sud-est de la région le sens principal de l'écoulement passe graduellement du NNE à l'ENE, sans influence marquée de la topographie, et, (b) dans la partie sud de la région les écoulements vers l'est et vers l'ESE dominent. La source de la glace responsable de ces écoulements se trouvait dans les hautes terres à l'ouest, entre les rivières Dartmouth et York. Ce secteur semble avoir marqué une zone de divergence des glaces entre les grands axes d'écoulements NNE et ESE, (c) des écoulements convergents dans les vallées des rivières York, St-Jean et Dartmouth vers la baie de Gaspé. Allard et Tremblay (1981) avaient déjà observé un écoulement parallèle à la rive nord de la baie de Gaspé dans la région de Forillon, tandis que Bédard (1985) avait observé un écoulement vers l'est sur la rive sud de la baie.

Ces observations résument assez bien la séquence d'écoulements de la région. Toutefois, l'absence de marques indiquant un écoulement régional ancien vers le sud ou le SSE, comme c'est le cas pour d'autres endroits de la péninsule où le glacier laurentidien a laissé des traces, est significatif. Des stries, vers le sud et le SE, ont été mesurées, mais elles ont toujours été associées à des glaciers de vallées, et datent probablement de la déglaciation tardive. Ainsi, au SE du lac Blanchet (fig. 8), une longue vallée en V, orientée du nord au sud et joignant le sommet du plateau, au nord, à la rivière Dartmouth, au sud, illustre bien l'influence locale du relief à des époques différentes. De nombreuses stries, dans la partie sommitale de la vallée et sur le plateau, indiquent un écoulement vers le NNE, tandis que dans la partie inférieure de la vallée, près de la rivière Dartmouth, des stries et des queues-de-rat indiquent un écoulement vers le sud, parallèle aux parois de la vallée. Les marques du sommet seraient donc reliées à une glace plus épaisse, indépendante de la topographie locale, et provenant du centre de la péninsule, c'est-à-dire au sud et au SO de la région. Il est probable que les trains de dispersion de roches volcaniques vers le SE, rapportés par LaSalle (dans Gray, 1987) dans le sud-est de la

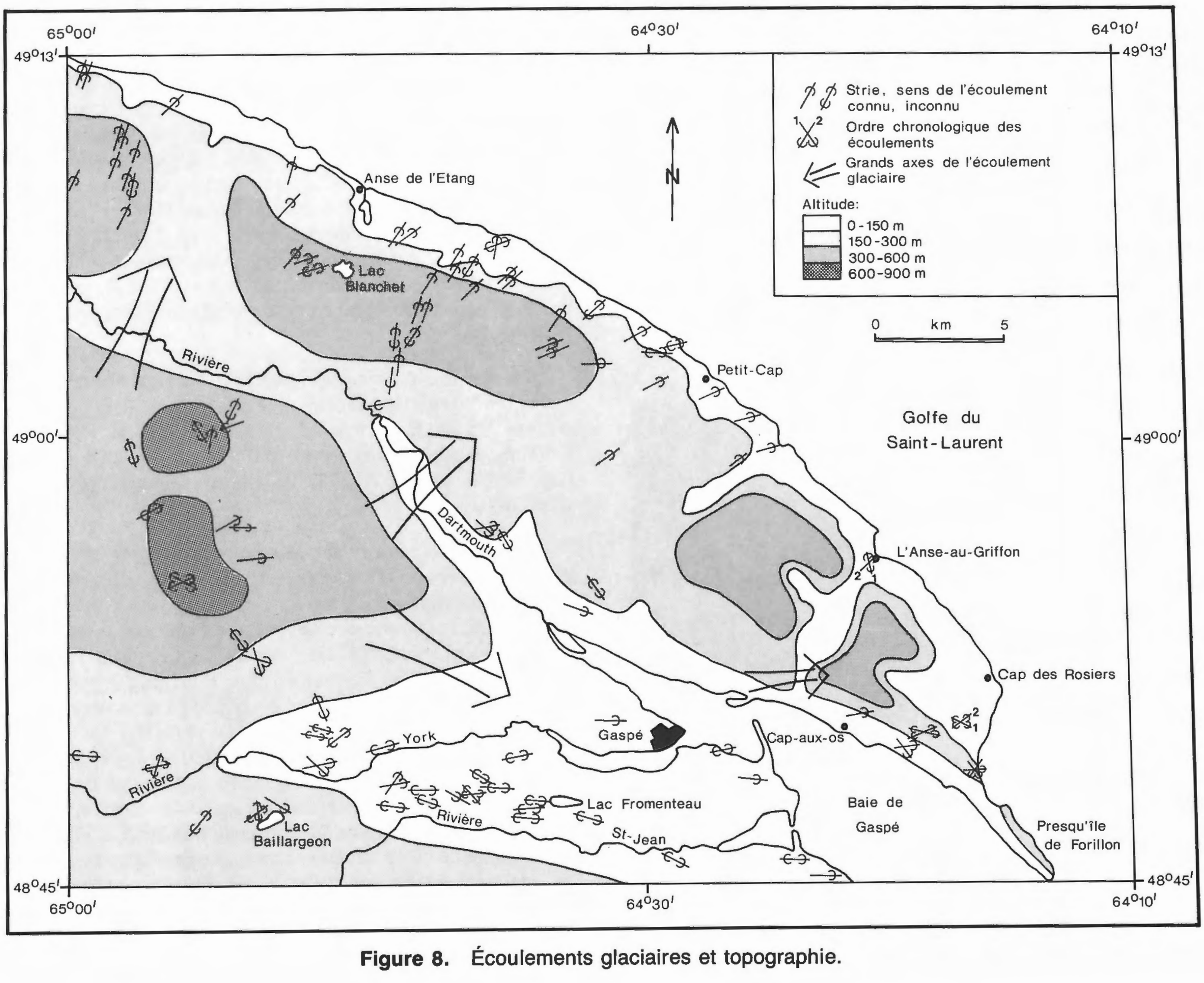

Figure 8. Écoulements glaciaires et topographie.

péninsule marquent le flanc sud de cette zone de divergence (ice divide) dans la calotte gaspésienne au Wisconsinien supérieur.

Le transport des débris glaciaires a été étudié dans la partie nord de la région seulement. Afin de vérifier l'influence des écoulements glaciaires nord et NNE sur le transport des débris glaciaires, les grès de Gaspé (grès du Dévonien inférieur et moyen; fig. 2; Brisebois, 1981) ont été choisis comme traceurs lithologiques (fig. 5). Environ 200 emplacements de pointage de blocs et d'examen de galets et cailloux sur le terrain ont été visités dans le quart nord-ouest de la région. Ces mesures jointes à des comptages en laboratoire des granules (plus de 2 mm) et des cailloux du till a permis de tracer une limite approximative de dispersion vers le nord et le NNE des grès de Gaspé (fig. 5). Malgré l'encaissement de la vallée de la rivière Dartmouth, plus ou moins perpendiculaire à l'écoulement glaciaire, ce déplacement vers le NNE est évident. Les grès sont facilement observables jusqu'à environ 5 km au nord du contact avec les roches cambro-ordoviciennes. Au nord de cette ligne, ils s'estompent rapidement. Ce phénomène peut être dû soit à un inventaire incomplet, soit à la fragilité de ces grès qui résistent sans doute mal au transport glaciaire. De toutes façons, leurs déplacements, même sur cette courte distance, correspond bien au sens de l'écoulement régional.

Enfin, malgré une recherche constante à l'intérieur du même secteur nord-ouest, un seul bloc de granite d'origine présumément précambrienne a été observé (fig. 9). Il s'agit d'un bloc d'environ 1,0 m de diamètre trouvé dans un rapide de la rivière Dartmouth à environ 170 m d'altitude (fig. 3). Il constitue le seul indice lithologique d'un apport laurentidien, au-dessus de la limite marine. Il se peut toutefois, étant donné sa position dans le cours d'eau, qu'il ait été déplacé sur une distance considérable d'abord par le glacier de vallée de la rivière Dartmouth, et ensuite par les glaces de la rivière.

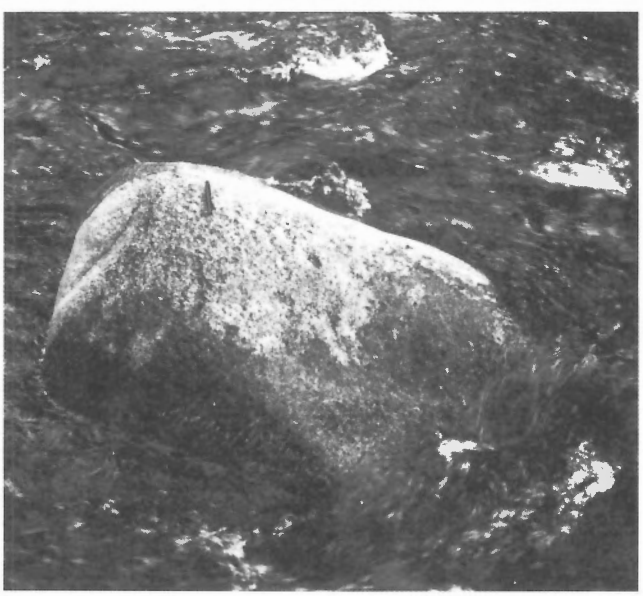

Figure 9. Granite précambrien dans un rapide de la rivière Dartmouth (fig. 3) à 170 m (le stylo fait 12 cm). (CGC 203990-V)

À l'est, dans le parc de Forillon, Allard et Tremblay (1981) proposent un glacier centré sur les crêtes du centre de la presqu'île pour expliquer le transport glaciaire vers le sud et vers le nord. Ainsi les blocs de calcaire des formations de Cap Bon-Ami et de Grande Grève provenant de la crête rocheuse médiane (fig. 2) se retrouvent à l'ensemble du parc de Forillon. Par contre, les blocs de grès de Gaspé, du nord de la baie de Gaspé, semblent restreints à leurs roches-mères, et ne se retrouvent pas sur la rive nord de la presqu'île.

LA LIMITE MARINE

Il est difficile de déterminer la limite de la dernière transgression marine sur le pourtour nord-est de la péninsule dû à une côte rocheuse élevée aux falaises trop abruptes d'une part et, d'autre part, à la confusion engendrée par la présence d'encoches, de terrasses et autres formes d'érosion marine plus anciennes que celles de la mer de Goldthwait. Sans tenter de distinguer entre ces marques anciennes et celles de la mer de Goldthwait, Allard et Tremblay (1981) estiment un niveau marin maximal de 76 m au cap Bon Ami, tandis que les premiers travaux (Bell, 1863; Coleman, 1905) rapportent 73 m dans la région de Gaspé.

Les estimés récents de la transgression goldthwaitienne sont beaucoup plus conservateurs. Prest et coll. (1968) la place à 23 m à la pointe de la presqu'île de Forillon et à 29 m sur la rive SE de la baie de Gaspé. Dionne (1976) rapporte 30 m à Cap des Rosiers (plages) et suggère des niveaux probables à 46 m à l'Anse-au-Griffon et à Rivière-au-Renard. Les sédiments fossilifères les plus élevés, à 30 m, au sud de Cap-des-Rosiers, ont donné un âge de 9980 ± 130 BP (QU-795, Allard et Tremblay, 1981). Ces derniers auteurs rapportent aussi des lignes de rivage à une altitude comparable (30 m) à d'autres endroits sur la rive nord du parc de Forillon, mais considèrent comme probable une limite marine plus élevée.

La dernière transgression marine (avec présence de sédiments fossilifères) a atteint un niveau supérieur à ces estimés récents. Un diamicton fossilifère de la presqu'île de Forillon à 40 m, donne un âge de 13 100 ans ± 120 CGC-4497 (fig. 3). Une analyse détaillée de la micro-faune à cet endroit se poursuit.

La frange côtière étant peu propice à la conservation des marques d'érosion littorale, le relevé de blocs paléoglaciels d'origine précambrienne a été utilisé pour estimer le niveau maximal de la submersion marine. La technique a été bien éprouvée par Dionne (1972) le long du Saint-Laurent, en aval de Québec. De plus, l'extrême rareté d'erratiques précambriens à l'intérieur du nord-est de la péninsule confère une certaine crédibilité à son utilisation dans la région. Un relevé systématique des replats en terrain ouvert, pâturage ou autres, à plusieurs endroits entre Cloridorme et Cap-des-Rosiers, entre 30 et 100 m d'altitude, indique une concentration anormale de blocs précambriens dans cette bande (fig. 3). Une quarantaine de blocs ont été comptés, en prenant soin d'éliminer tout bloc susceptible d'avoir été déplacé par l'homme. Le plus élevé dans l'ouest de la région, à Cloridorme, repose à 40 m, bien que des terrasses avec galets (suspects) discoïdes ont été trouvées au niveau de 79 m. On retrouve de fortes concentrations de blocs dans les baies ou paléo-baies profondes comme celle de L'Anse à l'Étang (fig. 3). Au sud du Cap-des-Rosiers, on a relevé trois blocs précambriens, reposant à la surface, à 79 m. Dans ce même secteur un galet de granite, bien que suspect, a été trouvé à 129 m. Par contre, les blocs paléo-glaciels précambriens semblent absents de la baie de Gaspé. Il est probable que cette baie ait été occupée par un glacier de vallée durant la transgression initiale de la mer de Goldthwait. Le niveau d'argile marine le plus élevé est à seulement 15 m dans l'estuaire de la rivière Saint-Jean (fig. 3), mais aucun indice sûr n'est disponible pour établir le niveau maximal de transgression qui devrait être de beaucoup supérieur à 15 m.

CONCLUSION

Des travaux récents dans le nord-est de la péninsule gaspésienne ont rendu possible l'achèvement de la cartographie des formations en surface pour la demie nord de la péninsule afin de procéder à une compilation finale à 1/250 000. Les contributions suivantes sur la géologie glaciaire de la région sont à retenir.

1. Une extrême rareté d'erratiques glaciaires d'origine précambrienne à l'intérieur des terres au-dessus de la limite marine. Un seul erratique a été trouvé.
2. Conjointement à cette absence, aucune trace d'écoulement glaciaire pouvant être reliée à une provenance laurentidienne, c'est-à-dire du nord ou du NO, n'a été observée.
3. Des sens d'écoulements glaciaires dominants vers le NNE dans la partie nord-ouest de la région, et vers l'ESE dans les parties sud et sud-est de la région. Il est probable qu'un vaste système de glaciers de vallées, impliquant les vallées des rivières Dartmouth, York et Saint-Jean, a alimenté un glacier centré dans la baie de Gaspé lors de la déglaciation.
4. La dernière transgression marine a atteint un niveau d'au moins 40 m, et probablement plus élevé, dans la presqu'île de Forillon, et le long de la côte nord de la péninsule. Le niveau de 40 m a été daté à 13 100 ± 120 (CGC-4497).
5. Un diamicton glaciomarin fossilifère présent à la base de la séquence stratigraphique dans de nombreuses baies et échancrures de la côte sur la rive nord, résulte de l'avancée de glaciers de vallées qui ont remanié les dépôts marins de la zone sub-littorale.

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Re-examination of the Frozen Ocean Group: juxtaposed middle Ordovician and Silurian volcanic sequences in central Newfoundland.[1]

H. Scott Swinden[2]

Swinden, H.S., Re-examination of the Frozen Ocean Group: juxtaposed middle Ordovician and Silurian volcanic sequences in central Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 221-225, 1988.

Abstract

The Frozen Ocean Group, as previously defined, is bipartite. The northern part consists of Ordovician pillow lavas and interbedded epiclastic rocks which are tentatively correlated with the Big Lewis Lake basalts of the Wild Bight Group to the north. No felsic volcanic rocks are present. These rocks are not correlated with massive sulphide-bearing sequences in the Wild Bight Group and are not belived to be prospective, although this remains to be tested by whole-rock analyses. The southern part includes red to orange, subaerial rhyolites and mafic volcanic rocks, interpreted as Silurian on the basis of lithological correlations with the Springdale Group to the west. There may be a potential in the southern area for epithermal gold mineralization.

The Frozen Ocean Group name should be dropped. Formal definition of these rocks awaits systematic mapping. No rocks within this sequence are correlatives of the Buchans-Robert's Arm Belt.

Résumé

Le groupe Frozen Ocean, tel qu'il a auparavant été défini, est bipartite. La partie nord se compose de laves en coussins ordoviciennes et de roches épiclastiques interstratifiées qui ont été provisoirement corrélées aux basaltes de Big Lewis Lake du groupe de Wild Bight plus au nord. Aucunes roches volcaniques felsiques ne sont présentes. Ces roches ne sont pas corrélées à des séquences massives riches en sulfures au sein du groupe de Wild Bight et on ne les estime pas prometteuses, bien que ceci reste à démontrer par des analyses effectuées sur des échantillons de roche totale. La partie sud comprend des rhyolites subaériennes dont la couleur varie de rouge à orange, ainsi que des roches volcaniques mafiques, que l'on attribue au Silurien en s'appuyant sur des corrélations lithologiques avec le groupe de Springdale situé plus à l'ouest. Il peut exister un potentiel de minéralisation aurifère épithermal dans la zone sud.

Le nom « Frozen Ocean Group » (groupe de Frozen Ocean) devrait être abandonné. Une définition officielle de ces roches exige une cartographie systématique préalable. À l'intérieur de cette séquence, aucune des roches n'est corrélative de la zone de Buchans et Robert's Arm.

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

Previous metallogenic investigations for this project have been carried out in the Robert's Arm and Buchans Groups (Swinden and Sacks, 1986; Swinden, 1987). During the 1987 season, work was concentrated in the Frozen Ocean Group (Dean, 1978).

The Frozen Ocean Group, although not contiguous with the Robert's Arm and Buchans groups (Fig. 1) has previously been correlated with them by Dean (1978) and Kean et al., (1981). The only published map of the group is the compilation of Dean and Strong (1976) and this provides little information regarding internal stratigraphy or mineral potential. Exploration company mapping in this area on the late 1970's and early 1980's (e.g. Fenton, pers. comm. 1981) outlined the internal stratigraphy of the sequence and identified substantial amounts of both volcaniclastic sedimentary rocks and felsic volcanic rocks. The possible presence of felsic volcanic accumulations, and their possible economic significance, prompted the present field investigations.

PREVIOUS WORK

Rocks currently assigned to the Frozen Ocean Group were first mapped on a regional scale by Williams (1964) who included them in the Wild Bight Group, an early Ordovician sequence of oceanic volcanic and sedimentary rocks, which outcrops extensively between the Frozen Ocean Group and the coastline to the north (see also Swinden, 1984).

Dean and Strong (1976) compiled the geology for this area at 1:50 000 scale, showing the Frozen Ocean Group as a lithostratigraphic unit separate from the Wild Bight Group. Dean (1978), in a detailed description of the Dean and Strong (1976) compilation, formally defined the Frozen Ocean Group as comprising volcanic and lesser sedimentary rocks in the area west and southwest of New Bay Pond. He noted that most of the northern part of the group consisted of mafic volcanic and pyroclastic rocks with minor interbedded sedimentary rocks, while felsic volcanic rocks of unknown extent, with features suggesting subearial deposition, were present in the southern part. He suggested that these overlay the subaqueous rocks to the north, perhaps forming the top of the sequence. Dean (1978) correlated the Frozen Ocean Group with the Buchans — Robert's Arm belt to the north and west based on inferred regional stratigraphic relationships. At that time, the Buchans — Robert's Arm belt was interpreted by many to be post-middle Ordovician in age, based mainly on an inferred stratigraphic relationship between it and underlying, fossiliferous shale and turbidites (the

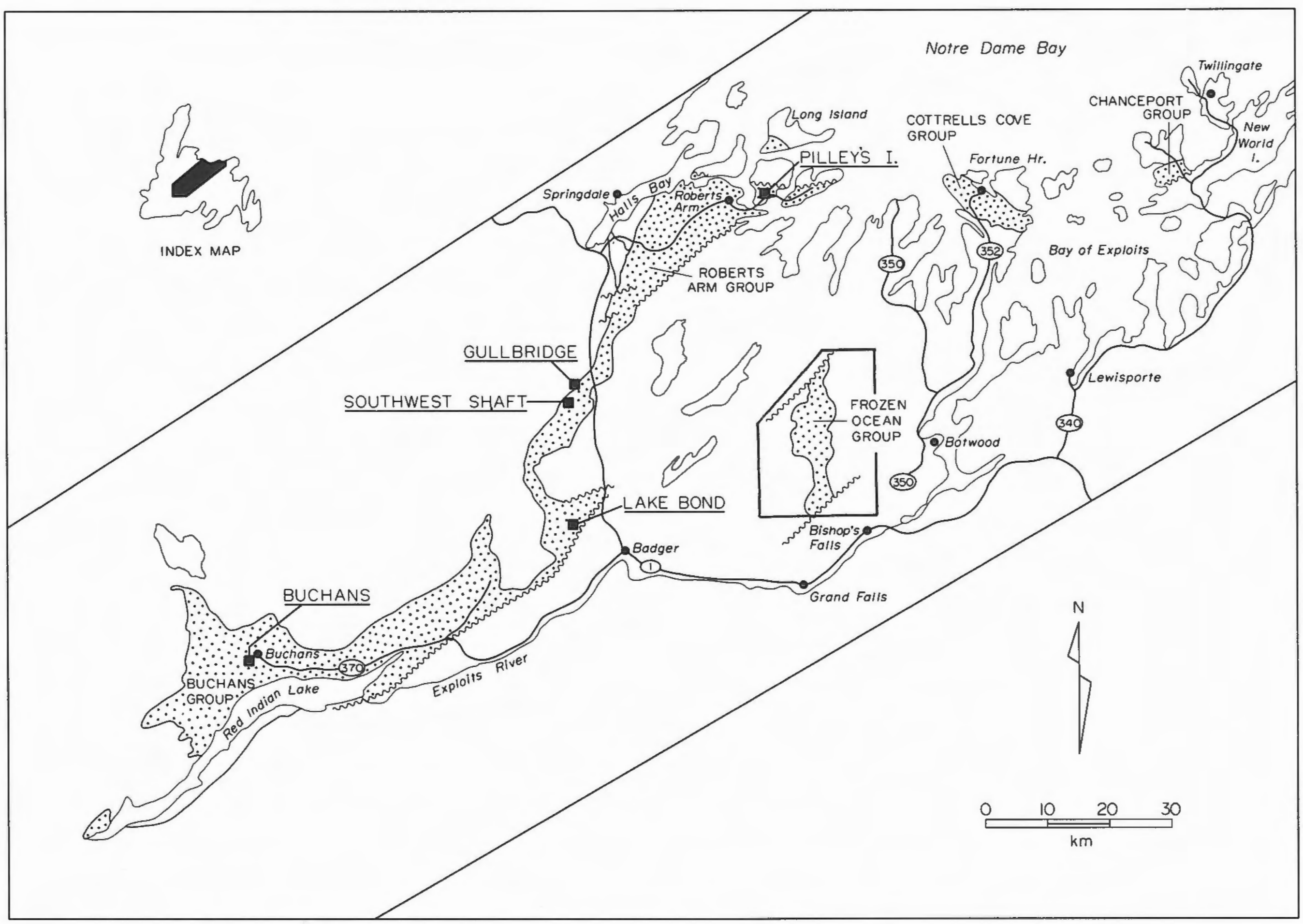


Figure 1. Location of the Buchans — Roberts arm Belt and rocks generally considered to be correlative (stippled). The Frozen Ocean Group, as previously defined, is indicated. Area of Figure 2 is enclosed by solid lines.

"Caradocian shale" and overlying Sansom Greywacke) of Middle Ordovician to early Silurian age.

Based on the identification of this stratigraphic succession in a southwest-facing sequence north of New Bay Pond (where Ordovician rocks of the Wild Bight Group were interpreted to be overlain successively by Caradocian shale, Sansom Greywacke and Frozen Ocean Group mafic volcanic rocks), the Frozen Ocean Group was interpreted as post-Middle Ordovician in age. It was, therefore, considered to be a stratigraphic correlative of the Buchans — Robert's Arm Belt and this correlations was supported by observed lithological similarities between the mafic pillow lavas in the Frozen Ocean and Robert's Arm Groups (Dean, 1978).

Subsequent work, however, has raised further questions about both the age of the Buchans — Robert's Arm belt and the stratigraphic setting of the Frozen Ocean Group (and hence the correlation of the two). Evidence from fossils in the Buchans Group (Nowlan and Thurlow, 1984) and radiometric (U/Pb in zircon) age determination of the Buchans and Robert's Arm Groups (Dunning et al., 1987) have unequivocally demonstrated that the Buchans — Robert's Arm Belt is early Ordovician in age. The contact between it and the middle Ordovician to early Silurian sedimentary rocks (the Caradocian shale and Sansom Greywacke) must, therefore, be structural. This, of course, negates the stratigraphic line of reasoning that was the basis for separation of the Frozen Ocean Group from nearby strata. Furthermore, the contact between the Frozen Ocean Group and the Sansom Greywacke has recently been reinterpreted as structural, according to Kusky and Kidd (1985). Because there is no longer a stratigraphic necessity to correlate the Frozen Ocean Group and the Buchans — Robert's Arm Belt, the question is raised as to what the Frozen Ocean Group should be correlated with (i.e. whether a correlation with the Buchans Robert's Arm Belt is still tenable) and whether a separately-defined Frozen Ocean Group is even necessary.

GEOLOGY OF THE "FROZEN OCEAN GROUP"

Mapping in the area west and southwest of New Bay Pond shows that the Frozen Ocean Group, as it is currently defined, is composite. It consists of two distinct lithostratigraphic assemblages (Fig. 2) apparently separated by an arm of the Hodges Hill Granite west of the southern end of New Bay Pond. For convenience in the following discusssion, the northern unit is informally termed the "New Bay Pond sequence" and the southern unit the "Charles Lake sequence". These are described separately below.

New Bay Pond sequence

The New Bay Pond sequence consists of a succession of northeast-facing, marine volcanic and sedimentary rocks, It can be subdivided into three units, a lower mafic volcanic unit, a middle epiclastic turbidite unit and an upper mafic volcanic unit.

The lower mafic volcanic unit consits mainly of massive basalt and mafic volcanic breccia. The mafic flows are rarely pillowed, although locally, there are possible pillow fragments in volcanic beccias. The main rock type is massive, aphanitic basalt which locally contains large amygdales filled with epidote. There is considerable interbedded volcanic breccia, consisting of massive to vesicular basalt fragments, generally unsorted and angular to subangular, in an aphanitic green matrix that is probably tuffaceous. There are isolated outcrops of more silicious rocks, particularly immediately west of New Bay Pond, but these are locally clearly intrusive and it is not clear whether any are actually volcanic.

The medial epiclastic unit consists mainly of greywacke and pebbly conglomerate with lesser argillite and coarse, polylithic conglomerate. The greywacke is commonly well bedded, and rock fragments, where identifiable, are generally mafic volcanic rocks. Volcanic rocks also dominate the clast assemblages of the coarse-grained rocks, which are generally unsorted and matrix supported. The strata locally display partial Bouma (1962) sequences, and are interpreted as turbidites which have reworked primary volcanogenic material. Abundant sedimentary structures provide top determinations and demonstrate that, although there are local reversals, the sequence is dominantly northeast-facing.

The upper volcanic sequence includes a larger proportion of pillow lava, and recognizable pillow breccia, which is the dominant lithology. Red, ferruginous chert and interbedded magnetite were noted in one locality.

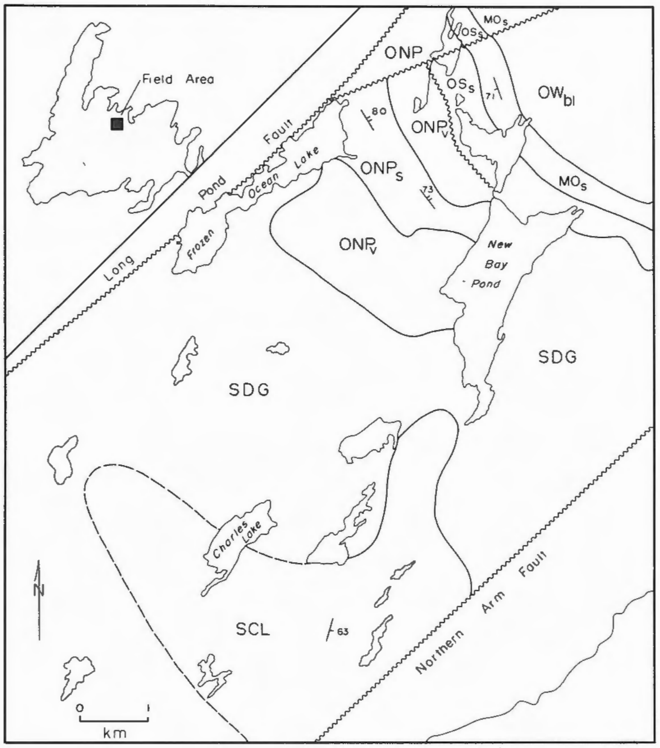
Because the New Bay Pond sequence apparently faces northeast and the adjacent Sansom Greywacke faces southwest (Fig. 2), the contact must be a fault. This fault is exposed on a bush road near the north end of New Bay Pond, where intense shearing is accompanied by silicification and minor pyritization. To the northwest, the New Bay Pond sequence is faulted against volcaniclastic rocks of the Wild Bight Group along the Long Pond Fault. All other contacts are intruded by plutonic rocks of presumed Devonian age related to the Hodges Hill granite.

Charles Lake sequence

The Charles Lake sequence consists of a generally northeast-striking volcanic assemblage that is lithologically very different from the New Bay Pond sequence. The dominant lithologies are red to orange, silicic volcanic rocks and massive pale to dark green mafic volcanic rocks. The two appear to be present in approximately equal amounts, although exposure is not good enough, nor mapping detailed enough, to estimate the proportions accurately. The felsic volcanic rocks include ash-flow crystal tuff, locally containing well-developed lithophysae, massive, fine-grained rhyolite and rheoignimbrite. Red silicic quartz-feldspar porphyries which are lithologically very similar to the volcanic rocks locally show good intrusive relationships to surrounding rocks and it appears that the silicic rocks in this unit represent a high-level intrusive-extrusive volcanic complex.

Mafic volcanic rocks are generally aphanitic, locally plagioclase-phyric or amygdaloidal.

Coarse grained, polymictic conglomerate and/or volcanic breccia containing a clast assemblage representative of all local rocks types occurs here and there. The clasts are generally poorly sorted, subangular to subrounded and texturally, rocks range from clast- to matrix-supported.



SILURIAN OR DEVONIAN
SDG — Plutonic rocks. Includes granite, granodiorite, diorite and gabbro.

SILURIAN
SCL — Charles Lake sequence: red to orange rhyolite, rhyolitic ash flow tuff, massive, intermediate to mafic volcanic rocks, lesser polylithic conglomerate and/or volcanic breccia.

UPPER ORDOVICIAN TO SILURIAN
OSs — Sansom Greywacke: volcanically — derived turbiditic greymacke, lesser argillite.

MIDDLE ORDOVICIAN AND OLDER
MOs — Shoal Arm Formation: carbonaceous shale, chert.

OWb1 — Wild Bight Group (Big Lewis Lake volcanics): dominantly pillow breccia, lesser pillow lava, epiclastic rocks.

ONP — New Bay Pond sequence: v — dominantly pillow breccia, lesser pillow lava, epiclastic rocks; s — epiclastic turbiditic greywacke, lesser polymictic conglomerate, minor argillite.

Figure 2. Geology of the area west and southwest of New Bay Pond showing, emphasizing rocks previously assigned to the Frozen Ocean Group.

No evidence of submarine environments (e.g. marine sedimentary rocks, pillow lavas) was observed in the Charles Lake sequence. This, coupled with the presence of strongly welded ignimbrites in the felsic volcanic rocks, suggests a subaerial environment (a conclusion previously reached by Dean, 1978).

The contacts between the Charles Lake sequence and the adjacent plutonic rocks are generally not exposed but are inferred to be intrusive as dykes of the plutonic bodies cut the volcanic sequence. To the south, the volcanic rocks are apparently faulted against the Botwood Group along the Northern Arm Fault (Fig. 2).

REGIONAL CORRELATIONS

The two lithostratigraphic units which make up the Frozen Ocean Group, as it is now defined, clearly represent different physical and magmatic environments. Both have correlatives in the nearby Ordovician and Silurian record that suggest an age difference as well.

The lithology of New Bay Pond sequence is very similar to the Big Lewis Lake basalt in the Wild Bight Group (Swinden, 1984), which outcrops immediately to the northeast (Fig. 2). The lithological similarity is particularly evident in the relatively abundant pillow breccia and fine-grained hyaloclastite in the two sequences, and in the relative scarcity of pillow lava. Epiclastic rocks are virtually identical in the two units. The Big Lewis Lake volcanic rocks are stratigraphically overlain by the southwest-facing Caradocian shale and Sansom Greywacke and on this basis are inferred to be Llandeilian or Caradocian in age. The correlation with the New Bay Pond sequence suggests that this sequence is also of this age.

The lithologies in the Charles Lake sequence contrast sharply with both the New Bay Pond sequence and also with Ordovician volcanic and sedimentary sequences elsewhere in Central Newfoundland. Subaerial red rhyolite with associated intermediate to mafic volcanic rocks and fluviatile conglomerated are generally associated with Silurian events in Central Newfoundland. Some rocks in the Charles Lake sequence are very similar to those in the Springdale Group to the west. In particular, the lithophysae-bearing ash flow tuff, flow-banded red rhyolites and quartz-feldspar porphyry and strongly-welded, lithophysae-bearing ash flow tuff show a close lithological correspondence with rocks in the Springdale Caldera, recently described by Coyle et al. (1985) and Coyle and Strong (1987). Similar lithologies may also occur within the Silurian Botwood Group which adjoins the Charles Lake sequence to the southeast across the Northern Arm Fault.

The northeast-trending fault separating the New Bay Pond sequence from the Sansom Greywacke is a structural anomaly in this part of central Newfoundland where most documented faults have a northeasterly to easterly strike. This raises the possibility that there may be other faults with orientations close to bedding in the sequence, and that significant, unrecognized, structural breaks may be present, for example, in the adjacent Ordovician successions of the Wild Bight Group.

ECONOMIC IMPLICATIONS

The mapping reported here has implications for the mineral exploration potential of the area west of New Bay Pond. The Wild Bight Group, with which the New Bay Pond sequence is correlated, is host to four volcanogenic massive sulphide deposits (Swinden, 1984). One of these, the Point Leamington deposit (Noranda Mines Staff, 1974; Swinden, 1984), associated with a felsic volvanic dome within the Side Harbour volcanic unit (Swinden, 1984), is only 5 km to the northwest of the most northerly New Bay pond sequence exposures. The previous suggestion by Fenton (pers. comm. 1981) that a felsic volcanic centre was present in the central part of the New Bay Pond sequence, was taken as evidence that these rocks might be similarly prospective. The present study, however, indicates that the felsic rocks in this area are, at least in part, intrusive and of very limited extent. There does not appear to be a felsic volcanic centre in the southern part of the Wild Bight Group. Furthermore, the Big Lewis Lake basalt flows, which are a better lithological correlative of the New Bay pond sequence than the Side Harbour sequence, are stratigraphically and geochemically distinct from volcanic rocks associated with the Point Leamington massive sulphide deposit (Swinden, 1985). If their correlation with the New Bay Pond sequence is correct, then stratigraphic equivalence with the ore-bearing rocks is unlikely. Geochemical studies of the volcanic rocks are in progress to ascertain whether rocks similar to those associated with mineralization elsewhere in the Wild Bight Group are present.

The suggestion of a Silurian age for the Charles Lake sequence raises the possibility that gold-bearing, epithermal, mineralization similar to that described by Tuach (1986, 1987) in coeval rocks in White Bay, may be present. Although no evidence of such mineralization was encountered during the present study, it is emphasized that the mapping was of a reconnaissance nature and there is considerable scope for additional mapping and prospecting in the area.

CONCLUSIONS

The Frozen Ocean Group, as previously defined, is considered to be an inappropriate lithostratigraphic name. The group is actually bipartite, comprising a northern unit of Middle Ordovician mafic volcanic and epiclastic rocks (the "New Bay Pond sequence") and a southern unit of probably Silurian felsic and mafic volcanic rocks (the "Charles Lake sequence"). It is suggested that the name "Frozen Ocean Group" be dropped. The New Bay Pond sequence is more properly considered as part of the Wild Bight Group, and need not be assigned a more specific name until such time as this group is systematically mapped and subdivided. The Charles Lake sequence should be formally assigned formation or group status, but this awaits systematic and detailed mapping of the whole sequence, so that the boundaries and stratigraphic succession can be properly described.

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Preliminary kinematic analysis of the Bass River Complex, Cobequid Highlands, Nova Scotia[1]

R. Damian Nance[2] and J. Brendan Murphy[3]

Nance, R.D. and Murphy, J.B., Preliminary kinematic analysis of the Bass River Complex, Cobequid Highlands, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 227-234, 1988.

Abstract

The Bass River Complex forms part of the northern Appalachian Avalon terrane and comprises the Great Village River Gneiss, the platformal(?) metasedimentary Gamble Brook Formation, and the mafic metavolcanic Folly River Formation. Three major periods of deformation within the complex are considered to be Precambrian. The earliest (D1) affects only the gneiss and suggests the existence of basement-cover relations. The second additionally affects the Gamble Brook Formation and records polyphase ductile shear (D2a/D2b) on basement-cover contacts attributed to sinistral transtension at about 630 Ma. Late kinematic, locally sheeted mafic dykes suggest limited crustal extension and may herald the development of possible back-arc basinal conditions in the Folly River Formation. A third deformation (D3a/D3b) may record dextral transpression and polyphase, NW-directed thrusting in the Folly River Formation. D3 is tentatively attributed to basin closure during the latest Precambrian and may have accompanied emplacement of the late Precambrian Jeffers Group.

Résumé

Le complexe de Bass River forme une partie du nord du terrane d'Avalon de l'Appalachien et comprend le gneiss de Great Village River, la formation plate-formale(?) métasédimentaire de Gamble Brook et la formation métavolcanique mafique de Folly River. Trois périodes importantes de déformation dans le complexe sont considérées comme étant précambriennes. La première (D1) concerne seulement le gneiss et suggère l'existence de relations entre le soubassement et la couverture. La seconde touche en plus la formation de Gamble Brook et présente un cisaillement ductile polyphasé (D2a/D2b) aux contacts de la couverture et du soubassement attribué à une transtension sénestre d'environ 630 Ma. Plus récemment, des dykes mafiques stratifiés localement et cinématiques suggèrent une extension limitée de la croûte et pourraient annoncer le développement possible d'un bassin marginal dans la formation de Folly River. Une troisième déformation (D3a/D3b) pourrait témoigner d'une pression transversale dextre et d'un chevauchement multiphasé dirigé vers le nord-ouest dans la formation de Folly River. D3 est attribuée hypothétiquement à la fermeture du bassin pendant la fin du Précambrien et peut avoir accompagné la mise en place du groupe de Jeffers à la fin du Précambrien.

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INTRODUCTION

The Bass River Complex of the eastern Cobequid Highlands, Nova Scotia (Fig. 1), occupies a narrow belt between the dextral Cobequid and Rockland Brook Faults and is thought to form part of the late Precambrian metamorphic infrastructure of the northern Appalachian Avalon terrane (Donohoe, 1983; Cullen, 1984). Based on structural criteria within the complex, Donohoe and Wallace (1980, 1985) recognized an older, amphibolite facies "basement" unit (the Great Village River Gneiss) and two younger, greenschist facies "cover" units (the metasedimentary Gamble Brook Schist and the mafic metavolcanic Folly River Schist). Both cover units have since been assigned formational status (Murphy et al., 1988). In addition, basement-cover contacts between the Great Village River Gneiss and the Gamble Brook Formation were shown to be associated with younger, syntectonic granite gneisses. Detailed descriptions of these units and their polyphase structural history are given by Cullen (1984) and Donohoe and Wallace (1985), and are only briefly summarized here. However, the kinematic history of the complex, which forms the focus of this initial report, has received little attention since Eisbacher (1969, 1970) attributed the deformation of the complex to Devonian dextral shear. Wider aspects of the ongoing re-investigation of the Cobequid Highlands, of which this study is part, are described by Murphy et al. (1988).

LITHOLOGIC UNITS

Great Village River Gneiss

Basement lithologies of the Great Village River Gneiss include massive to quartz-plagioclase layered, hornblende amphibolites (Bass River Amphibolite; Cullen, 1984), hornblende granitoid orthogneisses (Great Village River Orthogneiss; Cullen, 1984), and biotite/garnet-rich, psammitic paragneisses (Donohoe and Wallace, 1985). To these, Cullen (1984) added the pelitic Portapique River Schist which locally contains sillimanite and staurolite and occurs in isolated outcrops adjacent to basement-cover contacts. Cullen considered the schist to be a member of the basement on the basis of its metamorphic grade and the pretectonic nature of the high-grade phases relative to the earliest deformation of the cover. However, their lithologic similarity to units of the otherwise lower grade Gamble Brook Formation, their stratigraphically and structurally restricted occurrence, and their association with granite gneiss bodies that intrude the Gamble Brook Formation without the development of chilled margins, suggest that the Portapique River Schist originated through high-grade metamorphism of cover lithologies adjacent to the basement gneisses. The Great Village River Gneiss has been correlated with the Mount Thom Complex (unit 1a; Fig. 1) of the eastern Cobequid Highlands (Donohoe and Cullen, 1983), the orthogneisses of which have yielded a questionable Rb/Sr whole-rock age of 934 ± 82 Ma (Gaudette et al., 1984).

Syntectonic Granite Gneisses

The strongly mylonitic contacts between the Great Village River Gneiss and the Gamble Brook Formation are commonly associated with heterogeneously deformed granite gneiss bodies which range in size from veins and dykes to narrow plutons that follow the contacts for several kilometres. However, in contrast to their host rocks, the intensity of mylonitization in these granite gneisses varies widely over very short distances. Larger bodies commonly crosscut the principal (D2) mylonitic fabric in the basement and the cover without apparent offset, yet show well-developed C-S fabrics and asymmetric augen that are coplanar with the stronger fabrics of their host rocks. Smaller bodies range from concordant, strongly mylonitic sills to crosscutting networks of essentially undeformed veins. These relations, coupled with the geochronological constraints on the timing of deformation discussed below, support Cullen's (1984) contention that the emplacement of the granite gneisses was broadly syntectonic with respect to D2. The gneisses have yielded Rb/Sr ages of 642 ± 15 Ma and 626 ± 22 Ma (Gaudette et al., 1984).

Gamble Brook Formation

The metasedimentary Gamble Brook Formation represents the older of the two cover units and separates basement from the younger Folly River Formation. The contact between the Gamble Brook Formation and the Great Village River Gneiss is interpreted to be a ductile shear zone (Murphy et al., 1988). Adjacent to the basement, lower portions of the Gamble Brook formation are locally arkosic but are dominated by mylonitic quartzites, quartz-biotite-garnet schists and thin carbonates. Structurally and stratigraphically overlying portions, however, are distinctly more pelitic and contain significant amounts of biotite-garnet and biotite-muscovite schists in addition to quartzites and quartz-rich schists (Cullen, 1984). Murphy et al. (1988) respectively refer to these successions as the lower and upper Gamble Brook formations (Table 1). The depositional setting of the formation is uncertain. However, the orthoquartzite protolith of much of the formation and the association of thin carbonates and overlying pelitic horizons are consistent with its deposition in a shallow-marine, platformal setting. If this is correct, the Gamble Brook Formation may be broadly correlative with other Precambrian platformal successions of the Avalon terrane such as the Green Head Group of southern New Brunswick (Wardle, 1978) and the George River Group of Cape Breton Island (Milligan, 1970).

Folly River Formation

Metavolcanic rocks of the Folly River Formation comprise schistose, biotite-bearing greenstones; greenschist facies mafic flows, pyroclastics and porphyritic hyaloclastites; and minor volcanogenic sediments including purple, jasperitic iron formations and pyritiferous quartz-sericite schists (Cullen, 1984). However, much of the formation comprises locally sheeted, plagiophyric mafic dykes that intrude the Gamble Brook Formation and closely resemble higher grade, crosscutting amphibolite dykes within the basement. The mafic complex locally contains pillow basalts that are associated with, and stratigraphically overlain by, phyllites and banded quartz-rich schists that resemble distal turbidites.

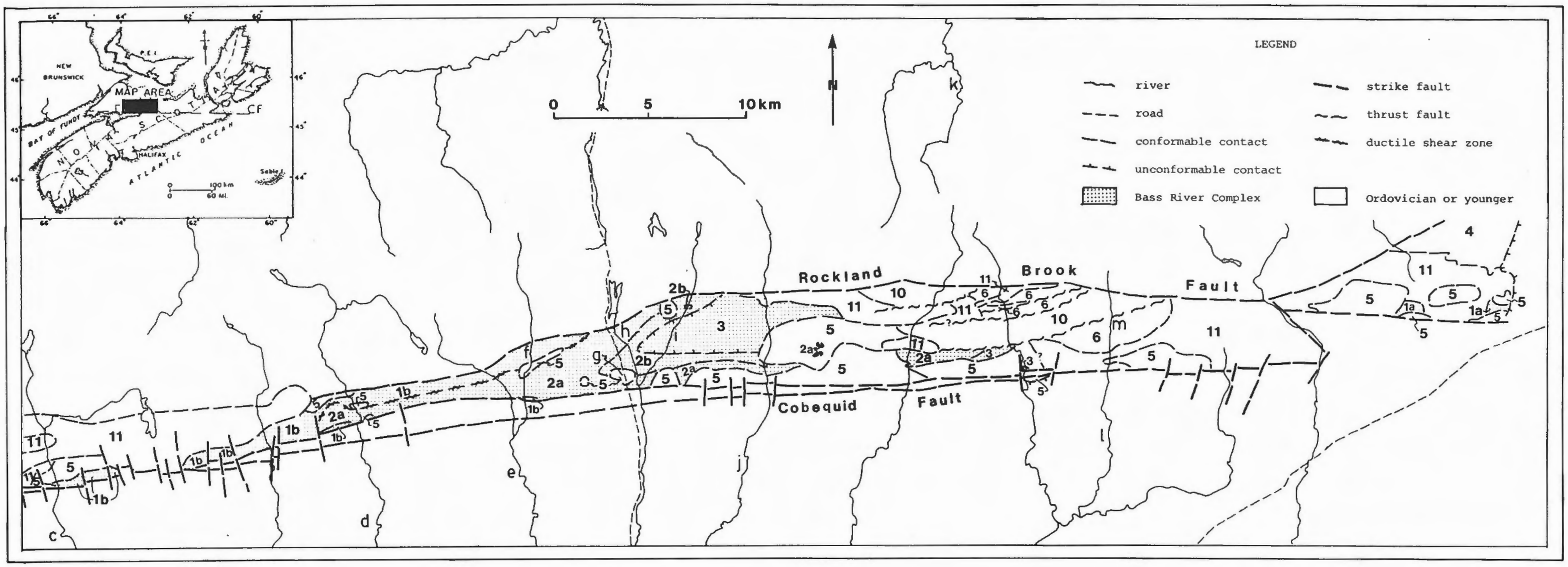

Figure 1. Geological map of the Bass River Complex and part of the eastern Cobequid Highlands, Nova Scotia (simplified after Murphy et al., 1988). See Table 1 for legend. Identified locations are as follows: (c) Economy River, (d) Portapique River, (e) Great Village River, (f) Rockland Brook, (g) Folly Mountain quarry, (h) Folly Lake quarry, (i) East Folly River, (j) Debert River, (k) Miller Brook, (l) North River, and (m) Nutby.

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

DEVONO-CARBONIFEROUS	11	Devono-Carboniferous plutons
	10	Nutby Formation
ORDOVICIAN-SILURIAN	6	undivided
PRECAMBRIAN-CAMBRIAN	5	Late Precambrian-Cambrian plutons
	4	Dalhousie Mountain volcanics
		BASS RIVER COMPLEX
	3	Folly River Formation
HADRYNIAN		unconformity
	2	Gamble Brook Formation (a) lower (b) upper
		unconformity?
HADRYNIAN?	1	(a) Mount Thom Complex (b) Great Village River Gneiss

In view of these associations, the volcanics are likely to be related to an extensional tectonic setting although their geochemistry is presently unknown. If correct, the Folly River Formation, which is intruded by several appinitic diorite and syenogranite bodies with Rb/Sr and K/Ar ages in the range 540-625 Ma (Donohoe and Wallace, 1985; Donohoe et al., 1986), may be broadly correlative with mafic volcanics within the late Precambrian Georgeville Group in the northern Antigonish Highlands (Murphy and Keppie, 1987).

DEFORMATIONAL HISTORY

Within the Bass River Complex, Donohoe and Wallace (1980, 1985) and Cullen (1984) recognized three phases of Precambrian deformation which they designated DB1, DB2 and DB3. They further showed that the earliest phase (DB1), which is represented by an amphibolite facies foliation, affects only the Great Village River Gneiss and interpreted this to demonstrate the existence of basement-cover relations between the gneiss and the Gamble Brook Formation. Subsequent deformation was considered to affect all units of the Bass River Complex such that the second phase of deformation in the basement (DB2) coincided with the first phase in the Gamble Brook and Folly River formations. We also recognize these phases and similarly attribute Precambrian structures to three major deformational episodes (D1, D2 and D3). However, while D1 corresponds to DB1, DB2 and DB3 are interpreted as two phases (D2a and D2b) of a single progressive deformation that records the development of a ductile shear zone between the Gamble Brook Formation and the basement, but which is largely absent in the Folly River Formation. Deformation in the latter formation is kinematically distinct to D2 and is assigned to a younger, polyphase (D3a and D3b) progressive deformation as described below.

D1 Structures

Rare evidence of the earliest phase of deformation (D1) occurs within the Great Village River Gneiss where the earliest recognizable folds (to which the D2a mylonitic fabric is axial planar) deform an older metamorphic fabric (Fig. 4 in Donohoe and Wallace, 1985). In contrast, folds to which the mylonitic fabric is axial planar in the Gamble Brook Formation deform only the bedding. Mylonitic overprinting of an earlier metamorphic fabric has also been observed in suitably oriented basement xenoliths within the younger granite gneisses (Fig. 3:2:3 in Cullen, 1984). Elsewhere, the S1 metamorphic fabric has been completely overprinted by S2a. Where visible, S1 is an amphibolite facies foliation defined by hornblende-plagioclase compositional banding in basement amphibolites and by quartzofeldspathic layering in the basement gneisses. D1 folds and linear fabrics have not been observed and the age and significance of the deformation is unknown. The correlation proposed by Donohoe and Cullen (1983) between the Great Village River Gneiss and the Mount Thom Complex would imply a minimum age for D1 of about 900 Ma based on available age data (Gaudette et al., 1984). However, given the uncertainties in this age and correlation, it is also possible that D1 records the earliest phase of the progressive deformation that later produced D2. If so, the existence of basement-cover relations between the Great Village River Gneiss and Gamble Brook Formation could no longer be argued on structural grounds although their obvious lithological contrasts would remain. However, until the results of planned geochronological studies become available, the balance of the evidence supports a significant age difference between the two units.

D2 Structures

D2 represents the principal tectonic episode throughout much of the Bass River Complex and resulted in two phases of deformation (D2a and D2b). These are equivalent to the DB2 and DB3 events of Donohoe and Wallace (1980, 1985) and Cullen (1984) but are considered to have formed during a single progressive deformation that records the development of a ductile shear zone along the basement-cover contact. The earlier (D2a) phase produced strong L-S fabrics (Fig. 2a) under greenschist to amphibolite facies conditions and is thought to have accompanied the emplacement of the syntectonic granite gneisses in the interval 625-640 Ma (Gaudette et al., 1984).

In the basement, S2a is the dominant planar fabric and ranges from an amphibolite facies metamorphic foliation to an intense mylonitic schistosity developed under strongly heterogeneous ductile shear. The associated L2a mineral lineation lies within S2a and, in outcrop, is defined by a strong dimensionally preferred orientation of hornblende and flattened quartzofeldspathic augen. Hence L2a is a stretching lineation and is believed to record the transport direction during D2a deformation. L2a plunges gently to moderately east and southeast. S2a dips broadly southeast at moderate to steep angles and is axial planar to small, tight to isoclinal folds (F2a) that deform the earlier S1 metamorphic foliation and can be locally shown to be sheath structures. F2a are typically asymmetric, similar folds and plunge to the northeast and southwest at moderate to gentle angles. Their sense of asymmetry throughout the exposed basement matches that defined by associated C-S fabrics and asymmetric augen but the implied sense of shear on L2a can be locally towards the west-northwest or the east-southeast due to younger, coaxial megascopic folding and the local development of second order folds on the flanks of mesoscopic F2a structures. However, within the ductile shear zone that defines the contact between the Great Village River Gneiss and the Gamble Brook Formation (Fig. 1), the sense of shear implied by all indicators consistently suggests oblique-slip, normal and sinistral movement towards the the present southeast (Fig. 2a).

In the granite gneisses, the earliest recognizable deformation is assigned to D2a since it takes the form of a heterogeneous, LS fabric that is coplanar, colinear and continuous with the D2a fabric of the host basement. F2a folds have not been observed in the granite gneisses, although stringers of the gneiss that are discordant to but contain the S2a fabric frequently define F2a folds within basement amphibolites. Elsewhere, however, dykes and veins of granite gneiss that cut the basement S2a fabric at high angles are themselves only mildly influenced by D2a deformation. Such relations suggest that the emplacement of the gneisses at around 630 Ma (Gaudette et al., 1984) occurred immediately prior to and during the development of the D2a fabric as concluded by Cullen (1984). Well-developed C-S fabrics and fractured, asymmetric quartzofeldspathic augen (the elongation of which defines an L2a stretching lineation) again imply L2a-parallel shear and within the ductile shear zone, where the granite gneisses intrude both the basement and the Gamble Brook Formation, are consistent with oblique-slip transport towards the present southeast (Fig. 2a).

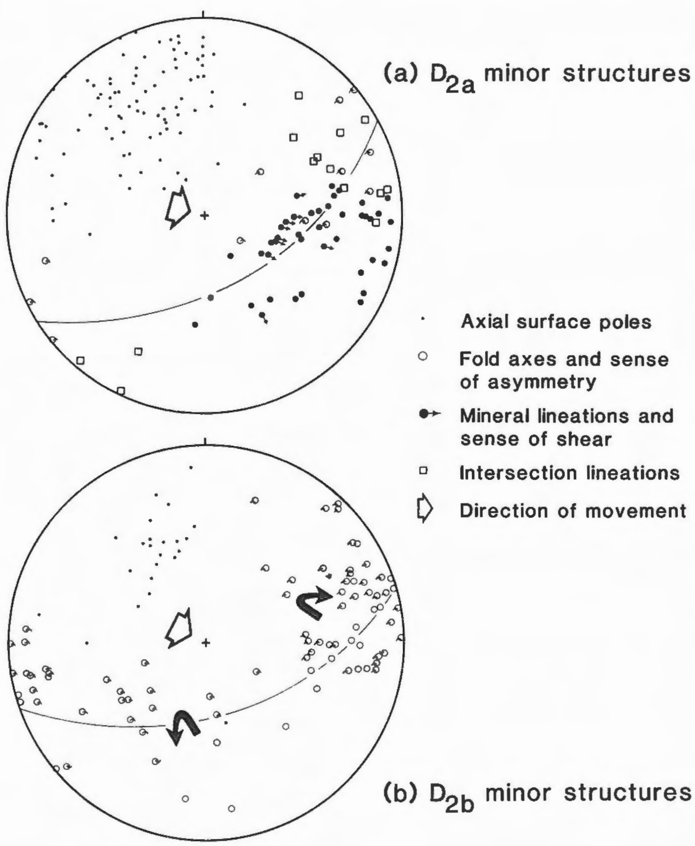

Figure 2. Synoptic equal area stereographic projections of D2 structural elements of the Great Village River Gneiss, syntectonic granite gneisses and the Gamble Brook Formation within the basement-cover ductile shear zone.

The first and principal phase of deformation within the Gamble Brook Formation is also assigned to D2a since the associated heterogeneous, LS tectonite fabric is coplanar, colinear and continuous with the D2a fabric in both the basement and intrusive granite gneisses. The D2a fabric in the Gamble Brook Formation intensifies towards the basement-cover ductile shear zone and developed largely under greenschist facies conditions. However, the absence of contact effects adjacent to the granite gneisses, and the presence of staurolite and sillimanite in the Portapique River Schist, implies a rapid increase in metamorphic grade to the high amphibolite facies as the basement is approached. In the basal Gamble Brook quartzites within the ductile shear zone, S2a is a locally intense mylonitic fabric defined by quartz ribbons, the elongation of which also defines a strong L2a-stretching lineation (Fig. 2a). However, the intensity of these mylonitic fabrics decreases with increasing distance from basement-cover contacts. In pelitic lithologies removed from the contacts, S2a is a muscovite-biotite or muscovite-chlorite schistosity. S2a dips south and east at moderate to steep angles and is axial planar to asymmetric F2a folds that appear to deform only the bedding (S0) and can be locally demonstrated to be sheath structures. Within the ductile shear zone, F2a folds plunge northeast and southwest at moderate to gentle angles parallel to a locally well-developed S0/S2a intersection lineation (Fig. 2a). These plunges steepen to almost vertical attitudes with increasing distance from basement-cover contacts. The L2a mineral lineation in mylonitic quartzites plunges east to southeast at gentle to moderate angles and, as in the basement, is considered to record the direction of D2a transport. The sense of shear on L2a implied by F2a fold asymmetries and mylonitic C-S fabrics is locally toward either the west-northwest or the east-southeast for the same reasons that exist within the basement. However, within the basement-cover ductile shear zone, all indicators of shear sense are once again consistent with oblique slip towards the present southeast with left-lateral and normal components (Fig. 2a).

D2b is considered to represent the second phase of the progressive (D2) deformation and resulted chiefly in the production of widespread asymmetric folds. Both inside and outside the basement-cover ductile shear zone, these F2b folds are coaxial, coplanar and spatially associated with those of F2a and, in like fashion, can be locally shown to be sheath structures. Yet they differ in structural style from F2a folds in that they fold S2a in addition to S0 and S1, and only locally develop axial planar fabrics (S2b). Associated linear features are restricted to local, L2b intersection lineations between S2b and S2a, S1 and S0. Within the ductile shear zone, F2b folds plunge at varying angles to the northeast, east, south and southeast and, stereographically (Fig. 2b), define a partial great circle girdle that is broadly coplanar with the mean orientations of S2a, F2b axial surfaces, and locally developed S2b axial planar crenulation cleavages. The sense of movement implied by the stereographic distribution of F2b fold asymmetry groups is parallel to that of D2a. Thus the kinematics and structural geometry of D2b mimic that of D2a and the two are considered phases of a single progressive deformation. Contact relations between the Great Village River Gneiss and the Gamble Brook Formation are consequently interpreted as the product of heterogeneous ductile shear involving repeated, oblique, top-to-the-southeast movement with left-lateral and normal components of slip.

D3 Structures

For reasons discussed below, polyphase deformation within much of the Folly River Formation appears to reflect a single progressive deformation (D3) that is younger than the D2 episode within the underlying Gamble Brook Formation. D3 is tentatively correlated with the thrust emplacement of the late Precambrian volcanics of the Jeffers Group which lies west of the Bass River Complex (Pe-Piper and Piper, 1987). A presently accepted minimum age for D3 is set by an Rb/Sr isochron of 596+/-70 Ma (Donohoe et al., 1986) obtained from the Debert River pluton which intrudes the Folly River Formation on Debert River (Fig. 1).

Penetrative LS tectonite fabrics produced during the earlier phase of D3 deformation (D3a) take the form of a low-grade schistosity (S3a) that dips broadly southeast at gentle to moderate angles; a mineral lineation (L3a) that plunges gently east-southeast and defines a stretching direction that is considered to record the line of D3a transport; and an S3a/S0 intersection lineation that lies parallel to the axes of associated F3a folds and plunges gently to moderately south and east (Fig. 3a). S3a is a greenschist facies foliation in the mafic volcanics and a bedding-subparallel, muscovite-chlorite cleavage in the overlying pelitic sediments. The L3a mineral lineation lies within S3a and is defined by the elongation of low-grade metamorphic minerals. S3a is axial planar to tight, asymmetric folds (F3a) which plunge southwest, south and east. These F3a folds are probably sheath structures although this has yet to be unequivocally demonstrated in the field. If correct, the sense of movement implied by the stereographic distribution of F3a asymmetry groups is everywhere consistent with northwest-directed shear parallel to the L3a mineral lineation. This vector, which implies oblique-slip movement with right-lateral and thrust components, is further supported by occasional C-S fabrics and asymmetric epidote augen in the metavolcanics.

Closely associated F3b folds differ in style from those of F3a in that they fold the S3a fabric and possess only a weakly developed, spaced axial planar fracture cleavage (S3b) and an occasional axis-parallel intersection lineation. However, they are broadly coaxial and coplanar with F3a folds and suggest a similar sense of shear (Fig. 3b). Hence they are tentatively attributed to a later phase (D3b) of the same tectonic episode.

Younger structures

Younger deformation locally affects the entire Bass River Complex and is likely of Paleozoic age. In fact, the regional form of the complex (Fig. 1) is dictated by a major synclinal structure which shows a right-lateral, en echelon orientation with respect to the bounding Cobequid and Rockland Brook faults, and may be as young as the Carboniferous. Other structures include occasional open, southerly plunging folds and conjugate sets of east-west kink bands, but these have not been studied systematically. Locally intense but spatially restricted penetrative fabrics associated with Paleozoic movement on major faults that border the Bass River Complex have not been examined but are described in detail by Cullen (1984).

SUMMARY AND INTERPRETATION

Three episodes and five generations of Precambrian folding have been provisionally identified in the Bass River Complex on the basis of style, overprinting relations and kinematic analysis. The first three generations (D1, D2a and D2b) correspond to those recognized and described by Donohoe and Wallace (1980, 1985) and Cullen (1984). However, as argued below, the last two generations (D3a and D3b) appear to affect only the upper portions of the complex and are tentatively correlated with the latest Precambrian deformation of the Jeffers Group (Pe-Piper and Piper, 1987).

The earliest episode (D1) represents an amphibolite facies event that affects only the basement lithologies of the Great Village River Gneiss and presumably developed prior to the deposition of the overlying platformal(?) succession now represented by the Gamble Brook Formation. The age of this episode is uncertain although correlation of basement in the Bass River and Mount Thom Complexes (Donohoe and Cullen, 1983), if supported by further age dating, would imply a minimum age of about 900 Ma (Gaudette et al., 1984).

The coplanar and coaxial D2a and D2b events are interpreted as phases of a single progressive deformation that developed in response to an episode of greenschist to amphibolite facies ductile shear with components of sinistral and normal slip. The timing of this event is presently set by the circa 625-640 Ma, Rb/Sr age of the syntectonic granite gneisses (Gaudette et al., 1984).

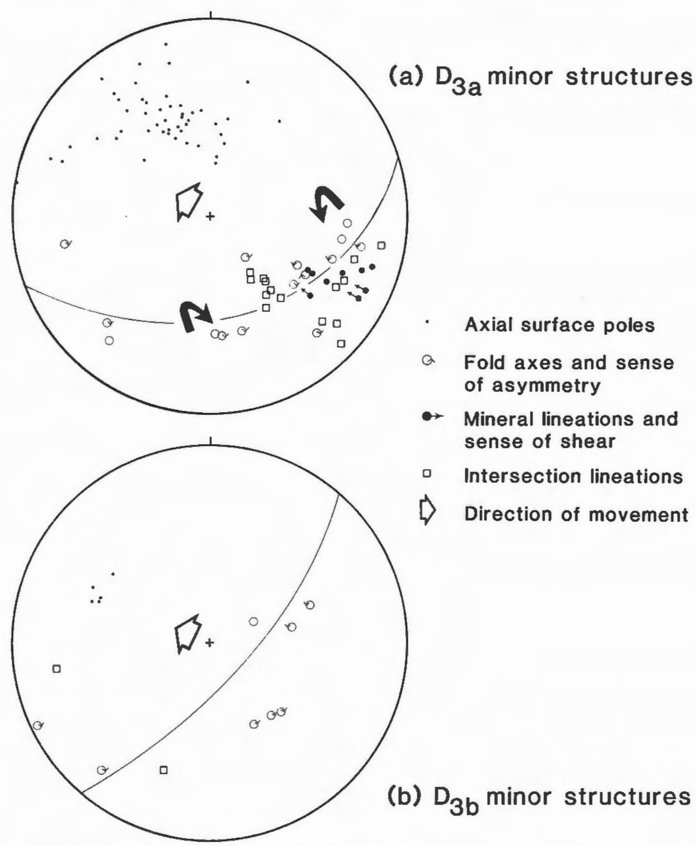

Figure 3. Synoptic equal area stereographic projections of D3 structural elements in the Folly River Formation.

D3a and D3b are provisionally interpreted as phases of a younger progressive deformation that developed under lower greenschist conditions during an episode of repeated, dextrally oblique, northwest-directed thrusting. A minimum age for this episode, which is kinematically similar to that which accompanied the thrust emplacement of the late Precambrian Jeffers Group, is presently set by the circa 600 Ma, Rb/Sr isochron from the post-tectonic Debert River pluton (Donohoe et al., 1986).

In the Great Village River Gneiss, D2 kinematics are consistent with basement remobilization under high amphibolite facies conditions. Associated, parautochthonous syntectonic granite gneisses imply that temperatures locally exceeded those required for crustal melting. At the same time D2 kinematics in the Gamble Brook Formation are consistent with significant extensional, left-lateral (transtensional) ductile shear along basement-cover contacts under largely greenschist facies conditions. Extensive foreshortening of the metamorphic gradient through structural telescoping within the extensional shear zone is unlikely due to the presence of parautochthonous granite gneiss in the Gamble Brook Formation. The geothermal gradient is therefore likely to have been steep. In fact, with the exception of a significant strike-slip component of motion, many of the features of the D2 tectonic episode (such as the development of a gneissic "core", a strong LS tectonite fabric; a steep geothermal gradient; a mylonitic "décollement" separating a gneissic infrastructure from a metasedimentary suprastructure; and polyphase folding during protracted deformation) broadly parallel those of Cordilleran, retroarc metamorphic core complexes (e.g. Davis, 1980). These similarities, together with the apparent syntectonic development of the Folly River volcanics with respect to D2 and the possible correlation of these volcanics with those of the Georgeville Group in the Antigonish Highlands, suggests a more regional significance to the development of D2. If correct, D2 kinematics would be consistent with a transtensional back-arc setting such as those proposed for other areas of the Avalon terrane during the late Precambrian (e.g. Nance, 1986). In fact, many of the features of the Folly River Formation (such as its uniformly mafic composition; the presence of a dyke complex, pillow lavas, hyaloclastites and jasperitic ironstones in the volcanic assemblage; and the close association of the volcanics with pelitic metasediments that resemble distal turbidites) strongly suggest an extensional tectonic environment. It is therefore tempting to attribute the Folly River Formation to an episode of local crustal extension that may have culminated the transtensional tectonics of D2. Such a tectonic setting may also be supported by the mafic dykes discussed below but has yet to be tested geochemically.

But for their sense of shear, structures in the Folly River Formation closely resemble those of the Gamble Brook Schist in form and orientation. As a result, Donohoe and Wallace (1980, 1985) and Cullen (1984) attributed deformation of the Folly River Formation to D2 and equated its principal fabric to that of the Gamble Brook Formation. However, a number of field observations suggest that the deformation of the Folly River Formation may have occurred during a younger, kinematically distinct episode of tectonic activity (D3), although D2 may have been penecontemporaneous with the formation's development. Thus mafic dykes that intrude both the basement and Gamble Brook Formation, and which may well be feeders to the Folly River Formation since they closely resemble those within the Folly River dyke complex, are themselves locally intruded by granite gneisses whose emplacement is thought to be syntectonic with respect to D2a (see above). The mafic dykes also increase in metamorphic grade to the amphibolite facies near basement-cover contacts in a similar fashion to the D2a metamorphic gradient of the Gamble Brook Formation/Portapique River Schist. However, although many of the dykes show a marginal D2a-parallel fabric, they are rarely intensely deformed and, in the basal quartzites of the Gamble Brook Formation, their relatively mild deformation contrasts strongly with the mylonitic fabric of their host. Those which contain sharply defined xenoliths of basement orthogneiss that bears a strong S2a mylonitic fabric but which themselves possess only a weak, coplanar penetrative fabric, suggest dyke emplacement prior to or during D2b. Yet other dykes sharply truncate F2b fold closures. Hence the emplacement of these lithologically similar dykes apparently spanned the interval of D2 deformation.

However, the intrusive contact of the Folly River dyke complex on East Folly River (Fig. 1) is discordant to and clearly postdates the mylonitic (S2a) fabric of the host Gamble Brook Formation since the fabric does not cross the contact and is oblique to a weaker planar fabric (assigned to S3a) that is present in the dyke complex. Thus the penetrative greenschist facies fabrics that are widely developed in the structurally and stratigraphically higher Folly River volcanics, which are believed to have been fed by these dykes, are presumably younger than D2a and differ in style from those of D2b (see above). Similarly, if the onset of Folly River magmatism was broadly syntectonic with respect to D2, a reasonable time lapse might be expected before syntectonic, greenschist facies metamorphism could occur in surface manifestations of this magmatism. More importantly, the syntectonic onset of Folly River magmatism would be consistent with the normal component of D2 shear in underlying units but conflicts with the consistent indication of repeated, oblique-slip thrusting suggested by our preliminary kinematic analysis of the Folly River Formation (Fig. 3). For similar reasons, the role of D2b in the deformation of the Folly River Formation is likely to be of limited extent. Neither F2b folds nor S2b crenulation cleavages have been observed in the mafic dykes, while the obvious style distinctions and apparent kinematic contrasts between S2b and S3a (see above) argue against a correlation between these fabrics. Instead, S3a appears to have resulted from a younger deformational event that affected only high structural levels of the Folly River Formation. If so, the age of this event is uncertain but must predate the 596 ± 70 Ma Debert River pluton as this contains xenoliths of foliated Folly River volcanics (Donohoe et al., 1986). Broadly parallel senses of repeated shear of similar age occur in the low-grade, arc-related volcanics of the late Precambrian Jeffers Group, based on our preliminary analysis of asymmetric folds and C-S fabrics. These fabrics are reported to predate the appinitic Jeffers Brook Diorite (Pe-Piper and Piper, 1987) which has yielded several recalculated K/Ar ages in the range 544-628 Ma (Donohoe and Wallace, 1985). It is therefore conceivable, although the two units are nowhere in contact, that D3 deformation in the upper portions of the Folly River Formation is associated with the emplacement of the Jeffers Group by northwest-directed thrusting during the latest Precambrian.

Taken together, these observations suggest that the extent to which D2 affects the Folly River Formation decreases with structural and stratigraphic height, while the reverse is true of D3. At the level of the basement-cover ductile shear zone, the initial intrusion of mafic dykes apparently occurred during and immediately prior to D2a, perhaps as part of a bimodal suite that included the penecontemporaneous granite gneisses. However, the overlying dyke complex is not strongly deformed and may in fact form a deformational buffer between structural levels of the Bass River Complex affected by the D2 ductile shear zone and those influenced only by D3. Hence the injection of mafic dykes appears to have spanned the interval of D2 transtension and could have culminated in the development of a dyke complex with continued extension.

In this scenario, the reversal of tectonic vectors implied by the D3 kinematics could be attributed to the transpressional closure of this short-lived basin, perhaps as the result of a collisional event associated with the emplacement of the late Precambrian Jeffers Group. Post-tectonic granites such as the Debert River pluton might then be interpreted as post-collisional intrusions.

If these interpretations are substantiated, the Folly River Formation is less closely linked to the Bass River Complex than was implied by Donohoe and Wallace (1980, 1985) and Cullen (1984), although mafic magmatism thought be associated with the formation is tied to the complex by the onset of D2a deformation. However, it must be emphasized that these tentative conclusions rest heavily on the proposed correlation of mafic dykes in the Gamble Brook Formation and the basement with those of the Folly River Formation. Confirmation of this correlation, which is presently based on their field relationships and strong lithologic similarity, must await the results of geochemical analysis.

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Summary of multidisciplinary geophysical research in the Charlevoix seismic zone, Québec

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Abstract

Various and extensive geophysical monitoring began in 1974 in the Charlevoix Seismic Zone. The historical record and a major earthquake in 1925 characterize this zone as the likely site of a future damaging earthquake (i.e., M = 5.5 or larger) and the aim of the monitoring was to develop the capability of predicting such earthquakes. The parameters studied included: microseismicity, seismic travel times, electrical impedance, vertical movement, horizontal movement, tilt, gravity change, and strain through water well level changes. Although no clear precursor was detected in the months before the largest seismic event (M = 5.0) that occurred during the observation period, the multiparameter experiment provided new insights into the structure and mechanics of this active region.

Résumé

Une surveillance géophysique à la fois variée et détaillée a commencé en 1974 dans la zone sismique de Charlevoix. Les relevés historiques et un séisme d'importance survenu en 1925 font de cette zone le site le plus probable d'un futur tremblement de terre susceptible de causer des dégâts (c.-à-d., M = 5,5 ou plus) et l'objet de la surveillance est de mettre au point des moyens de prédire de tels séismes. Les paramètres étudiés sont notamment les suivants : microsismicité, temps de parcours sismique, impédance électrique, mouvement vertical, mouvement horizontal, inclinaison, variations de la pesanteur et déformation à travers les variations de niveau dans les puits d'eau. Même si aucun précurseur net n'a été décelé dans les mois ayant précédé le plus grand événement sismique (M = 5,0) survenu pendant la période d'observation, l'expérience pluriparamétrique a apporté de nouvelles lumières concernant la structure et la mécanique de cette région active.

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INTRODUCTION

Many large earthquakes have occurred over the last four centuries in the Charlevoix Seismic Zone (CSZ) between Quebec City and Saguenay River, but details on the small scale seismicity and geographical extent were incomplete, mainly because of the large errors in the determination of the epicentres. The first field experiment conducted to determine the extent and rate of the microseismicity was undertaken in 1968 (Milne et al., 1970) albeit with very few instruments and essentially inconclusive results. A larger experiment undertaken in 1970 (Leblanc et al., 1973) located the CSZ boundary to the southwest and an even larger survey in 1974 (Leblanc and Buchbinder, 1977) found the boundary to the northeast. More recent measurements from a permanent seismometer network have established that the CSZ is 80 km long, 35 km wide, and parallel to the St. Lawrence River with a depth range of from 2 to 28 km.

In the early 1970s earthquake prediction was viewed internationally as an achievable goal; Charlevoix was a suitable target for prediction studies as the area is readily accessible and the location of any possible large earthquake is confined to a relatively small zone. Motivated by earthquake prediction research in other countries the Earth Physics Branch (now part of the Geological Survey of Canada) decided to test the dilatancy hypothesis and measurements related to a number of geophysical studies commenced in the CSZ in 1974. The Charlevoix Geophysical Observatory was established on the north shore of the St. Lawrence River for continuous recording of tilt, magnetotelluric and seismic observations. Precisely timed explosions also commenced to monitor possible changes in seismic travel times. A six-element telemetered seismic array was installed in 1976 to record more earthquakes at lower magnitude than was possible with only one relatively insensitive station in the zone. Three additional magnetotelluric stations were established

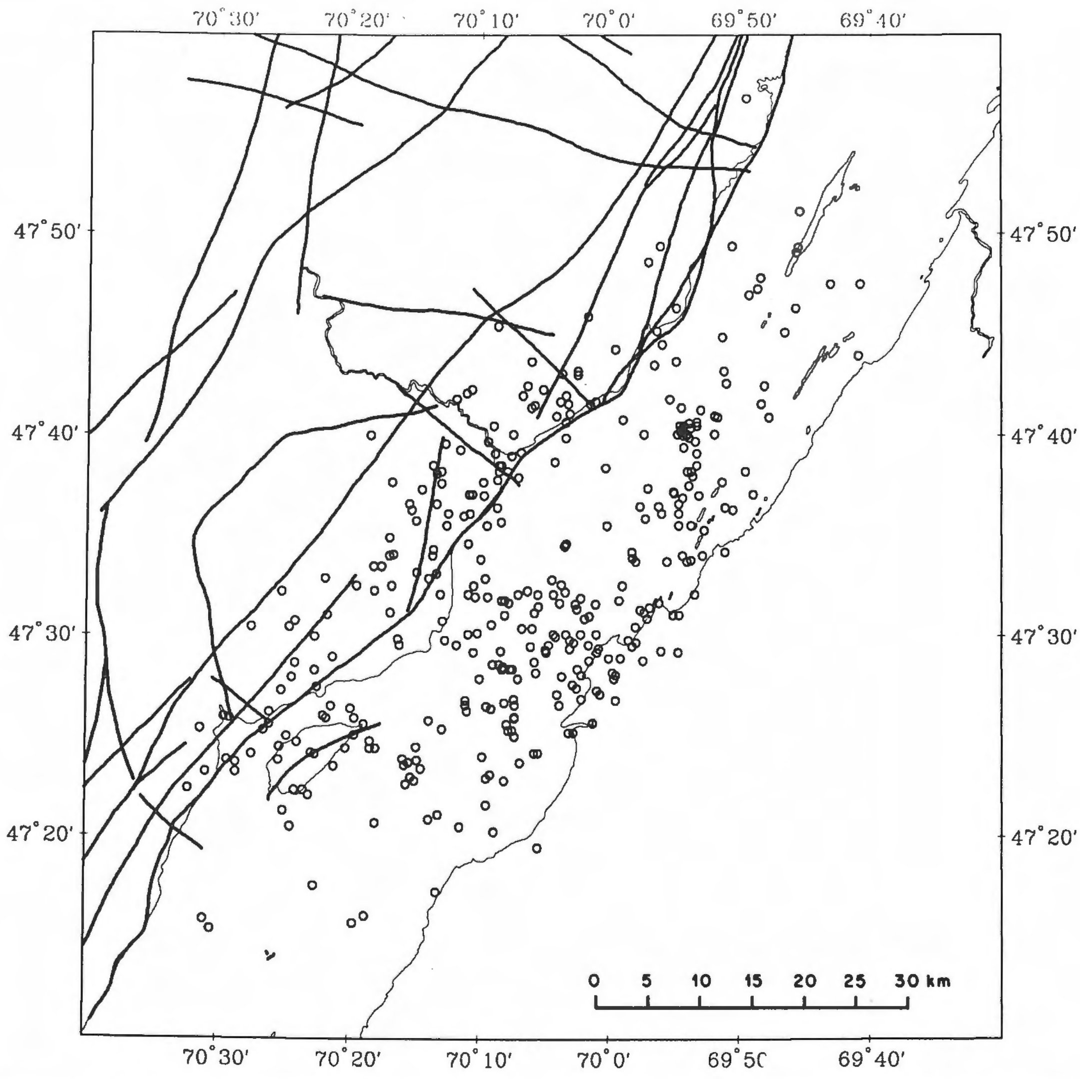

Figure 1. Epicentres (o) and faults from Rondot (1979). Epicentres are from array data where at least five stations and eight phases were used and where the location errors in latitude and longitude are equal to or less than 0.017 degrees and the depth error is less than 1 km. Data are from October 1977 to December 1986. Magnitudes range from 0 to 5 and depth from 2 to 28 km.

at about the same time. Various networks (levelling, gravity, horizontal control) for the study of crustal movement were established, older networks were resurveyed and water level monitoring in wells was added at the observatory in the late 1970s. All of this activity was spurred on by the large changes in geophysical parameters that had been reported in the literature (Whitcomb et al., 1973; Aggarwal et al., 1973).

This paper summarizes the twelve years of research that were undertaken by various researchers and institutions. Whilst a number of papers have already been published and that work will only be touched on here, much of the material is new. Knowing that changing concepts can bias the presentation of results, we first present the observations for each parameter separately with little or no attempt at synthesis or interpretation. Later, in the discussion section, a synthesis of the data is attempted together with an enumeration of the achievements of the project and recommendations for the future.

SEISMICITY

With the installation of the six-element short period vertical seismological array in October 1977, the hypocentres for the Charlevoix zone could be obtained for earthquakes with magnitudes as small as $M_L = 0.0$ on a routine basis (Fig. 1). The array geometry was set with three elements on the north shore and three on the south so as to bound the active zone. The data from this array were supplemented by the data from the regional analogue station LMQ on the north shore and the eastern Canadian telemetry network station LPQ on the south shore (Fig. 2).

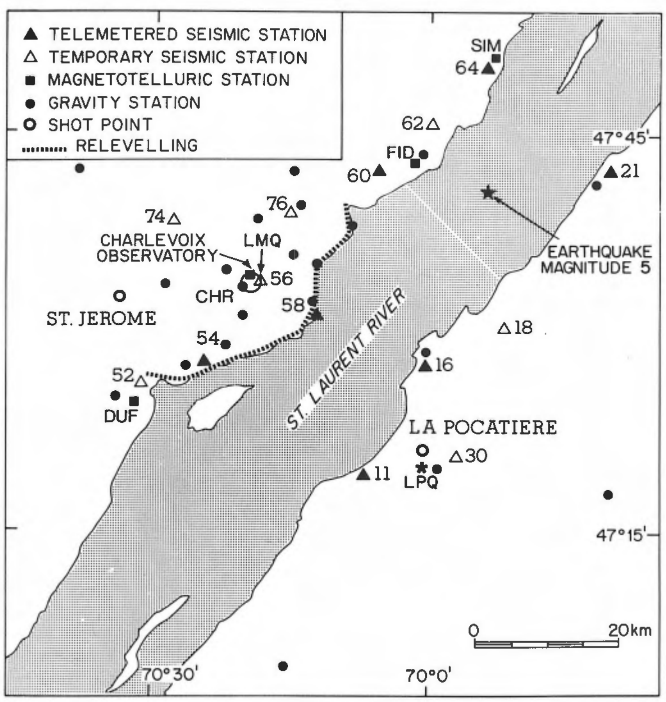

Figure 2. Geophysical recording sites in the Charlevoix Seismic Zone. Location of M 5 event 19 August 1979 is shown.

The data collected from the Charlevoix analogue seismic array have given a set of hypocentres that delineates many of the active subsurface faults. It is now well established that the active zone measures about 35 by 80 km and lies parallel to the St. Lawrence River. Based on these data, Basham et al. (1985) have reduced the size of the earthquake source zone and have hypothesized a maximum magnitude of 7.5 for it.

All of the microearthquake data collected support the view that the distribution of hypocentres is controlled by the ancient rift faults. Correlations between the hypocentre distribution and mapped surface faults on the north shore have been identified. A possible correlation also exists between hypocentres and the bathymetry in the St. Lawrence River to the southeast, suggesting that the bathymetry in this region may be fault controlled.

Temporal variations

During the 1974 microseismicity experiment (Leblanc and Buchbinder, 1977), it was noted that one station had a natural magnification of 0.8 magnitude unit over the average of the other seismograph stations. This sensitive site was chosen in October 1976 as the regional seismograph station LMQ and has operated to the present time. The number of very near earthquakes (within a radius of 50 km) detected for 30-day intervals is shown in Figure 3. The average for the whole histogram is 12 earthquakes per 30 days. For the first eight months the rate was well below the average with less than eight events per month. However, in the following three months that rate doubled and tripled. At the time this apparently significant increase in seismicity caused considerable concern since it was feared that it might herald a future large earthquake. The subsequent decrease in the rate alleviated that fear. The observations over the following years demonstrated that such excursions in the rate are not uncommon. If in the future the rate increases to significantly more than 30 events per month, this might indeed herald a future large event. However, such a rate would be readily detectable. In comparison to the numbers detected at this single station, the average rate of located events is six per month.

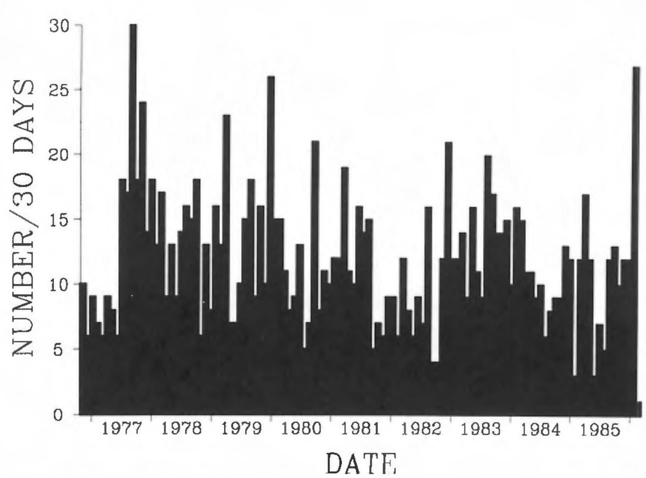

Figure 3. A 30 day histogram of earthquakes in the Charlevoix Seismic Zone recorded at station LMQ and having epicentral distances of less than 50 km.

P- AND S-WAVE EXPERIMENTS AND ANISOTROPY

P travel times from explosions

From 1974 to 1985 explosions were set off on both sides of the St. Lawrence River usually two times a year in order to monitor possible changes in seismic travel times. The experiment was designed so that the raypath crossed the seismically active area. The expectation was that, since changes in travel time of P-waves of 16 % and Vp/Vs decreases of 10 % had been reported in California (Whitcomb et al. 1973) and decreases in Vp/Vs of 13 % in New York (Aggarwal et al. 1973), changes of the order of 10 % might also be seen in the Charlevoix Seismic Zone. This was not the case; after a few years the largest change that was observed was an order of magnitude smaller, around a 1 % change in Vp. The travel time changes at two typical stations are shown in Figure 4 (Buchbinder and Keith, 1979; Buchbinder, 1981; Buchbinder et al., 1984).

S-waves from earthquakes

Crampin (1978) has suggested that dilatancy-induced anisotropy, resulting from parallel cracks, should manifest itself by the splitting of shear-waves which would be readily observable in polarization or particle motion diagrams. The splitting occurs because the plane shear-wave splits into two orthogonal components having different velocities. The hypothesis was successfully tested in Turkey (Crampin et al., 1980; Booth et al., 1985; Crampin and Booth, 1985). The hypothesis was also confirmed in the CSZ in 1984 (Buchbinder, 1985).

CRUSTAL DEFORMATION

Crustal deformation was monitored in the Charlevoix zone from 1976 to 1986 by three different types of tilt measurements at the Charlevoix Observatory and by periodic precise gravity and levelling surveys on the north shore. First-order horizontal control surveys across the St. Lawrence River both before and during the 1976-83 period provide an additional constraint on the regional deformation.

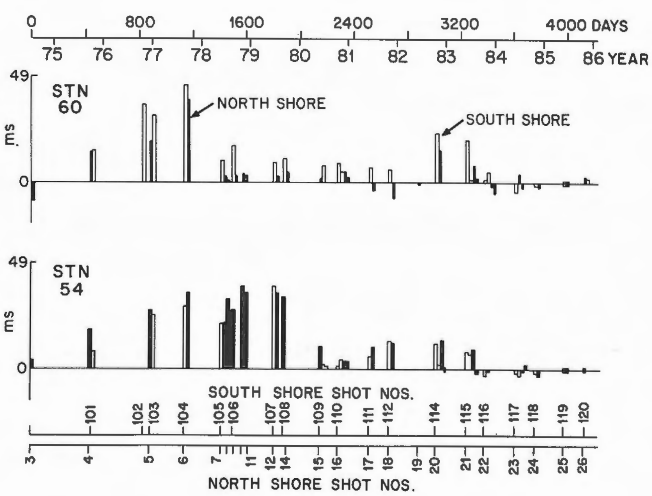

Figure 4. Seismic wave travel time residuals with respect to travel times measured in June 1985 as a function of time and shot number. Solid bars are for North shore shots, open bars for south shore shots.

Observatory measurements

Three mercury-level type tiltmeters were operated in a near-surface vault from 1976 to 1986. The vault tilt records were compared with the results from an adjacent 40 m-diameter bench mark array regularly levelled by Laval University from 1977 to 1986 and with results from borehole tiltmeters operated by Dalhousie University from 1983 to 1986 (Fig. 5). The vault data are characterized by large (amplitude 5-15 microradian) thermoelastic effects from annual surface temperature changes: the levelling array data are similarly affected (amplitude 1.8 microradian). All three methods of tilt measurement were affected by water level variations (determined in boreholes): the water-level effect varied from 1 microradian per metre at the surface to less than 0.1 microradian per metre at a tiltmeter depth of 110 m.

Regional crustal deformation

Regional deformation has been monitored in the CSZ from 1976 to the mid 1980s by means of precise gravity, levelling and horizontal control surveys. Semi-annual gravity surveys were carried out on the north shore of the River from 1976 to 1984 (Lambert and Liard, 1981). Levelling on the north shore was carried out by the Geodetic Survey of Canada in 1977, 1978, 1980 and 1982 and before the main monitoring period in 1965. Horizontal control surveys were conducted in 1965, 1978 and 1983. In 1983 the strategy was changed from a search for short-term earthquake precursors to a determination of long-term deformation and strain accumulation in the Charlevoix Seismic Zone. All three parameters were measured in 1983 and will be repeated in 1987. Taking the 1980 data as reference the spring-1977 elevation and gravity values along the north-shore appear to correspond (Fig. 6).

BOREHOLE TILT MONITORING

The Charlevoix borehole tiltmeter array represents an attempt to detect precursory tidal and secular tilt anomalies in the presence of near-surface meteorologically induced tilts. Theoretical studies by Beaumont and Berger (1974), Tanaka (1976) and Beaumont (1978) predict significant tidal tilt anomalies if there are precursory constitutive property changes in the rock mass surrounding earthquake source regions. The array (Fig. 5) consists of three boreholes lying in a triangle of approximately 80m each side. Results have been presented by Peters et al. (1983), and Peters and Beaumont (1985).

WELL-LEVEL MONITORING

Well-level monitoring began in the fall of 1978 when two 68 m (W1 and W2) and one 30 m deep (W3) vertical boreholes were drilled (Fig. 5). Current literature (Nur, 1972; Johnson et al., 1974; Scholz and Kranz, 1974) suggested that pore fluids play a major role in the mechanics of earthquakes and, inversely, are a sensitive indicator or diagnostic of this activity. The possible perturbing effect of local ground waters on the sensitive tilt measurements at the site was another factor which made it desirable to monitor pore pressure at depth.

Physical details of the boreholes are shown in Figure 5. The two identical observation boreholes W1 and W2 were drilled to provide evidence on the spatial coherence of the pore pressure measurements. The shallow hole, W3, was drilled to monitor variations in the water table overlying the partially confined monitored zones.

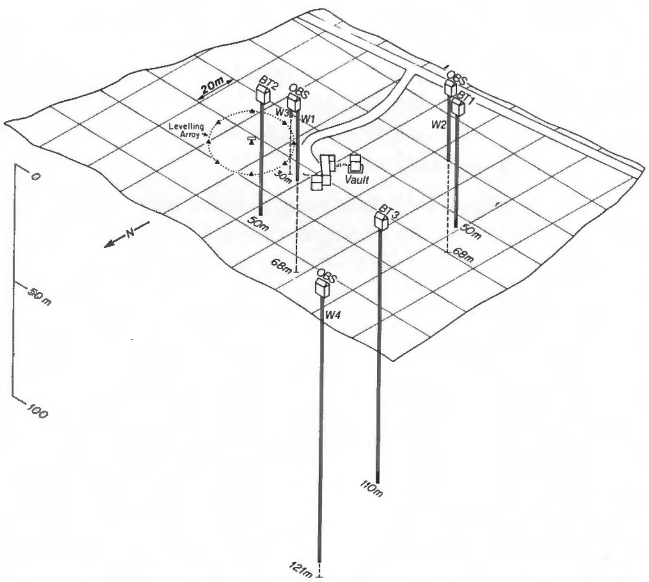

Figure 5. The Charlevoix Geodynamics Observatory near Charlevoix, Quebec. Water level monitoring was carried out in observation wells (OBS: W1, W2, W3). Cased and uncased sections of these wells are shown by solid and dashed lines respectively. Tilt was monitored in cased boreholes (at the bottom of BT1, BT2 and BT3), in a buried vault (vault) and by means of a leveling array (filled triangles).

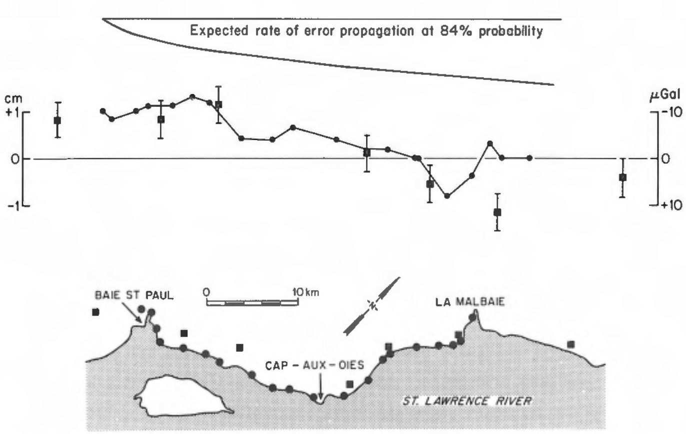

Figure 6. Comparison between gravity and elevation in spring 1977 with respect to autumn 1980. Elevation is positive upward on the graph; gravity is positive downward. Vertical bars on gravity values denote one standard error. Expected error propagation rate for the leveling is given in the top panel. The location of the leveling bench marks (solid dots) and the gravity stations (squares) is shown in the lower panel.

The meaningful interpretation of the water level data in terms of deformation related events has required borehole tests and the development of theoretical models to determine and remove 'normal' tidal, barometric and hydrological effects (Bower, 1986).

MAGNETOTELLURIC MONITORING

The bulk resistivity of crystalline rock is strongly dependent on the porosity of the rock and the degree to which cracks are connected. Electrical resistivity sounding techniques, such as the magnetotelluric (MT) method, may detect large scale changes in tectonic stress which modify the crack density and porosity. The first MT station, CHR, was established in October 1974, and since that time five additional stations have recorded data for various lengths of time (Fig. 2). Station BAT, which was 70 km northwest of CHR, served as a reference station since its location was remote from St. Lawrence River and the seismic zone. Stations FID and DUF were in a setting similar to CHR and were established to determine if the changes in electrical parameters observed at CHR were a local or more regional effect. Station LEB, which was only 200 m south of CHR, was intended to confirm the large changes observed at CHR. Station SIM was located at the northeast end of the seismic zone where many of the larger earthquakes have occurred in the past.

Electric impedance changes

Figure 7 shows the amplitude of the impedance measured in the major axis of anisotropy at a period of five minutes. The bars represent two standard deviations. The prominent time dependent changes that have occurred at most stations appear to be a local phenomenon when compared with the results from BAT. In fact, at BAT, there has been no significant variation other than a long-term increase of approximately 10 % over 6.5 years. The changes at FID and SIM appear to last for a number of months. However, significant

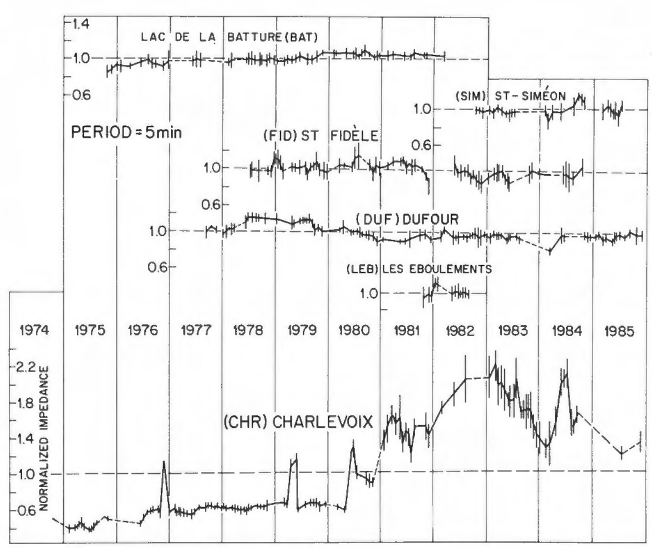

Figure 7. Normalized impedance measured in the direction of the major axis of anisotropy for a period of 5 minutes. Error bars are two standard deviations in length.

long term changes with periods of four or five years have occurred at DUF. The drop in impedance in 1979 was consistent with the disappearance of dilatancy in a significant part of the Charlevoix zone (Buchbinder et al., 1984).

The most striking changes are the large impedance increases that occurred at CHR in November 1976, April-May 1979 and from June 1980 until at least the end of 1985. There is no clear correlation between these events and seismic activity.

DISCUSSION

The Charlevoix experiment was conceived at a time when evidence was mounting that the phenomenon of dilatancy (Brace et al., 1966) may be a common feature in the crust prior to earthquakes (e.g., Semenov, 1969; Aggarwal et al., 1973; Whitcomb et al., 1973). The monitoring of various parameters at Charlevoix was designed to delineate changes associated with dilatancy preceding an earthquake of magnitude M = 5.5 or larger. It was recognized that smaller magnitude earthquakes would probably require more frequent observations because of the shorter associated "precursor" times. In fact, changes were seen in nearly all the monitored parameters, although the largest earthquake to occur over the monitoring period was an M = 5.0 event. Assessing the possible significance of these changes in terms of earthquake processes associated with the Charlevoix earthquakes has not been straightforward as a result of the inadequacies of existing theory and the variety of the signal sources.

"Tectonic" versus "local" signals

Most of the monitored parameters were observed with unprecedented accuracy in the presence of unknown nontectonic, usually local, effects. Instrumental accuracies were estimated in most cases through the internal consistency of results (e.g., gravity networks) and by laboratory or field tests (e.g., seismic wave travel times, water level monitors). The problem of separating local noise from tectonic signals was tackled mainly by estimating the effects of known meteorological inputs and by studying the coherence of signals from different sensors. For example, the predominance of meteorological effects was revealed in the case of near-surface tilt. A substantial effort went into the modelling of local, surface effects in order to improve the detectability of possible tectonic signals (e.g., seasonal hydrological effects on water levels, gravity, tilt, magnetotelluric parameters; variations in tidal loading parameters in the St. Lawrence estuary).

Evidence for dilatancy

The dilatancy model of earthquake preparation has been developed from a combination of laboratory and field results (Scholz et al., 1973; Anderson and Whitcomb, 1973; Brady, 1974; Stuart, 1974; Mogi, 1974; Mjachkin et al., 1975). No precursory effects of the kind predicted by the dilatancy model were detected before the M=5.0 event of 19 August, 1979, the largest seismic event encountered during the experiment. On the other hand, the shape of the coseismic well-level response to the M=5.0 and three other events at the Charlevoix observatory could be interpreted in terms of dilatancy collapse in spherical regions around the respective hypocentres. The radial strain within a dilatant zone 10 km in radius centred at the focus of the M=5.0 event is calculated to be 2.5×10^{-6} in order to produce the dilatation of 8.8×10^{-8} observed at the Charlevoix well. A dilatation of this magnitude is, however, one to two orders of magnitude larger than that allowed by elastic strain energy estimates for the M = 5.0 event assuming a seismic efficiency of 1 %. The same model predicts uplift, tilt and gravity changes too small to have been detected by the available observations. Model studies also show that if a dilatant zone were confined entirely under the river, the expected conductivity changes would be such that no effect would have been observed at magnetotelluric stations on the north shore. Another possible explanation for the coseismic well-level changes is triggering by local earthquakes of aseismic movement on a nearby active fault (e.g., Sacks et al., 1981).

A widespread drop in seismic P-wave travel times was detected in the Charlevoix seismic zone from 1977 to 1978 and again from 1979 to 1980 after the M = 5.0 event (Fig. 4). The magnitude of these changes (1 % or less) and the azimuthal dependence of the 1979 to 1980 changes (Kirsch et al., in press) suggest the phenomenon of extensive-dilatancy anisotropy (Crampin et al., 1984). According to this theory cracks grow throughout an earthquake preparation zone either by stress corrosion at the crack tips or by elastic bowing at stresses much lower than those required for conventional dilatancy. No differential change in gravity was seen at the time of this widespread drop in P-wave travel time but a change in apparent electrical impedance was seen at station DUF.

There is further evidence for deformation of microfractures that may not correspond to classical dilatancy nor to a distinct seismic event. A coincidence of changes in gravity, P-wave travel time changes and electrical impedance occurred during 1977 around seismic station 54 (Fig. 2) at the time of an unusual number of M=2 to 3 earthquakes. These changes could be explained by the flow of fluid into the region beneath station 54 along the NE-SW striking faults from the region northeast of seismic station 58. This flow coincided with anomalous behaviour in P-wave travel times at station 54 compared to other stations from August 1977 to May 1978. A rise in electrical impedance beginning at station DUF in late 1977 corresponded to an increase in electrical conductivity in the region beneath station 54 that may have been associated with the percolation of fluids from larger fissures to the surrounding microfractures.

Further insights and achievements

In general we have gained experience in making long-term, precise and accurate geophysical measurements and, in attempting to detect and interpret tectonic changes, we are beginning to appreciate the complexity of the region. In the travel time variation experiment from explosions, initially the errors were large and poorly known and may have been as large as ± 20 ms. With changes in software, recorder hardware, time keeping and the method of analysis, the errors were identified and were reduced to ± 4 ms.

Initially, with relatively little data in the interpretation of travel time changes from explosions, the theory of randomly oriented cracks (O'Connell and Budiansky, 1977) was used. Later, results were found to be more consistent with vertical, aligned, water-saturated cracks (Crampin, 1978). From the variations of travel times between 1979 and 1980 the azimuth of the cracks was determined to be about 35° in the Paleozoic (Kirsch et al., in press). The crack density $\in \doteq 0.03$ was determined from the observation of split shear waves from earthquakes in the Precambrian.

The information that can be obtained from these experiments with respect to crack orientation, ϕ, and density $\in$ may be summarized as follows. The absolute value of orientation, say ϕ, can be obtained from the S-waves from earthquakes, the value of $\phi$from P-wave velocity changes from explosions is indeterminate by 90°. For the crack density $\in$ an absolute value may be obtained from the S-waves, if an assumption is made about the path length involved. A change in $\in$ may be determined from P-waves velocity changes but the change is a function of an assumed starting value of $\in$ as mentioned above, from the shear-wave data values of $\in$ of about 0.02 to 0.03 and d = 30° were obtained (Buchbinder, 1985).

The utility of well-level monitoring is critically dependent on the degree with which barometric and meteorologically induced effects can be removed from the observations. Considerable success has been achieved in the Charlevoix study in modelling and removing these extraneous effects and anisotropy has been revealed which is consistent with the existence of the northeast-trending faults revealed by hypocentral locations (Anglin, 1984). In particular, the amplitude and phase of the well-level tide during the period 1981 through 1985 show large coherent variations about a general trend which changed from an isotropic elastic response in 1981 to a response consistent with predicted horizontal strain in an azimuth of 70° east of north in 1985.

Coseismic water well-level changes appear to be too large to be explained by the calculated residual strain field of local earthquakes. Calculations based on reasonable but poorly constrained parameter values, indicate that local volumetric strain due to immediate postseismic loss of dilatancy in the earthquake region may be the mechanism accounting for the observed well-level changes. Alternatively, triggered aseismic movement on a nearby fault could also account for these changes.

The techniques and instruments employed in the Charlevoix gravity network have achieved an unprecedented level of accuracy. The standard deviation of the station gravity values with respect to the mean of the network is about 3 μ Gal, equivalent to about 1 cm in vertical height. Only two gravity stations in the network show a reasonable correlation with ground moisture estimates derived from the precipitation record (Tanner and Lambert, in press). The 1977 gravity change may have been a delayed reaction to above average precipitation in 1976 but later maxima in precipitation did not produce the same effect.

The general lack of vertical crustal movement along the north-shore inside the Charlevoix Seismic Zone from 1965 to the present contrasts with the relative displacements of the order of 10 cm from the 1930s to the 1960s reported for the area peripheral and to the north and west of the zone (e.g. Frost and Lilly, 1966; Vanicek and Hamilton, 1972). If further horizontal control surveys across the CSZ restrict the long-term strain rate to still lower values, this would provide indirect support for the "uncompensated crustal loading" hypothesis of Dunbar and Garland (1975).

The inherent redundancy of the levelling array observations allowed the short wavelength vertical noise of the subsurface bedrock to be quantified which led to realistic estimates of long term stability of the tilt measurements. Records from the more stable borehole tiltmeters suggest a tilt rate of 0.1 microradian per year or less at the Charlevoix observatory from 1983 to 1986. No significant changes were seen in other parameters and there were no events larger than M = 4.0 in the Charlevoix zone during this period.

The magnetotelluric study demonstrates the sensitivity of electrical parameters (especially impedance and polarization azimuth of earth electric fields) to groundwater levels at CHR. The large increase in impedance at this station remains an enigma but may result from local changes in earth resistivity very near the electrode array (A.G. Jones, pers. comm., 1987).

Temporal changes in the number and distribution of located earthquakes in the Charlevoix zone were seen. A donut pattern of microseismic events has persisted for most of the period of monitoring. However, activity appeared to diminish during 1980 after the M = 5.0 event. This provides evidence that an event of this magnitude can influence the effective stress, possibly through the pore pressure, over a region 20 x 30 km in area - a much larger region than would normally be suggested for an event of this magnitude.

The horizontal control network across the Charlevoix seismic zone has seen considerable improvement in survey accuracy from 1965 to 1983 with the introduction of meteorological corrections for along-line refraction. However, taking into account current uncertainties existing data merely require the strain rate to be less than 0.2 microstrain per year, a rate typical of or a little larger than found in California, Japan and the West Coast of Canada (Lisowski, 1985).

CONCLUDING REMARKS

There is a strong possibility of a large, damaging earthquake in the Charlevoix Seismic Zone in the foreseeable future, but we cannot predict when it will occur. This is a similar state of affairs to the situation in other seismically active regions of the world where earthquake prediction research has fallen far short of its promises of a decade ago. In China, for example, there has not been a successful prediction of a major damaging earthquake since the 1976 Lungliu and Yen Yuan earthquakes. In the Soviet Union a prediction was made for a magnitude 7 earthquake in the Pamir (Simpson, 1979) in an uninhabited region. The tectonic environment also appears to play a role. In thrust regions in Soviet Central Asia, China and Japan the precursory effects appear to extend over wide areas (Simpson, 1980) whereas in strike-slip areas, such as the Andreas fault in California these appear to be lacking.

In Japan, because of the high rate of seismicity, only major events are being considered for prediction. The Tokai district in central Honshu Island (Mogi, 1986) is currently identified as having a high potential for a damaging earthquake in the medium term.

Therefore, in spite of a very large effort that has been expended during the past dozen years, after-the-fact 'predictions' are still more plentiful than before-the-fact ones, and in some regions despite being heavily instrumented, even this has not proved possible. Thus the Coalinga, California earthquake has produced no before the fact nor after the fact predictions.

In the Charlevoix Seismic Zone, as elsewhere, some progress has been made. With the present (1987) reduced data gathering program now underway, only the seismicity rate is being monitored closely enough to provide a short-term warning of a future large earthquake. Periodic gravity, levelling and horizontal control resurveys are aimed at quantifying the long-term strain accumulation.

When earthquake prediction was in its infancy it was hoped that research results from one geographic, tectonic or stress regime would be of help to other areas. It is now clear that there is more dissimilarity than similarity between regions, and that short term accurate prediction of earthquakes is going to be difficult in any region.

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Spatial relationship of mineral occurrences with geological and LANDSAT-derived lineaments, northeastern New Brunswick.

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Rencz, A.N. and Watson, G.P., Spatial relationship of mineral occurrences with geological and LANDSAT-derived lineaments, northeastern New Brunswick; in Current Research, Part B. Geological Survey of Canada, Paper 88-1B, p. 245-250, 1988.

Abstract

Analysis of the relationship between mineral deposits and lineaments derived from LANDSAT imagery and geological maps for an area in northern New Brunswick illustrates a spatial dependence. This is interpreted to reflect structural controls to mineralization in this part of the Canadian Appalachians.

Résumé

L'analyse de la relation entre des gisements et des linéaments minéraux dérivés d'une imagerie LANDSAT et de cartes géologiques pour une zone du nord du Nouveau-Brunswick illustre une dépendance spatiale. Les auteurs y voient un reflet des contrôles structuraux qui s'exercent sur la minéralisation dans cette partie des Appalaches canadiennes.

INTRODUCTION

Several studies have utilized remotely sensed imagery, including radar and LANDSAT, to detect geological lineaments. Generally the studies indicate that remotely sensed results compare closely with lineaments mapped by aerial photography.

Recently Bonham-Carter (1985) has demonstrated a quantitative technique to evaluate the spatial relationship between lineaments and mineral occurrrences. In several studies there was a demonstrable dependence between mineral occurrences and lineaments trending in a specific direction. For example in the Meguma Terrane of Nova Scotia, Bonham-Carter et al. (1985) showed a spatial dependence between the occurrence of gold deposits with LANDSAT-derived lineaments that trend in an ENE-NE direction and SEASAT radar-derived lineaments trending NNW-NW. Lineaments oriented in other directions did not prove to be significantly related to the spatial distribution of gold occurrences.

The objectives of the current paper are twofold. First, to test the ability of LANDSAT imagery in defining major structural and lithological elements of the regional geology, and to detect lineaments in a section of northern New Brunswick, an area of current mineral exploration activity. The orientation and density of the resulting lineaments are compared to geological features mapped by traditional ground-based methods. Secondly, to determine the significance of the spatial relationnship between lineaments (geological and LANDSAT) and mineral occurrences. For this purpose the locations of and commodity data for 62 precious and base-metal occurrences known in the study area were taken from CANMINDEX files at the Geological Survey of Canada GSC and recent mapping by New Brunswick provincial geologists (Philpott and Davies, 1985; Burton and Philpott, 1986; Philpott, 1987a, b).

GEOLOGICAL SETTING OF THE STUDY AREA

The study area, northwest of Bathurst, New Brunswick encompasses a region of 30 × 30 km (Fig.1). It is underlain in the south by polydeformed rocks of the Ordovician Tetagouche Group. Farther north, an Ordovician ophiolite complex, termed the Fournier Group (Rast and Stringer, 1980) is conformably overlain by turbidites of the Ordovician Elmtree Group. These rocks are unconformably overlain by less deformed sedimentary rocks of the Silurian Chaleurs Group in the central portion of the study area and by volcanic and sedimentary rocks of the Devonian Dalhousie Group to the northwest. The supracrustal rocks are intruded by two Devonian granitic intrusions, the Antinouri Lake and Nicholas Denys stocks, and numerous felsic and mafic dykes and sills (Fig. 2).

The structure of the region is dominated by the northeasterly trending Rocky Brook-Millstream Fault system. A cluster of epigenetic mesothermal to epithermal base and precious metal occurrences lies in an area between this fault system and the western margin of the Antinouri Lake granite stock. In addition to gold and silver, associated minerals include pyrite, arsenopyrite, galena, chalcopyrite, sphalerite and pyrrhotite.

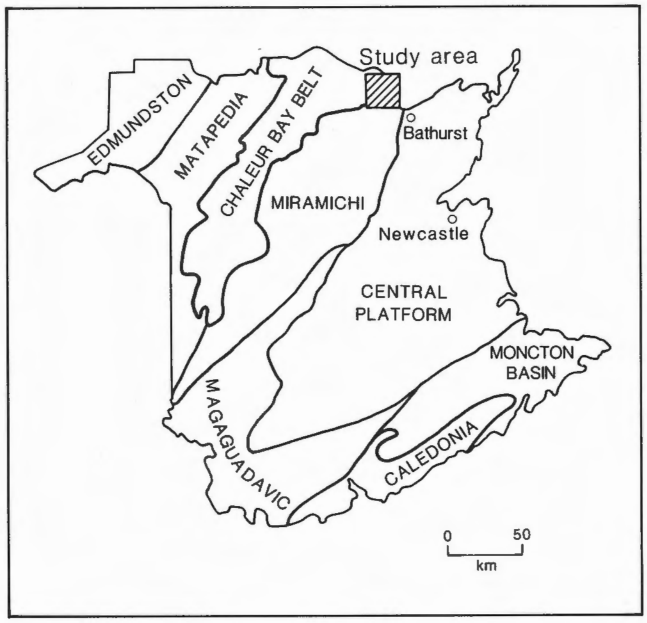

Figure 1. Location of study area with tectonic zones of New Brunswick.

METHODS

Lineament data

Two files of lineament data were constructed for this study: one set was derived from a LANDSAT Thematic Mapper (TM) image and the other from published geological maps For the purposes of this study a lineament derived from LANDSAT imagery is defined as any linear to curvilinear feature expressed tonally on remotely sensed imagery in a continuous or discontinuous manner (Harris, 1985). The LANDSAT lineaments were derived from a visual analysis of a colour enhanced TM image. The lineaments were then compiled onto a 1:50 000 map sheet and these lines were manually digitized creating a data file in vector format. The vectors were expressed in terms of UTM co-ordinates.

The geological lineaments were derived from 1:20 000 geological maps (Philpott, 1987a, b). These lineaments represent either mapped or inferred geological contacts and faults. These lines were also digitized.

Lineament mineral occurrence relationship

The method for evaluating the relationship between deposits and lineaments is described in Bonham-Carter (1985). In general the procedure tests the spatial dependence of mineral occurrences and lineaments by calculating a distribution which describes the frequency with which the observed mineral occurrences lie within a specified distance of the nearest lineament in a given direction. A Monte-Carlo simulation is then

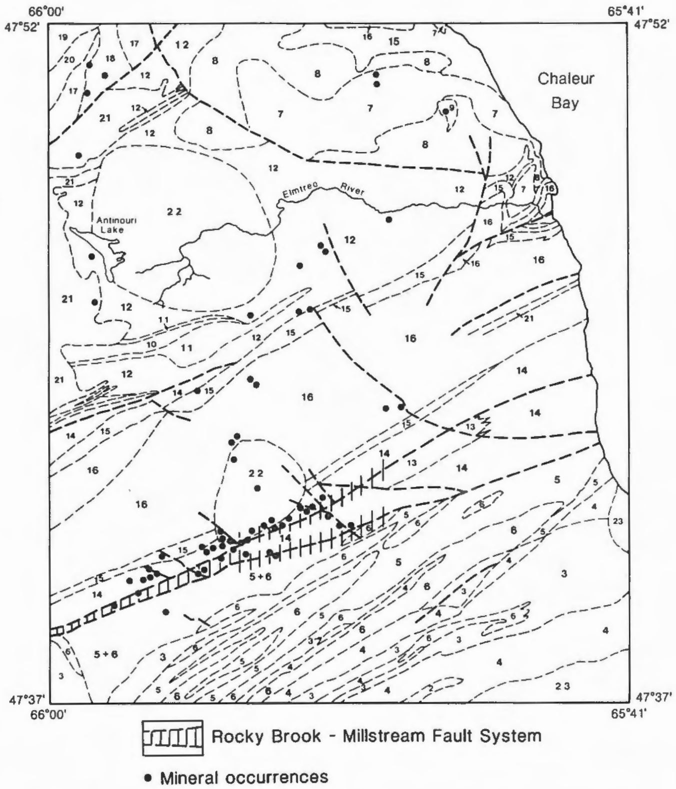

Figure 2. Geology of the study area (generalized from Philpott, 1987a, b).

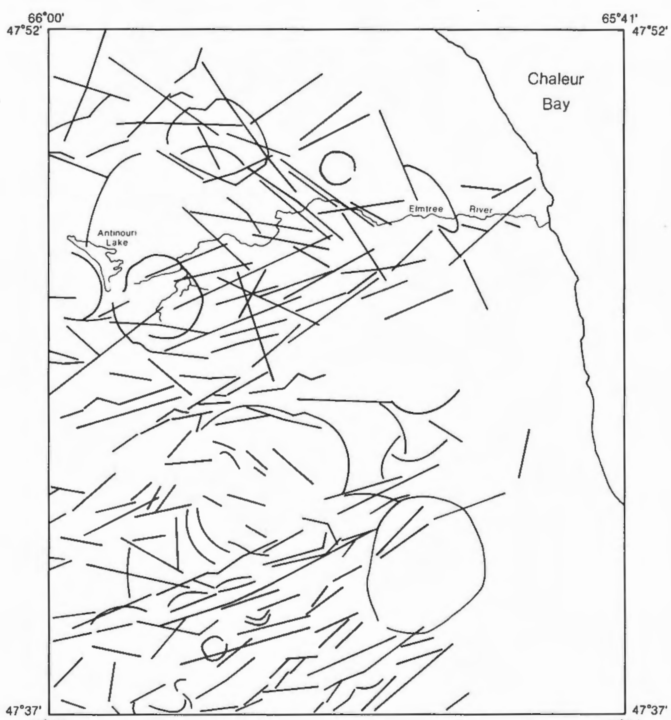

Figure 3. Lineaments derived from LANDSAT TM image of the area.

LEGEND

CARBONIFEROUS	
23	Reddish Brown Conglomerate, sandstone and siltstone (includes Bonaventure Formation).
LOWER DEVONIAN OR YOUNGER	
22	Felsic intrusive rocks:
21	Mafic intrusive rocks:
LOWER DEVONIAN	
DALHOUSIE GROUP	
20	Orange, massive and flow banded rhyolite and agglomerate
19	Green and brown Basalt, Basaltic tuff and Agglomerate, Red and Grey
18	Grey, green, finely laminated siltstone and mudstone grey calcareous sandstone
17	Dark Grey, Amygdaloidal Basalt
SILURIAN	
CHALEURS GROUP	
16	Greenish gray Slate, Phyllite Sandstone, and limestone, Hornfels and skarn within the thermal aureole of granitic stocks, (Includes Petit Rocher and Lavielle Formations)
15	Maroon and gray-green Conglomerate, grit, Greywacke and Argillite, minor interbeds of Limestone, Slate and Hornfels (Includes Armstrong Brook Formation)
14	Greywacke, Argillite, Slate, Minor Conglomerate, reddish Brown sandstone and hornfels
13	Greywacke, argillite, conglomerate, siliceous limestone and skarn
	ELMTREE ZONE
ORDOVICIAN OR OLDER	
ELMTREE GROUP	
12	Dark grey Phyllite, metagreywacke, grey quartzose metasandstone, minor conglomerate, grey feldspathic metagreywacke and impure limestone, hornfels and skarn within the aureole of granitic stock
11	Dark green porphyritic metabasalt, minor metagabbro
10	Red and maroon manganiferous slate, chert, grey chert, dark grey phyllite, buff metarhyolite, minor green metabasalt
FOURNIER GROUP	
9	Grey and reddish brown feldspathic metagreywacke, dark grey slate, grey chert, limestone and pillowed metabasalt
8	Dark green pillowed metabasalt
7	Dark green foliated metagabbro, amphiobolite, diabase dykes,
	MIRAMICHI ZONE
CAMBRIAN (?) TO LATE ORDOVICIAN	
TETAGOUCHE GROUP	
6	Mafic intrusive rocks:
5	Grey and Dark Grey Phyllite, minor quartzose metagreywacke, impure limestone, dolomitic limestone and graphitic schist
4	Red and Maroon manganiferous slate and argillite, grey and green chert, argillite, black slate, fedlspathic and lithic metagreywacke, calcareous slate
3	Massive and pillowed metabasalt, basalt metatuff and breccia
2	Quartz and quartz-feldspar metaporphyry, felsic metatuff, minor felsic agglomerate. interbedded with red, maroon and green manganiferous slate and chert
1	Quartzose metagreywacke, dark grey phyllite, slate and metaquartzite

used to generate a second set of random points (representing mineral occurrences) whose distribution is independent of the lineaments and the corresponding frequency distribution is also calculated. Whether the observed distribution indicates a spatial dependence between points and lines is determined by comparing observed and calculated distributions using a one-tailed Kolmogorov-Smirnov test. This statistic provides a measure of whether or not a hypothesis of point-line independence can be rejected at a given level of confidence.

RESULTS

General observations

The locations of the mineral occurrences and lineaments derived from the LANDSAT image and the geological map are presented in Figure 3. Comparison of the LANDSAT TM image with topographic and geological maps of the study area permits a number of general observations on the kinds of information which can be reliably interpreted from remotely sensed data.

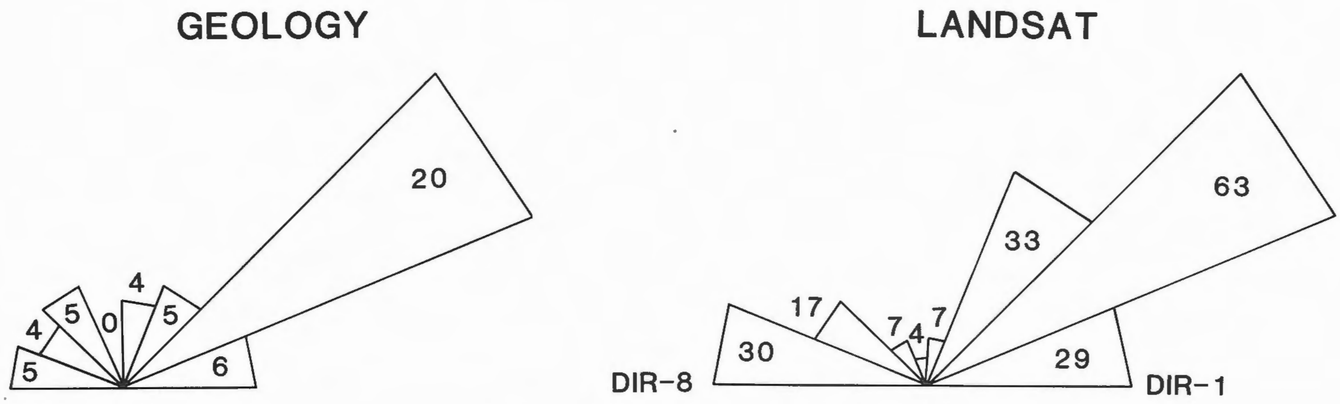

Figure 4. Rose diagram of orientation distribution for LANDSAT and geological lineaments.

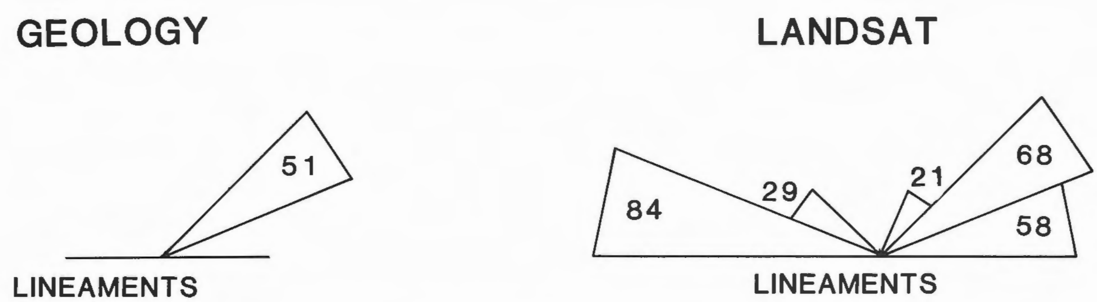

Figure 5. Rose diagram of orientation classes significantly related to mineral occurrences (numbers express degree of significance).

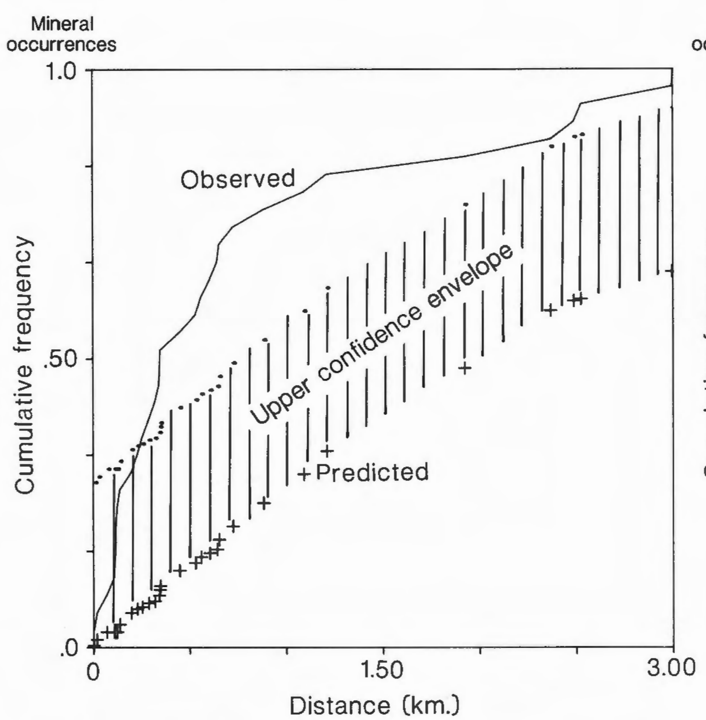

Figure 6. Distribution of deposit population (n = 62) versus distance from LANDSAT lineaments in DIR-2 (045 to 067.5°). The hypothesis of lineament/mineral occurrence independence is rejected when the curve of the observed points surpasses the 99% confidence envelope.

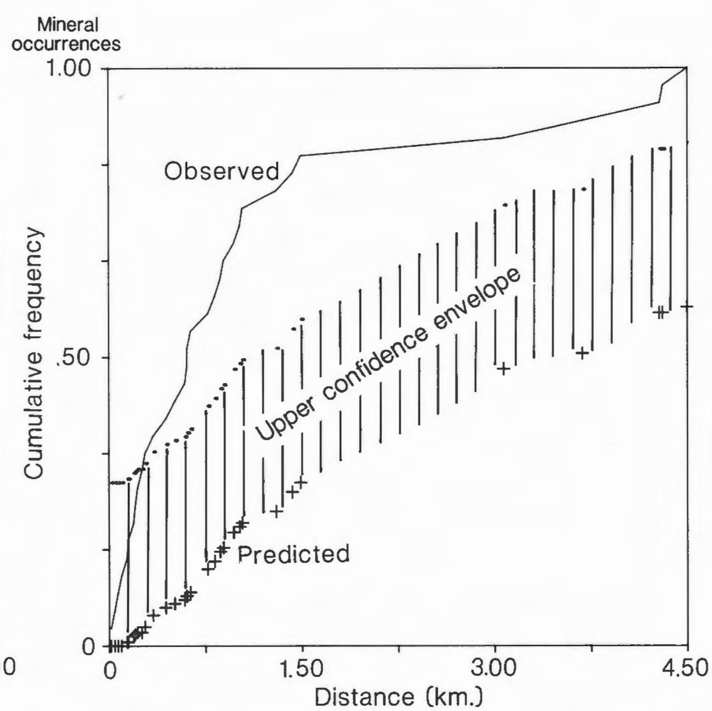

Figure 7. Distribution of deposit population (n = 62) versus distance from LANDSAT lineaments in DIR-8 (270 to 292.5°). The hypothesis of lineament/mineral occurrence independence is rejected when the curve of the observed points surpasses the 99% confidence envelope.

a) The courses of major rivers and large streams (i.e. Elmtree, Millstream and Tetagouche rivers) reflect underlying bedrock structures and orientations. Similarly, major faults such as the Rocky Brook-Millstream system are clearly marked by linear features which are in part expressed by drainage patterns.

b) Parts of the image area which are underlain by Silurian and Devonian rock formations have lineaments generally oriented northeast-southwest (045 to 065°), a direction which corresponds to the orientation of regional fold axes. These are often transected and offset by shorter lineaments trending northwest-southeast (225 to 245°), a direction which corresponds to the orientation of several quartz vein occurrences (ie Nigadoo Zn-Pb-Cu deposit).

c) Areas underlain by Tetagouche Group rocks have a more complex pattern of lineaments reflecting their polyphase deformation.

d) Devonian granitic stocks are clearly recognizable. For the Antinouri Lake stock, there is a distinct border zone which may reflect an alteration or contact metamorphic aureole extending into the country rocks.

d) There is an obvious clustering of mineral occurrences within and around the Rocky Brook-Millstream Fault system along the southern margin of the Nicholas Denys stock.

Lineaments

The lineaments were grouped into eight orientation classes at 22.5° intervals. In this study DIR-1 refers to lineaments in orientations from 067.5 to 090°, while DIR-8 refers to lineaments trending between 270 and 292.5°. Analysis of the LANDSAT image produced more lineaments in each of the eight classes that the geological analysis (Fig. 4). However there was a strong correlation between orientations in both data sets. In both data sets DIR-2 (045 to 067.5°) was the dominant direction class while there were very few lineaments detected in a northerly direction (Fig. 4).

A rose-diagram of the Kolmogorov-Smirnov statistic for each of the eight classes is shown in Figure 5. The statistic reflects the degree of spatial association between the mineral occurrence locations and lineament data such that the greater the number, the lower the probability that the association is a random event. The two data sets illustrate similar trends. For the geological data, results were significant at the 99 % confidence level for DIR-2. There were insufficient lineament data in the other directions to calculate meaningful relationships. The LANDSAT data showed similarly significant results with DIR-2 and additionnally for DIR-8 and DIR-1. In both sets of data there was no significant relationship when considering northerly trending lineaments.

The most significant cases are examined in more detail in Figures 6-8. These figures provide information on the distance at which the observed number of occurrences exceeds the upper confidence envelope and thereby signify the distance (from a occurrence to a lineament) at which the hypothesis of lineament/occurrence independence will be rejected. For example, in the LANDSAT relults for DIR-2 (Fig. 6) the two curves cross at approximately 170 m, a distance which encompasses 33 % of the observed occurrences. At this distance only 4 % of the total number of occurrences were predicted for a random situation. The other figures illustrate that the hypothesis of lineament/occurrence independence must be rejected at about 200m for LANDSAT DIR-8 (Fig. 7) and 100 m for geological data DIR-2 (Fig. 8).

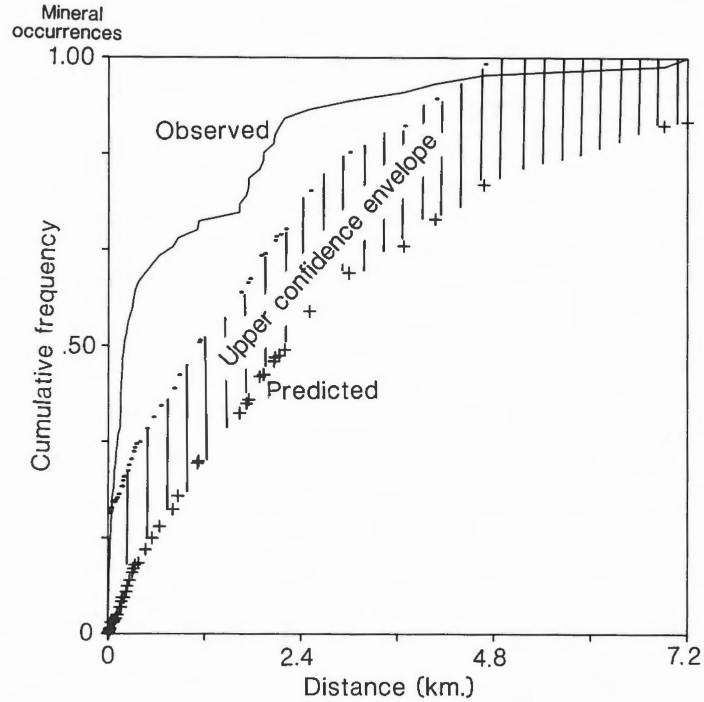

Figure 8. Distribution of deposit population (n = 62) versus distance from geological lineaments in DIR-2 (045 to 067.5°). The hypothesis of lineament/mineral occurrence independance is rejected when the curve of the observed points surpasses the 99% confidence envelope.

It should be noted that the calculation of the degree of spatial association between occurrences and lineaments is independent of the actual number of occurrences. For example, when considering the LANDSAT results, the direction interval of highest significance (DIR-8) contained the third highest number of lineaments.

CONCLUSIONS

1. LANDSAT TM data were effective in defining major structural and lithological elements of the regional geology.

2. The analysis of lineaments from LANDSAT produced more lineaments than existed on the geological map, however, the spatial orientation of the lineaments in both data sets was very similar.

3. A strong, non-random correlation exists between the location and orientation of known geological boundaries and linear elements derived from LANDSAT imagery and the location of known mineral occurrences. Of particular importance were those linear elements lying in three interval classes: DIR-1 (067.5 to 090°), DIR-2 (045 to 067.5°) and DIR-8 (270 to 292.5°).

4. The LANDSAT data were effective in illustrating the relationship for all three directions; while the geological results only showed a significant relationship with DIR-2. DIR-2 corresponds with the orientation of Acadian F1 fold axes whereas DIR-8 represents the orientation of late or post-tectonic extensional faults. The relationship of known mineral occurrences with lineament oriented in DIR-2 has been previously understood simply on the basis of their close physical proximity. This analysis has shown an equally important association of mineral occurrences and LANDSAT lineaments in two other directions. This clearly underscores the ability of LANDSAT data to provide an enhanced perspective of geological structure on a regional scale.

5. In all the cases the distance at which the lineament/occurrence relationship became significant was short (less than 200 m), illustrating that the majority of occurrences were situated on or very close to lineaments. The spatial association between lineaments with particular orientation and mineral occurrence locations emphasizes the significance of structural controls to mineralization for this portion of the Canadian Appalachians.

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The lithochemistry of metal-enriched coticules in the Goldenville-Halifax transition zone of the Meguma Group, Nova Scotia[1]

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Graves, M.C. and Zentilli, M., The lithochemistry of metal-enriched coticules in the Goldenville-Halifax transition zone of the Meguma Group, Nova Scotia; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 251-261, 1988.

Abstract

The transition between the sandy Goldenville and shaly Halifax formations of the Cambro-Ordovician Meguma Group appears to be a significant control for metal concentration. Manganiferous calcareous argillite and black slate at the base of the Halifax Formation are preferentially enriched in Mn, total C, Ba, Pb, Zn, Cu, Mo, W, and Au over average crustal values and over other lithologies of the Goldenville-Halifax transition zone.

The transition zone coticules are the product of manganese carbonate precipitation from pore fluid near the sediment-water interface during early diagenesis by oxidation of organic matter. Regional metamorphism has developed spessartine garnets at the expense of the carbonate. Metal enrichment appears to accompany sedimentary-diagenetic processes during anoxic conditions prevalent at the time of GHT development.

Résumé

La transition entre la formation sableuse de Goldenville et la formation schisteuse d'Halifax du groupe cambro-ordovicien de Meguma semble avoir grandement favorisé la concentration de métaux. L'argilite calcaire manganifère et l'ardoise noire à la base de la formation d'Halifax ont été enrichies de façon préférentielle en Mn, C total, Ba, Pb, Zn, Cu, Mo, W et Au, au-delà des moyennes crustales et d'autres lithologies de la zone de transition de Goldenville-Halifax.

Les coticules de la zone de transition sont le produit de la précipitation de carbonate à manganèse depuis le fluide interstitiel près de l'interface sédiment-eau, pendant la diagénèse initiale, par oxydation de la matière organique. Le métamorphisme régional a produit des grenats à spessartine au détriment du carbonate. L'enrichissement en métaux semble accompagner des processus sédimentaires-diagénétiques dans les conditions anoxiques qui ont prévalu à l'époque de la transition Goldenville-Halifax.

[1] Contribution to the Canada-Nova Scotia Mineral Development Agreement 1984-89. Project carried by Geological Survey of Canada, Mineral Resources Division, Project 700059.
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INTRODUCTION

The Cambro-Ordovician Meguma Group outcrops throughout southern Nova Scotia. It consists of at least 12 km of clastic metasedimentary rock varying from greenschist to amphibolite grade (Schenk and Lane, 1983; Schenk, 1983). The Meguma Group is made up of two major parts: a lower unit of thick-bedded metawacke interbedded with slate and an upper unit of slate interbedded with thin-bedded metasiltstone.

Analyses of the patterns of mineral resource distribution (Zentilli and Graves, 1977; Graves and Zentilli, 1982; Zentilli, et al., 1984), indicate that a disproportionate number of occurrences of gold, tungsten, arsenic, antimony, and zinc-lead occur within less than 10 % of the apparent stratigraphic thickness, specifically to the transition between the coarser grained Goldenville Formation and the overlying, pelitic, Halifax Formation, referred to here as Goldenville-Halifax transition or the GHT.

The GHT is marked by a finely laminated, manganiferous unit, locally rich in calcareous or calc-silicate nodules, spessartine quartzites (coticules of Keenan and Kennedy, 1983), and sulphides. This marker unit is overlain by a black, carbonaceous slate (Zentilli and MacInnis, 1983). Recent work on lead-zinc (Binney et al., 1986; Cameron, 1985), manganese (Hingston, 1985), tungsten (Shaw, 1982; Fisher, 1984), and tin occurrences (Wolfson, 1983) indicates that the GHT and its associated manganiferous enrichment constitute a significant metallogenetic control.

The characterization of the GHT should provide not only a guide to potential mineral occurrences in southern Nova Scotia but also insights into mineralizing processes in pelitic sequences elsewhere (Graves and Zentilli, 1976). An internally-consistent lithochemical database has been acquired that characterizes the chemical composition of stratigraphic and lithological units within the GHT. Definition of average or representative analyses of the major components of the Meguma Group might enable comparisons with other sedimentary sequences and possibly define depositional and tectonic environments for the Meguma Group.

STRATIGRAPHIC SETTING

LaHave River area lithologies

The stratigraphic subdivisions proposed by O'Brien (1986) have been partially adopted for this report as discussed in Waldron and Graves (1987):

- Meguma Group
  - Halifax Formation
    - Feltzen Member
    - Cunard Member
    - Moshers Island Member
  - Goldenville Formation
    - West Dublin Member/ Tancook Island Member
    - Rissers Beach Member
    - New Harbour Member

The New Harbour Member of the Goldenville Formation consists of thickly bedded, buff-weathering metawacke with thin interbeds of slate and slaty metawacke. The Rissers Beach Member consists of pervasively cross-laminated metawacke of variable thickness. The West Dublin Member in the LaHave River area and the largely equivalent Tancook Member in the Mahone Bay area to the east are similar to the New Harbour Member. The Moshers Island Member of the Halifax Formation consists of green to grey-green, calcareous, manganiferous, parallel-laminated argillite. The spessartine garnets which have formed at the expense of manganese and carbonate minerals during greenschist grade metamorphism give the rock a sandy feel and hardness. This mineralogy suppresses the slaty cleavage seen in less manganiferous rock of the same primary grain size. The Cunard Member consists of black slate interbedded with thinly bedded pyritiferous metasiltstone. The Feltzen Member consists of blue-grey slate interbedded with cross-laminated, often calcareously-cemented metasiltstone.

The thickness of these units is hard to estimate in the case of the uppermost and lowermost units as the base of the New Harbour Member is not seen and lower Paleozoic units overlying the Meguma Group are not seen in contact with the Feltzen Member. The upper units of the Goldenville (the West Dublin or Tancook members) range in thickness from 200 to 800 m (O'Brien, 1985a, 1986). The Moshers Island Member varies considerably in thickness: it is up to 300 m thick in the Dublin Shore area, 115 m in the Caribou core, and 10 m in the Eastville deposit. Metalliferous black slate at the base of the Cunard Member is difficult to measure due to rare or severely weathered outcrop and structurally disrupted drillcore, but is estimated to range in thickness from 10 to 30 m at Lake Charlotte and at Eastville.

The contact between the Goldenville and Halifax formations

The consistent placement of the contact between the Goldenville and Halifax formations is critical to the problem of the nature of the GHT as a geochemical marker as well as the practical use of the contact as an exploration guide. Care must be taken, however, in assessing the location of the contact which has been defined using two different criteria. Faribault (1914) and Taylor (1967, 1969), for example, used the highest exposed bed of greywacke as the position for the contact. Alternatively, sand/shale ratios have been used (Schenk, 1970); the coarser grained rocks being Goldenville and the finer grained Halifax, though the precise ratio chosen has usually not been stated. In areas of gradational lithological change, application of each of the two methods has resulted in considerable variation in the location of the contact. In the eastern portion of the outcrop area, the contact is generally sharp; whereas in the central and western part of the province, contacts are more gradational.

SAMPLING AND ANALYTICAL METHODS

The rocks of the GHT have been sampled at fourteen localities shown in Fig. 1, on transects selected on the basis of good stratigraphic and structural control, good exposure, and different metamorphic grade. The following localities have been sampled: (1) Sanford, (2) Cranberry Point, (3) Chebogue Point, (4) Broad Cove River, (5) Dublin Shore, (6) Tancook Island, (7) Lake Charlotte (in limited outcrop section in 1985 and in drillcore LC-86-1 in 1986), (8) Ship Harbour, (9) Liscomb Harbour, (10) Lundy, (11) Fogerty Head,

(12) Queensport Harbour, (13) Caribou Gold District (drillcore LL81-5A of Sherrit Gordon Mines Limited), and (14) Blockhouse (drillcore BH-9 of Golden Shadow Resources Limited). Outcrop sections have been briefly described previously (Zentilli et al., 1986). Only the sections at Dublin Shore (5), Caribou (12), and Lake Charlotte (13) are discussed here.

Dublin Shore samples (Fig. 2) are from nine sample stations. The New Harbour Member was sampled within 50 m of its contact with the Rissers Beach Member at the headland to the west of the village of Green Bay (location A, Fig. 2). The Rissers Beach Member was sampled at the landward end of Crescent Beach in the middle of the unit (location B, Fig. 2). The West Dublin Member was sampled on the east shore of Bush Island in the upper third of the unit (location C, Fig. 2) and at Sperry Cove. The Moshers Island Member was sampled within 20 m of the top of the unit on the south side of Bells Cove (location D, Fig. 2) at three separate but closely spaced sample stations within the top half of the unit. The Cunard Member of the Halifax Formation was sampled on the north side of Bells Cove within the bottom 50 m of the unit (location E, Fig. 2). All sample sections are considered to be representative of the lithological character of the respective units sampled.

Two N-diameter diamond drillholes in excess of 600 m in length were drilled by Sherrit-Gordon Mines Ltd in 1981 to the south of the Caribou Gold district in central Nova Scotia. One of these holes records a complete section through the basal Halifax Formation from a 20 m thickness of unambiguous Goldenville lithology, through 115 m of Moshers Island calcareous argillite, to 328 m of Cunard black slate and metasiltstone.

Nova Scotia Department of Mines and Energy drilled LC-86-1 at Lake Charlotte (point #7 on Fig. 1; 70 km east of Halifax) in support of this project. The Lake Charlotte manganese occurrence represents extensive mineralization in the Goldenville-Halifax transition zone and the N-diameter drillcore was designed to assess the stratigraphic position and character of the host rock. The 275 m hole provides a section of highly manganiferous Halifax Formation, both Moshers Island and Cunard lithologies, but is disrupted structurally and does not represent a complete stratigraphic section. The top 50 m of the drillcore is calcareous and manganiferous Moshers Island Member, the next 25 m is black, iron sulphide-rich Cunard Member slate, and the lower 200 m is again Moshers Island Member calcareous argillite.

The transition lithologies are also recognized in the stratabound Eastville zinc-lead prospect (10 km strike, 2 – 10 m thickness, and 1 to 3 % Pb+Zn) described by Binney et al., (1986), and studied in detail by MacInnis (1986). The Moshers Island Member as used in this report is recognized in the Eastville deposit as the calcareous quartzites of Binney et al., (1986).

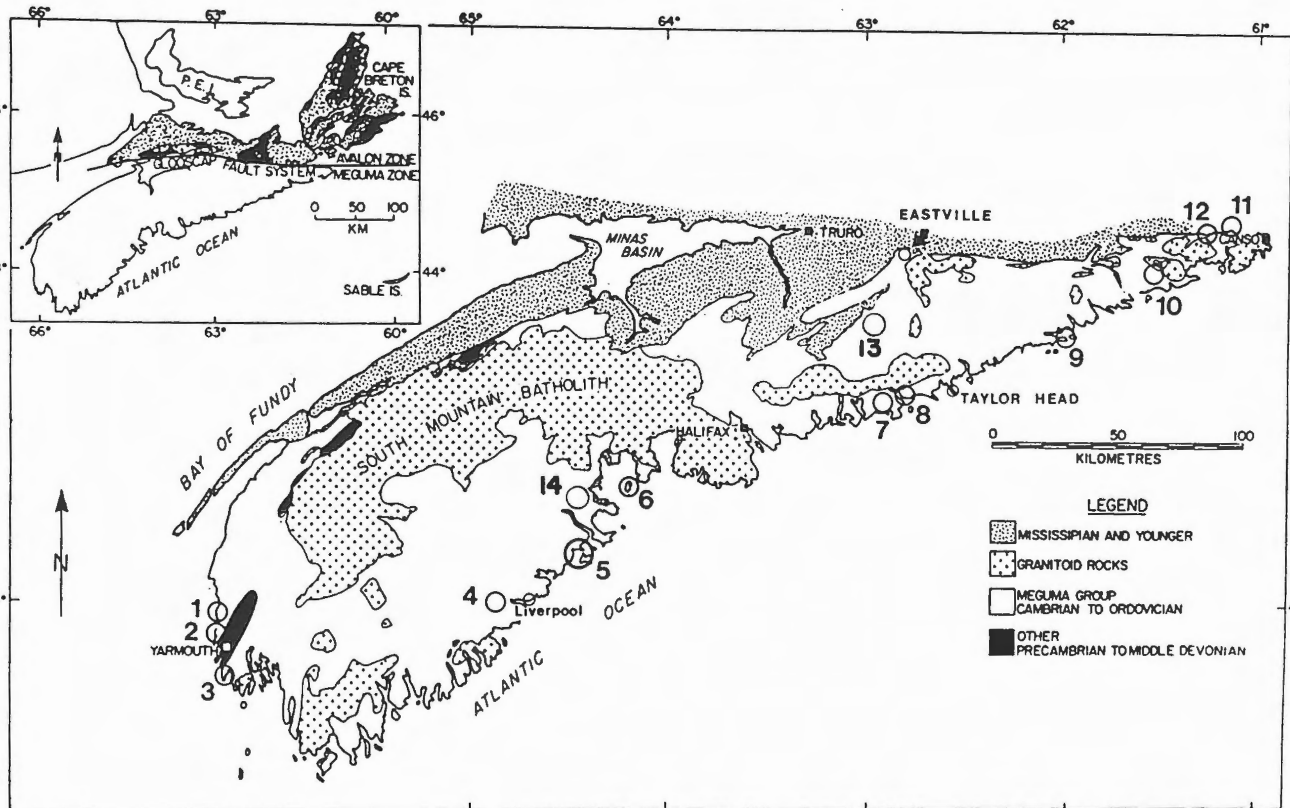

Figure 1. Location map of sample transects: (1) Sanford, (2) Cranberry Point, (3) Chebogue Point, (4) Broad River, (5) Dublin Shore, (6) Tancook Island, (7) Lake Charlotte (outcrop and borehole LC-86-1), (8) Ship Harbour, (9) Liscomb Harbour, (10) Lundy, (11) Fogerty Head, (12) Queensport, (13) Caribou (borehole LL81-5A), and (14) Blockhouse (borehole BH-9).

The Dublin Shore transect was sampled at stations consisting of 10 to 30 m of vertical stratigraphic section. Fresh rock was collected at each station bed-by-bed to obtain samples representing the lithological variation of that portion to the transect. Some attempt was made to sample for horizontal variation but sections rarely outcropped more than 50 m along strike. Ten to fifty samples were collected from each sample station. The spacing between sample stations was dictated by the quality and quantity of the outcrop and the variation within the transect.

All drillcore reported here is available for inspection at the Nova Scotia Department of Mines and Energy core storage facility in Stellarton, Nova Scotia. The drillcore was sampled at Stellarton and typical samples of each lithology were split by sawing.

Fresh samples were used for crushing, powdering, and analyses. Crushing and powdering for major element and trace element (by XRF) were done in tungsten carbide. Powders for neutron activation analyses were crushed and powdered by hand in an agate mortar.

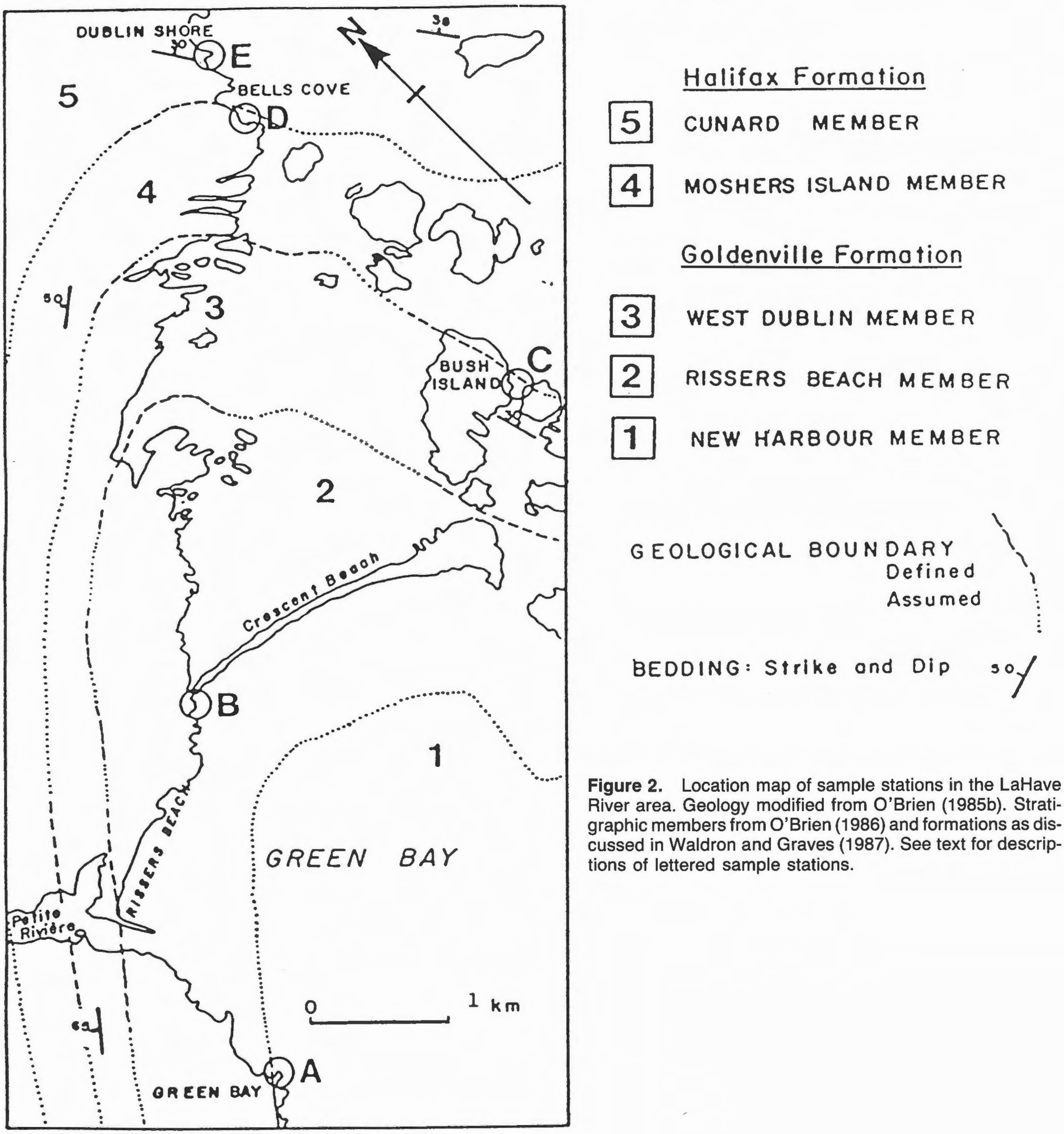

Figure 2. Location map of sample stations in the LaHave River area. Geology modified from O'Brien (1985b). Stratigraphic members from O'Brien (1986) and formations as discussed in Waldron and Graves (1987). See text for descriptions of lettered sample stations.

The samples were analyzed for major, minor, and selected trace elements including gold. Major elements (SiO_2, Al_2O_3, total Fe reported as Fe_2O_3, MgO, CaO, Na_2O, K_2O, and MnO) were analyzed using standard atomic absorption at Dalhousie University. P_2O_5 was determined using a colourometric technique also at Dalhousie University. Due to time limitations, some of the major element analyses were completed using fused glasses on the electron beam microprobe at Dalhousie University, following the method described by MacKay (1981). Most results were of similar reproducibility as atomic absorption analyses but the detection limit for MnO, CaO, Na_2O, K_2O, and P_2O_5 were considerably higher (see the following discussion of accuracy and precision). Trace element analyses (Ba, Rb, Sr, Y, Zr, Nb, Pb, Ga, Zn, Cu, Ni, TiO_2, V, Cr) were conducted by X-ray fluorescence at the XRF Regional Facility at St. Mary's University in Halifax, with good reproducibility and detection limits low enough to allow discrimination of crustal abundance levels for all elements in this package. Further trace element analyses (Sc, Co, As, Mo, Sb, Cs, La, Ce, Sm, Hf, Ta, W, Au, U, Th) were performed by non-destructive instrumental neutron activation analyses (INAA) by Bondar-Clegg and Company, Ltd., of Ottawa, Ontario.

Representative samples of Goldenville metawacke and Halifax slate remote from the GHT were analyzed for comparison. Every fourth sample was analyzed in duplicate, and internationally recognized rock standards were used for control of precision and accuracy.

PRELIMINARY DISCUSSION OF DATA.

Characterization

Table 1 summarizes results from three sections: the Dublin Shore area, Caribou, and Lake Charlotte. The geochemical data are presented for each lithology in the section mapped by O'Brien on the Dublin Shore of the LaHave River map area (O'Brien, 1985b).

The data show the similarity of the New Harbour, Rissers Beach, and West Dublin rocks of the Goldenville Formation. This similarity is emphasized by considering the relationship between SiO_2 and Al_2O_3 in these three units. The relationship is strongly linear and represents strong sedimentary fractionation of detrital mineral grains by size.

The Rissers Beach Member is geochemically intermediate between the sandstone and the shale of the surrounding two units. Pervasive crossbedding of the Rissers Beach sandstone indicates very rapid proximal deposition (Schenk and Adams, 1986; J. Waldron, unpublished manuscript). The Rissers Beach Member represents a mixture of the components which comprise the better sorted sandstone and shale of the enclosing units.

The Moshers Island Member is rich in manganese with values (volatile-free) ranging from 0.99 to 14.24 wt. % and averaging 4.49 wt. % oxide at Bells Cove on the Dublin Shore section; 0.26 to 12.09 % averaging 3.58 % in the Caribou core; and 1.78 to 23.55 % averaging 2.96 % in the lower part of the Lake Charlotte core and 11.17 % in the upper part as compared with values below 0.25 wt. % for the other units. The Moshers Island Member is high in Pb, Cu, Zn, As, Mo, W, and Au. Barium is also mildly anomalous when compared to adjoining lithologies.

A comparison of the mean trace element composition of the Moshers Island argillite Member with the average continental crust shows significantly higher values in the Moshers Island Member of As, Au, Pb, Ba, Mo, Sb, W, and U; and lower values of Na and Sr. When compared with average shale the low Na and Sr are still evident, and Rb, Sb, and Th are also low.

The ratio Zn/(Zn+Pb) shows a bimodal distribution, with a concentration between 0.8 and 0.9, and a more diffuse population less than 0.5. The samples in the group with lower ratios are crenulated metalliferous coticules with sulphide minerals and anomalous MnO (averaging 5.56 weight % for 11 samples). These "crusts" also have anomalous gold values (28 ppb average for 11 samples) and copper (270 ppm average).

The Cunard Member data again show a strong control by sedimentological mineral fractionation. The unit is high in Fe, S, and C compared to Goldenville pelitic beds. When compared to Cunard Member rocks in general (as represented by the 400 m section provided by the Caribou core), the silt-poor black slate at the base of the Cunard Member (as represented in the Lake Charlotte core) has high levels of Mn, S, Ba, Zn, V, Cr, Co, As, Mo, Sb, W, and Au and low values of Na, Ca, and Mg.

A suitable composite greywacke for trace metal comparison has not been compiled but the high alumina and the high Na_2O/K_2O of typical of greywacke characterizes Meguma Group wacke.

Discussion of metallogenetic problems in light of the data

Coticules are spessartine quartzites, first defined as such by Renard (1879), and were first recognized in Nova Scotia by Schiller and Taylor (1965). Coticules occur throughout the Moshers Island Member and intermittently in the stratigraphic units surrounding it. The origin of coticules is problematical. Their protolith has been suggested to be manganiferous sedimentary ironstone, manganese-rich sand, chert, siliceous sediment rich in montmorillonite clay, tuffaceous sediment (Fransolet and Kramm, 1983), or hydrothermal deposits. Spessartine quartzites have been recognized as genetically associated with stratabound sulphide deposits at Broken Hill, Australia (Stanton, 1976), and manganiferous haloes are known to surround many stratabound base metal deposits (Stumpfl, 1979).

The average composition of the manganiferous unit associated with the GHT for the La Have River area (Moshers Island Member) is given in Table 1, column 7. This is considered representative of most localities. It is fruitful to compare the GHT manganiferous rocks with marine iron manganese deposits of different oceanic environments. Figure 3 is a ternary Mn-Fe-(Cu-Ni-Co) diagram showing the distribution of these elements in various Fe-Mn rich deposits, after Bonatti (1975). Hydrothermal manganiferous sediments

Table 1. Lithochemical summary data Goldenville-Halifax transition

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
SiO_2	74.40	62.02	70.99	76.66	58.12	78.09	58.36	59.88	46.77	59.32	57.41	60.85	62.44	73.31	70.95	60.28	63.34	84.23
Al_2O_3	13.02	19.66	15.15	12.44	21.90	11.48	18.69	18.40	16.70	18.36	18.18	23.64	22.92	14.16	12.98	15.61	16.39	5.05
Fe_2O_3*	4.91	7.38	6.05	4.15	9.51	3.68	10.40	9.32	13.44	9.94	11.78	6.51	5.43	6.15	6.80	7.27	7.32	1.50
MgO	1.36	2.32	1.83	1.29	2.93	1.03	2.17	2.43	2.79	2.76	1.80	1.81	2.49	2.30	1.66	3.52	3.75	1.24
CaO	0.44	0.40	0.69	0.70	0.26	0.84	1.08	1.17	5.12	1.73	1.76	0.09	0.07	0.07	3.12	5.16	3.35	5.85
Na_2O	3.04	2.38	2.72	2.53	2.04	1.52	0.69	0.95	0.24	0.84	0.63	1.30	1.61	1.10	0.99	3.87	1.40	0.47
K_2O	2.06	4.57	1.62	1.45	4.05	2.59	2.70	3.65	2.32	2.92	3.51	4.22	3.44	1.88	2.11	3.17	3.47	1.38
TiO_2	0.61	0.94	0.63	0.53	0.84	0.56	1.08	0.92	1.10	1.04	1.05	1.24	1.25	0.77	0.68	0.74	0.84	0.27
MnO	0.06	0.09	0.20	0.13	0.24	0.15	4.63	3.17	11.03	2.93	3.68	0.27	0.25	0.22	0.64	0.12	0.12	0.01
P_2O_5	0.10	0.23	0.11	0.11	0.09	0.06	0.19	0.12	0.50	0.16	0.20	0.06	0.10	0.03	0.07	0.24	0.02	.00
total	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100
S	0.01	0.01	0.01	0.01	0.01	0.29	0.13	0.54	1.66	0.20	4.61	1.27	0.04	2.96	1.81	0.03	0.24	0.02
CO_2	0.21	0.05	0.04	0.10	0.03	0.63	0.21	2.04	7.36	2.20	2.61	0.04	0.03	0.00	2.80			
C	0.02	0.01	0.02	0.01	0.01	0.15	0.07	0.18	0.05	0.30	5.32	1.12	0.65	0.11	0.74			
Ba	507	1241	336	269	835	463	1260	1081	1024	897	1009	1205	871	499	603	425	580	50
Rb	74	154	71	56	148	92	98	127	83	108	129	160	135	71	80	90	140	60
Sr	198	181	218	148	108	220	76	115	113	188	205	385	257	104	171	375	300	20
Y	21	33	31	25	25	20	37	33	36	31	42	46	34	30	26	33	26	15
Zr	211	223	218	217	185	208	173	154	145	166	154	186	162	168	201	165	160	220
Nb	12	18	15	12	18	11	17	13	12	15	14	16	19	13	11	20	11	1
Pb	14	18	21	19	16	10	81	28	10	22	21	12	20	25	13	13	20	7
Ga	17	28	17	14	27	14	25	24	25	22	23	27	28	20	17	15	19	12
Zn	61	94	82	57	121	56	102	114	137	115	112	96	85	134	83	70	95	16
Cu	16	27	15	9	15	18	54	45	94	33	69	23	3	45	32	55	45	5
Ni	21	43	29	18	46	17	51	54	103	62	57	24	12	25	21	75	68	2
V	82	147	91	64	139	65	199	144	171	147	177	174	183	96	79	135	130	20
Cr	69	120	90	65	116	69	119	112	109	111	113	160	138	67	68	100	90	35
Sc						8.8	24.2	16.8	19.2	19.1	18.0	24.6	22.8	11.5	10.8	22.0	13.0	1.0
Cr	54	103	77	58	89	87	101	88	92	97	106	150	120	59	58	100	90	35
Co						13	34	31	88	58	78	25	11	82	27	25	19	1
As	13.7	23.4	18.1	12.0	20.3	9.0	134.9	58.8	93.0	48.9	18.9	5.3		42.0	52.2	2.0	13.0	1.0
Mo	1.0		2.5	4.3			16.8	20.3	30.5	2.1	35.4	10.0	4.5	3.0	3.0	1.5	2.6	0.2
Sb	0.3	0.5	0.3	0.3	0.2	1.1	0.7	1.4	1.2	0.8	2.9	1.9	0.8	4.3	0.9	0.2	1.5	0.1
Cs						5.0	5.7	6.0	3.6	5.2	6.5	11.6	9.8	5.5	5.3	3.0	5.0	0.5
La	30	34	39	37	20	36	38	47	45	42	47	70	11	5	22	30	24	16
Ce						71	48	96	109	86	95	129	20	9	47	60	50	30
Sm						5.0	4.1	7.0	6.8	6.2	6.8	8.9	2.0	1.0	4.1	6.0	5.8	3.7
Ta						1.0	1.6	1.0	1.0	1.3	1.3	1.6	1.4	1.7	1.2	2.0	0.8	0.1
W						5.0	4.1	4.1	7.0	5.7	6.6	3.3	2.0	2.0	2.5	1.5	1.8	1.6
Au	2.0	1.0	1.5	1.5	5.7	1.0	8.3	4.0	4.2	3.2	4.2	2.9	1.5	3.5	1.9			
Th	7.3	11.8	8.7	7.4	12.3	9.0	10.6	9.9	8.2	10.0	9.9	14.4	11.0	7.3	8.7	7.2	12.0	1.7
U	1.3	2.6	1.9	1.5	2.6	2.1	4.2	2.8	3.5	2.4	4.9	6.0	3.5	2.7	2.5	1.8	3.7	0.5
n	25	9	15	30	3	2	43	11	9	20	7	8	2	2	9			

Legend

1 = New Harbour Mbr wacke, Goldenville Fm, Green Bay, Lunenburg Co, NS
2 = New Harbour Mbr slate, Goldenville Fm, Green Bay, Lunenburg Co, NS
3 = Rissers Beach Mbr, Goldenville Fm, Crescent Beach, Lunenburg Co, NS
4 = West Dublin Mbr wacke, Goldenville Fm, Bush Island and Sperry Cove, Lunenburg Co, NS
5 = West Dublin Mbr slate, Goldenville Fm, Bush Island and Sperry Cove, Lunenburg Co, NS
6 = Goldenville Fm wacke, Caribou drillcore LL81-5A, Halifax Co, NS
7 = Moshers Island Mbr, Bells Cove, Lunenburg Co, NS
8 = Moshers Island Mbr, Caribou drillcore LL81-5A, Halifax Co, NS
9 = Moshers Island Mbr, upper portion Lake Charlotte drillcore LC-86-1, Halifax Co, NS
10 = Moshers Island Mbr, lower portion Lake Charlotte drillcore LC-86-1, Halifax Co, NS
11 = Cunard Mbr slate, Lake Charlotte drillcore LC-86-1, Halifax Co, NS
12 = Cunard Mbr slate, Caribou drillcore LL81-5A, Halifax Co, NS
13 = Cunard Mbr slate, Bells Cove, Lunenburg Co, NS
14 = Cunard Mbr wacke, Bells Cove, Lunenburg Co, NS
15 = Cunard Mbr wacke, Caribou drillcore LL81-5A, Halifax Co, NS
16 = average crust[1]
17 = average slate[1]
18 = average sandstone[1]

Major Elements: in weight per cent recalculated volatile-free
* total Fe as Fe_2O_3
all analyses by standard atomic absorption analyses except P_2O_5 (colorimetric analyses); FeO, S, CO_2, and C (wet chem titration) at Dalhousie University, Halifax, Nova Scotia

Trace Elements: in parts per million (ppm)
all analyses by X-ray fluorescence, Regional XRF Facility, St. Mary's University, Halifax, Nova Scotia

Trace Metals: in ppm except gold (ppb)
all analyses by instrumental neutron activation (INAA) by Bondar Clegg and Company, Ltd, Ottawa, Ontario

1. Mason, B and Moore, B
1982: Principles of Geochemistry, 4th edition; John Wiley & Sons.

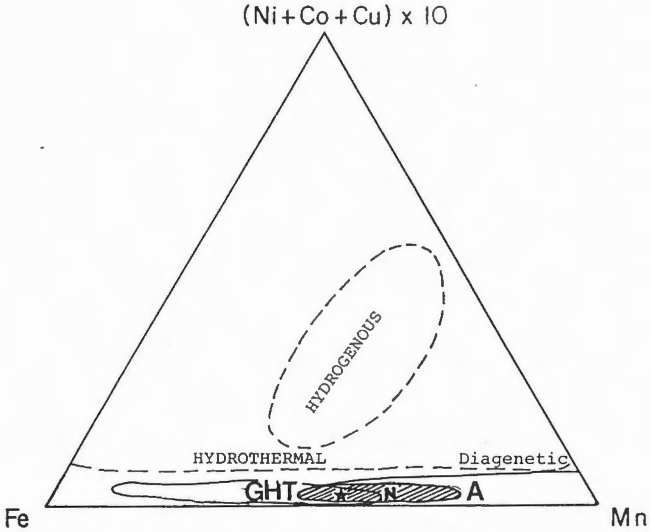

Figure 3. Fe – Mn – (Ni + Co + Cu) plot of manganiferous nodules and crusts after Bonatti (1975). A = coticules from the Ardennes (Krosse, 1983). GHT = field of metasediments of the Goldenville-Halifax transition zone (this study). N = Moshers Island Member nodules and sulphide-rich crenulated beds (this study). Small star = Eastville (MacInnis, 1986).

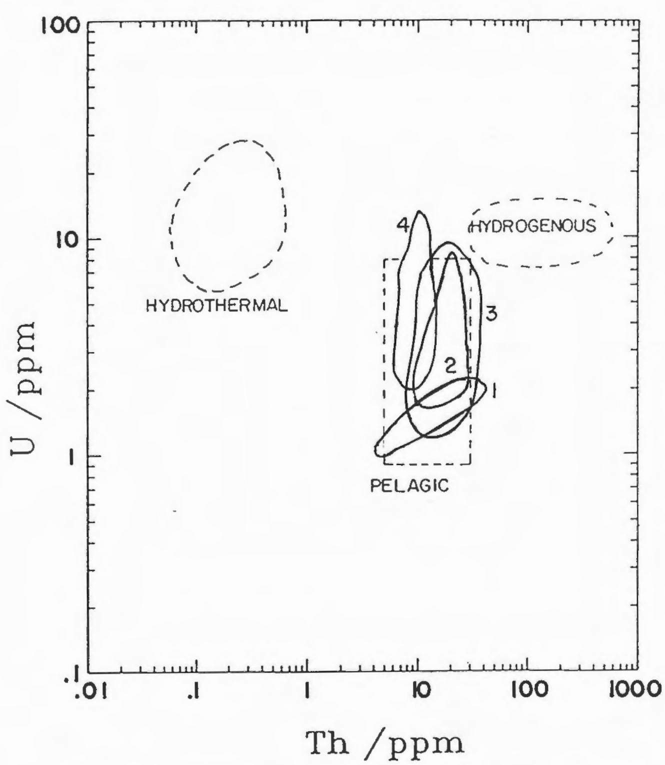

Figure 4. Relations between U and Th in manganese nodules and metalliferous sediment after Bonatti (1975). (1) = Lake Charlotte LC-86-1 lower Moshers Island member. (2) = Lake Charlotte LC-86-1 upper Moshers Island member. (3) = all other Meguma Group data reported in table 1. (4) = Eastville metasediment (MacInnis, 1986).

which formed rapidly may not have had time to scavenge metals such as Cu,Co, and Ni from seawater, and therefore plot at the bottom of the diagram. The Moshers Island deposits, as well as those from Eastville, N.S. (MacInnis, 1986), plot in the field of "hydrothermal" deposits, as do the coticules from the Ardennes, Belgium. However, Bostrom (1983) indicated that diagenetic nodules also plot in this field. A good diagram to test this possibility is the U-Th diagram of (Fig. 4), which places the GHT manganiferous deposits in the field of "pelagic sediments", clearly away from "hydrothermal" deposits; the latter generally show U/Th ratios well exceeding one (Bostrom, 1983). Greenschist metamorphism may have caused some loss of the fairly mobile U, i.e. the original U/Th ratios were probably higher, not lower than today. Figure 5 shows the relationships between Fe, Ti, Al, and Mn in various Fe-Mn-Al rich deposits. The Moshers Island rocks plot close to pelagic sediments, near the mixing line between East Pacific Rise deposits and terrigenous matter. It is interesting to note that the Eastville deposit, which has high base metals and organic carbon, overlaps the average for biological matter. If the metamorphism of the GHT has been isochemical (Cameron, 1985; MacInnis, 1986) it is suggested that the manganiferous sediments of the GHT correspond to marine deposits heavily influenced by terrigenous matter (from the erosion of a continental landmass) and by diagenetic processes involving organic matter.

The importance of the interaction of organic matter and minerals during diagenesis at Eastville has been demonstrated by isotopic studies (MacInnis, 1986). Carbonate in diagenetic concretions throughout the GHT and carbonate cement at Eastville show a restricted range of oxygen and carbon

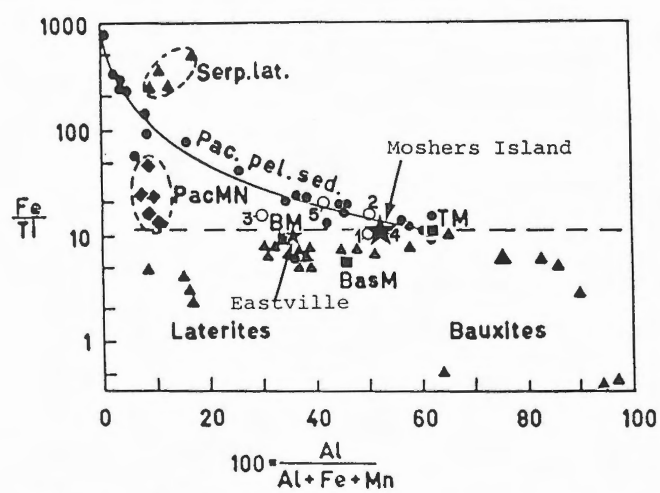

Figure 5. Relations between Fe, Ti, Al and Mn in various Fe – Mn - Al rich deposits. Bas M = basaltic matter, TM = terrigenous matter and BM = biological matter (all shown as large squares). Small triangles = laterite and bauxite; large triangle = mean bauxite. Serp. lat. = laterites formed on serpentinite. Circles = Pacific pelagic unconsolidated sediment. Diamonds = Pacific manganese nodules. Large star = Moshers Island average, this study. Small star = average of Eastville manganiferous unit (MacInnis, 1986). Solid curve represents the theoretical mixing line between East Pacific Rise deposits (top left corner), and TM (Bostrom, 1983).

isotopic ratios ($\delta^{18}O_{Smow}$ per mil; $\delta^{13}C_{PDB}$ per mil of 12.8 to 22.5 and -14.8 to -22.2 respectively). A carbonate rim surrounding a bituminous lens at Eastville is similarly depleted in $\delta^{13}C$ (MacInnis, 1986).

Strong depletion of $\delta^{13}C$ is considered indicative of carbonate formation from oxidation organic material (Hoefs, 1982) as opposed to precipitation from normal seawater. The oxidation of organic matter to produce authigenic carbonates is thought to be an important process in the diagenesis and early metamorphism of organic carbon-bearing sedimentary rocks (Anderson and Arthur, 1983), and in the genesis of metalliferous sediment (Graybeal and Heath, 1984; Kalhorn and Emerson, 1984).

MacInnis (1986) argued that Eastville carbonates formed from the diagenetic oxidation of organic material within the original sediments while suitable oxidants (O_2, NO_3-, MnO_2, Fe_2O_3, and especially $SO_4{}^{2}-$) were still available. Oxidation of organic matter to precipitate manganese-rich carbonates may have been accompanied by reduction of pore-water sulphate to form pyrite, possibly assisted by bacteria, as suggested by sulphur isotope data of A. Sangster (unpublished data). Force et al. (1983) discussed the process of intermittent oxidation of deep anoxic water (for example by renewed bottom currents): the anoxic water lies in the stability field of pyrite, where iron is removed from solution, and manganese is highly enriched. Oxidation of this water (Fig. 6) takes it through a solution field, a manganese carbonate field, and finally into manganese oxide (or oxyhydroxide) field if high Eh levels are attained. Within the GHT, the oxidized manganese field was probably rarely attained, judging by the amount of carbonaceous matter still preserved. Hingston (1985) suggested from petrographic evidence that spessartine in the coticules at Lake Charlotte evolved at the expense of manganese carbonate. At higher temperatures (ca.300°C) hydrolysis reactions are also capable of converting organic matter to carbonate, to form carbon dioxide and methane (Ohmoto and Rye, 1979). Similar values of $\delta^{13}C$ and $\delta^{18}O$ are found in metamorphogenic scheelite vein occurrences in the GHT elsewhere in the Meguma (Fisher, 1984).

In Figure 7, the composition of the GHT manganiferous metasediments are compared, on the basis of their Al-Fe-Mn ratios, with various oceanic metalliferous sediments. Notably, the GHT metasediments plot far from the fields of Nazca Plate nodules, and hydrothermal crusts, but coincide with the composition of many Pacific metalliferous sediments. As the GHT metasediments become less anomalous in manganese (i.e. from Eastville through Moshers Island Member to GHT slate), their composition converges to that of Average Shale and Average Deep Sea Sediment (Lalou, 1983), pointing towards the important influence of normal deep sea sedimentation in the formation of the GHT rocks.

The mineralogy, chemical composition, and Tremadocian age of the GHT coticules is similar to the coticules of the Ardennes, Belgium. Kramm (1973, 1976; evaluated by Krosse, 1983) has suggested that the Ardennes coticules represent the metamorphic equivalents of tuffs, which through halmyrolysis became montmorillonite-rich sediment, and

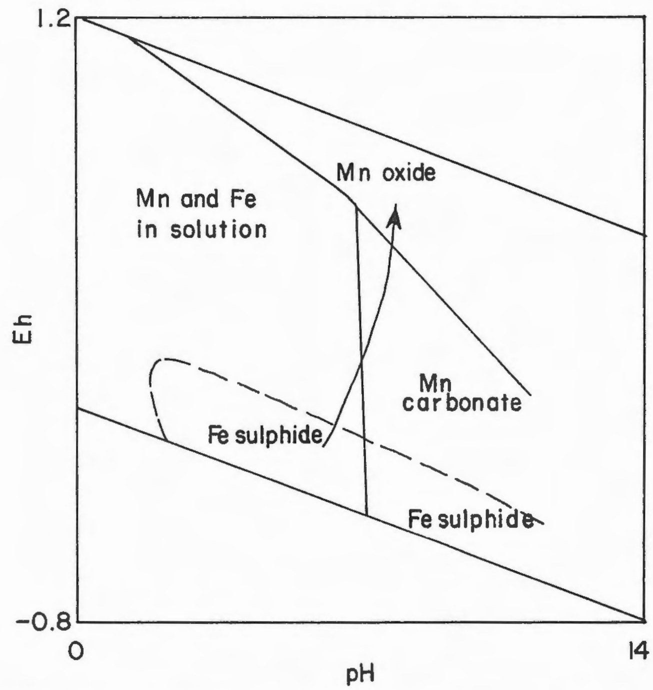

Figure 6. Generalized Eh-pH diagram showing oxidation path (arrow) of anoxic water (after Hem, 1972; Grasshoff, 1975, and Force et al., 1983) illustrating possible manganese carbonate precipitation from early diagenetic pore fluids.

through (hydrothermal?) exhalative processes became enriched in manganese and iron. Because volcanic components have not been recognized in the Meguma Group, and because volcanic material could have important implications in terms of tectonic evolution and metallogenesis of the Meguma, it is interesting to assess this possibility. Using the criteria devised by Leake (1964), based on Cr/TiO_2 and Ni/TiO_2, to evaluate the igneous versus sedimentary protolith of amphibolites, the GHT and Ardennes coticules clearly plot within the pelitic field. The criterion of Hughes (1973), which is based on alkalis, likewise suggests that the GHT coticules had no igneous protolith. The high Rb/Sr ratio of the GHT rocks (> one) is also more typical of pelites than of igneous rocks, which are generally one order of magnitude lower (Van de Kamp, 1970).

The Moshers Island Member has concentrations of Pb and Zn that are higher than the rest of the Meguma Group. The highest concentrations (755 ppm Pb, 183 ppm Zn) are found in rocks rich in iron sulphides. These are the same rocks which are high in Cu (up to 478 ppm) and Au (up to 67 ppb). Characteristically, the Zn/(Zn+Pb) ratio for samples with low concentrations of Pb and Zn is 0.65 to 0.97 (average = 0.88), whereas for metal rich samples the Zn/(Zn+Pb) ratio is less than 0.5 reflecting, primarily, higher Pb concentrations. The Moshers Island lithology at Eastville has an average ratio of 0.78. These values compare with 0.71 for the Earth's crust, 0.75 for ashes of oil, 0.88 for greywacke, and 0.96 for average pelitic sediments. The consistency of the Zn/(Zn+Pb) ratios throughout the GHT, and the

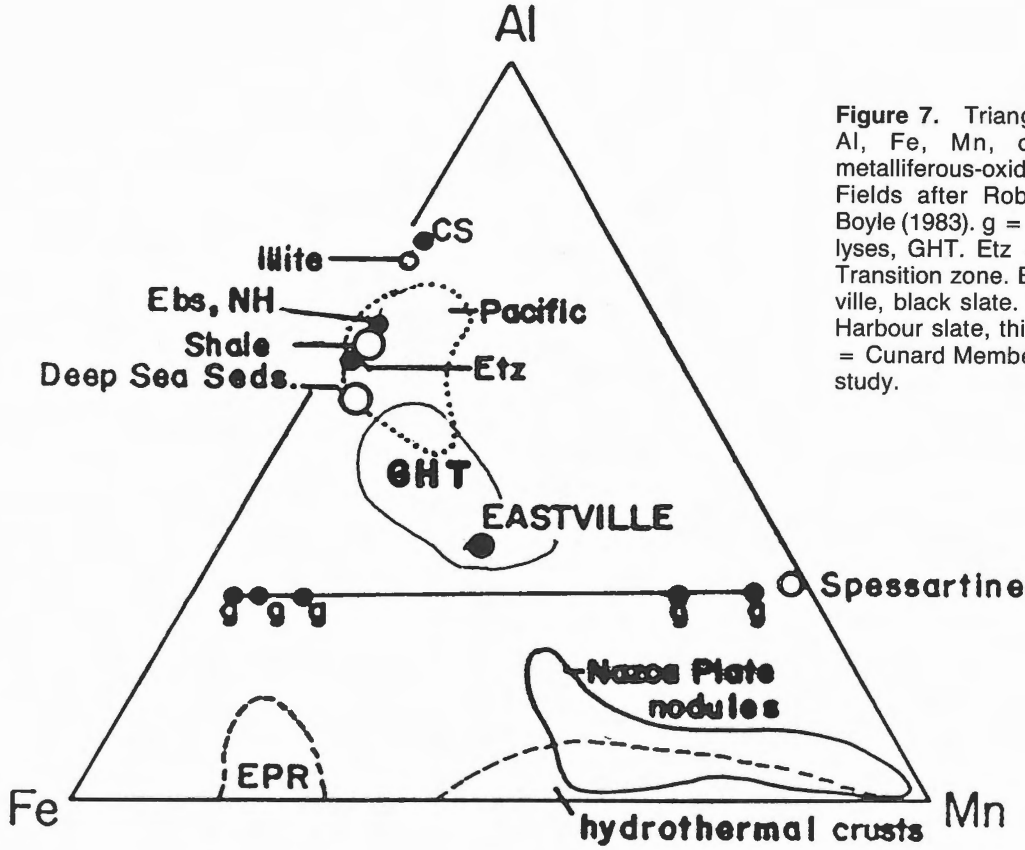

Figure 7. Triangular plot of Al, Fe, Mn, of oceanic metalliferous-oxide sediment. Fields after Robertson and Boyle (1983). g = garnet analyses, GHT. Etz = Eastville, Transition zone. Ebs = Eastville, black slate. NH = New Harbour slate, this study. CS = Cunard Member slate, this study.

similarity of its values with normal sedimentary products, indicate that the behaviour of these elements in the GHT could be explained by normal processes in sedimentary basins where greywacke is being deposited, though the concentration of metals in these metasediments is high.

Eastville has been described as a syngenetic stratabound Zn-Pb deposit (Binney et al., 1986). This work shows that lithochemically and isotopically the GHT is similar to Eastville at other sites along the contact. On his discussion of chemical parameters of stratiform lead-zinc deposits Lydon (1983) suggested that major Pb+Zn concentrations in which the Zn/(Zn+Pb) ratio in between 0.7 and 0.95, metal enrichment was due to chloride-rich, Cu-poor, ore fluids, probably at temperatures below 200°C, saturated with sphalerite and galena, which originated within the sedimentary pile. The abundance of organic matter at the GHT would suggest low fO_2 of the chemical system, and under those conditions Lydon (1983) would predict an abundance of Ba. Barium is 2 to 5 times higher in the GHT compared to average pelitic sediments. Lydon's model calls for a thick sedimentary pile with an impervious shale cap, similar to the configuration of the Halifax Formation overlying the Goldenville Formation. The GHT environment, even at considerably lower Zn+Pb concentrations, is therefore physically and chemically compatible with the genetic model proposed by Lydon (1983). What are missing are the loci for "third order basins" (Large, 1983) and syndepositional faults to localize outflow and syngenetic deposition. This remains a challenge for sedimentologists and structural geologists working in the Meguma Group.

PRELIMINARY CONCLUSIONS

Lithochemistry has been useful in characterizing lithological divisions in the Meguma metasedimentary rocks. It has confirmed the distinctiveness of the Moshers Island Member of the Halifax Formation. The Moshers Island Member is anomalous in concentrations of Mn, C, Ba, and many trace metals such as Pb, Cu, Zn, Mo, W, and Au. The metal enrichment of the Moshers Island Member supports the concept that the GHT may be of important metallogenic and exploration significance. The GHT has a distinct lithochemical signature which can be seen over several metres in most sections of the contact and can be stretched out vertically over several hundred metres of section in the south-central portion of the area. Mapping this unit should be an important prerequisite to assessment of the contact for mineralization.

It can be concluded that the manganiferous metasediments of the GHT had a sedimentary protolith characteristic of deep sea terrigenous sediments. The transition zone coticules are the product of manganese carbonate precipitation from pore fluids near the sediment-water interface during early diagenesis by oxidation of organic matter. Regional metamorphism has developed spessartine garnets at the expense of the carbonate. Metal enrichment in these rocks appears to accompany sedimentary-diagenetic processes during anoxic conditions prevalent at the time of GHT development.

Studies of this kind are not only useful as a guide to mineral exploration but are necessary in establishing base lines for regional geochemical surveys and environmental studies.

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New interpretation of the structural and stratigraphic setting of the Cutwell Group, Notre Dame Bay, Newfoundland[1]

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Szybinski, Z.A., New interpretation of the structural and stratigraphic setting of the Cutwell Group, Notre Dame Bay, Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 263-270, 1988.

Abstract

The rocks of the Cutwell Group record a two stage structural history. The first deformation (D1) was Silurian or earlier and resulted in north-directed thrusting and folding. Previously-interpreted lithostratigraphic sequences reflect structures related to this deformation. The second deformation (D2) resulted in east-west, dextral, strike-slip faulting, manifested mainly in movement on the Lobster Cove Fault. This deformation involved Silurian rocks and so the latest movement was Silurian or younger.

From stratigraphic sequence within thrust slices, the reconstructed stratigraphy of the Cutwell Group is: 1) a lower mafic to intermediate volcanic member; 2) an intermediate, dominantly felsic, volcanic member; and 3) an upper sedimentary member capped by Caradocian shale. There are no post-Caradocian volcanics in the group.

Volcanogenic sulphide deposits occur mainly within the Long Tickle Formation of the Cutwell Group. Correlation suggests that other felsic volcanic sequences in the Cutwell Group should have a similarly high exploration potential.

Résumé

Les roches du groupe de Cutwell offrent le profil d'une évolution structurale à deux stades. La première déformation (D1) s'est produite au Silurien ou avant et elle a provoqué un charriage et un plissement à direction nord. Des séquences lithostratigraphiques déjà expliquées confirment l'existence de structures liées à cette déformation. La seconde déformation (D2) a causé un décrochement dextre est-ouest qui s'est manifesté surtout par des mouvements sur la faille de Lobster Cove. Cette déformation a affecté des roches siluriennes, c'est pourquoi le dernier mouvement appartient au Silurien ou à une époque plus récente.

À partir d'une séquence stratigraphique comprise dans des lambeaux de charriage, la stratigraphie reconstituée pour le groupe de Cutwell est la suivante : 1) un niveau volcanique inférieur allant de mafique à intermédiaire ; 2) un niveau volcanique de type intermédiaire et surtout felsique et 3) un niveau sédimentaire supérieur coiffé par du schiste argileux caradocien. Il n'y a pas de roches volcaniques post-caradociennes dans le groupe.

Des dépôts de sulfure volcanogénique se rencontrent surtout à l'intérieur de la formation de Long Tickle du groupe de Cutwell. Une corrélation suggère que d'autres séquences volcanofelsiques dans le groupe de Cutwell devraient posséder un potentiel d'exploration aussi élevé.

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INTRODUCTION

An investigation of the Cutwell Group has been carried out as part of the project "Metallogeny of the Buchans — Roberts Arm Belt", a regional study of the metallogeny of non-ophiolitic rocks in the Central Mobile Belt of Newfoundland.

The Cutwell Group has previously been mapped on Long Island (Kean, 1973) as a continuous, south-facing sequence of volcanic rocks greater than 5 km thick. Contradictory, fossil-based ages, however, for different lithostratigraphic sequences within the Cutwell Group on Long Island suggest that there are major structural problems which need to be addressed in order to understand the stratigraphic and structural framework of the group. The presence of two significant volcanogenic sulphide deposits in the Cutwell Group emphasizes the importance of such a study for further assessment of the mineral exploration potential of this area.

The main objective of this field season has been to establish the structure and stratigraphy of the Cutwell Group and the stratigraphic setting of know mineral deposits and to determine the implications of the latter with regard to mineral exploration in the area.

Previous work

Espenshade (1937) first investigated the general geology and mineral deposits of the Pilley's Island area. Recognizing a major E-W trending fault which he named the Lobster Cove Fault, he assigned rocks to the north of it to "the Pilley's Series", and those to the south of it to "the Badger Bay Series". He further divided the Pilley's Series into a lower group of dominantly pyroclastic and volcanic rocks, which he named the Cutwell Group, and an upper group of mainly pillow lava, which he named the Lushs Bight Group.

Williams (1962), during reconnaissance mapping of the west half of the Botwood map sheet, found that rocks conformably overlying the Cutwell Group on Long Island were lithologically different from most of the rocks assigned to the Lushs Bight Group by Espenshade (1937). Williams (1962) included all rocks on Long Island in the Cutwell Group which he considered to be mid-Ordovician (Llanvirnian ?) based on fossils collected on Limestone Island near Little Bay Island.

Strong and Kean (1972) described newly-discovered fossil localities in the Cutwell Group on Long Island. Assemblages of conularids, crinoids, ostracodes and cephalopods from thin limestone lenses within pyroclastic rocks and pillow lava, were tentatively assigned to a Middle Ordovician age, supporting Williams (1962) suggested age for the Cutwell Group.

Kean (1973) mapped the Cutwell Group on Long Island in detail, describing it as a continuous, south-facing sequence of volcanic rocks, comprising, from bottom to top: 1) pillow lava, intrusive breccia and diabase of the Stag Island Formation; 2) deep-water chert and turbidite of the Pigeon Head Formation; 3) coarse agglomerate of the Quinton Cove Formation; 4) pillow andesite, intermediate pyroclastic rocks, cherty shale and tuff of the Burnt Head Formation; 5) shallow water limestone, greywacke and shale of the Parson's Point Formation; and 6) agglomerate, pillowed to massive lava, tuff and thin beds of fossiliferous limestone of the Long Tickle Formation. Intrusive and extrusive dacitic rocks that occur between the upper sequences of the Burnt Head Formation and the lower to middle sequences of the Long Tickle Formation, he assigned to the Seal Cove Complex. Kean (1973) interpreted these to represent a near-surface dacitic plug. On the basis of chemical composition and stratigraphy, he suggested that volcanic rocks in the succession are of island-arc affinity and in a later paper (Kean and Strong, 1975) interpreted the succession to represent a transition from immature to mature island arc volcanic activity.

Dean (1977, 1978) described the volcanic stratigraphy and metallogeny of the Notre Dame Bay area. On the basis of a Caradocian graptolite fauna collected from black shale of the Parson's Point Formation, he included formations underlying this shale in the pre-Caradocian "immature island arc" sequences and correlated them with the Western Arm, Catchers Pond, Pacquet Harbour and Wild Bight Group. Formations overlying the Parson's Point Formation, he included in post-Caradocian "mature island arc" sequences, correlating them with the Springdale, Mic Mac and Roberts Arm groups (then thought to be of Silurian age).

Stouge (1980) attempted correlation of Early and Middle Ordovician conodonts from Central Newfoundland with conodonts from the western part of the island. In accord with the conclusions of Williams (1962) and Strong and Kean (1972), the conodont fauna in Cutwell Group limestone suggested a Late Arenig-Early Llanvirn age.

Swinden and Thorpe (1984) presented lead isotope data for volcanogenic sulphide occurrences in the Cutwell Group (Oil Islands and Shamrock prospects). They suggested that lead in these deposits came from a similar source to that in deposits in the Buchans and Roberts Arm groups.

Kean (1984), during mapping of the Lushs Bight Group, showed that rocks assigned to this group on Pilley's Island and Triton Island including pillow lava, locally with inter-pillow chert, tuff breccia, tuff and red argillite with "iron formation", were lithologically unlike the typical Lushs Bight Group. He assigned these sequences to the Cutwell Group (Kean, 1987).

REGIONAL GEOLOGY

The Cutwell Group lies within the Dunnage Zone of the Central Mobile Belt of Newfoundland (Williams, 1964; 1979) which records the Cambrian and Ordovician history of an oceanic volcanic and sedimentary terrane commonly referred to as Iapetus. Rocks of the Cutwell Group outcrop in western Notre Dame Bay from League Rock (about 5 km east of Long Island) to the western shores of Halls Bay Head and Little Bay Island (Fig. 1).

Contacts with adjacent units are generally faulted. The Cutwell Group on Long Island is separated from the Lushs Bight Group on the northern shore of Pilley's Island by the Long Tickle Fault. On Halls Bay Head, the Cutwell Group is faulted against the Lushs Bight Group along the MacLean Fault (Donahoe, 1968; Dean and Strong, 1977).

The nature of the southern contact of the Cutwell Group in this area is of regional importance. Espenshade (1937) was first to recognize this east-west trending structural discontinuity and called it the Lobster Cove Fault. This fault separates the Cutwell Group from the Lower Ordovician Roberts Arm Group, which has a U-Pb (zircon) age of 473 $\pm$ 2 Ma (Dunning et al., 1987), on Pilley's and Triton islands. Shallow-water red sandstone and conglomerate, assigned to the Silurian Springdale Group, disconformably overlie the Roberts Arm volcanic rocks immediately south of the fault. Dean and Strong (1977) suggested, on the basis of regional geology, that the Lobster Cove Fault and other faults north of it (such as the Long Tickle and MacLean Faults) are folded, south directed thrust faults.

STRUCTURAL GEOLOGY

Detailed geological mapping has shown that the volcanic rocks of the Cutwell Group were affected by two major deformational events:

1) The first deformation (D_1) is recorded by generally north-directed thrusting and folding;

2) The second deformation (D_2) comprises dominantly dextral strike-slip faulting such as that recorded by the Lobster Cove Fault.

The following interpretation is based on a preliminary compilation of field data.

The Cutwell Group contains an extensive thrust system, hereafter referred to as the Cutwell Group thrust belt, which controls the present distribution of lithostratigraphic units. The over-all structural style of the group suggests that the thrust sheets define a relatively complicated duplex system. On Triton Island this duplex system forms a northeast-southwest trending, gently east-plunging, structural culmination, interpreted as an antiformal stack (Boyer and Elliot, 1982), which is herein informally termed the Triton antiformal stack. On Triton Island, individual thrust sheets contain contrasting lithostratigraphic assemblages. Andesitic tuff breccia typical of the upper sequences of the Long Island Formation forms the footwall of the antiformal stack. Pillow lava lies above the roof thrust of the antiformal stack and the imbricate zone comprises laminated red-green cherty sediment interbedded with reworked coarse tuff. The imbricate zone contains at least four horses arranged in a foreland-dipping duplex immediately north of the culmination axis. The internal structure of Pilley's Island is less clear, but mapping and reconnaissance structural observations indicate that it may be a hinterland-dipping duplex. Its relationship to the Triton antiformal stack is uncertain.

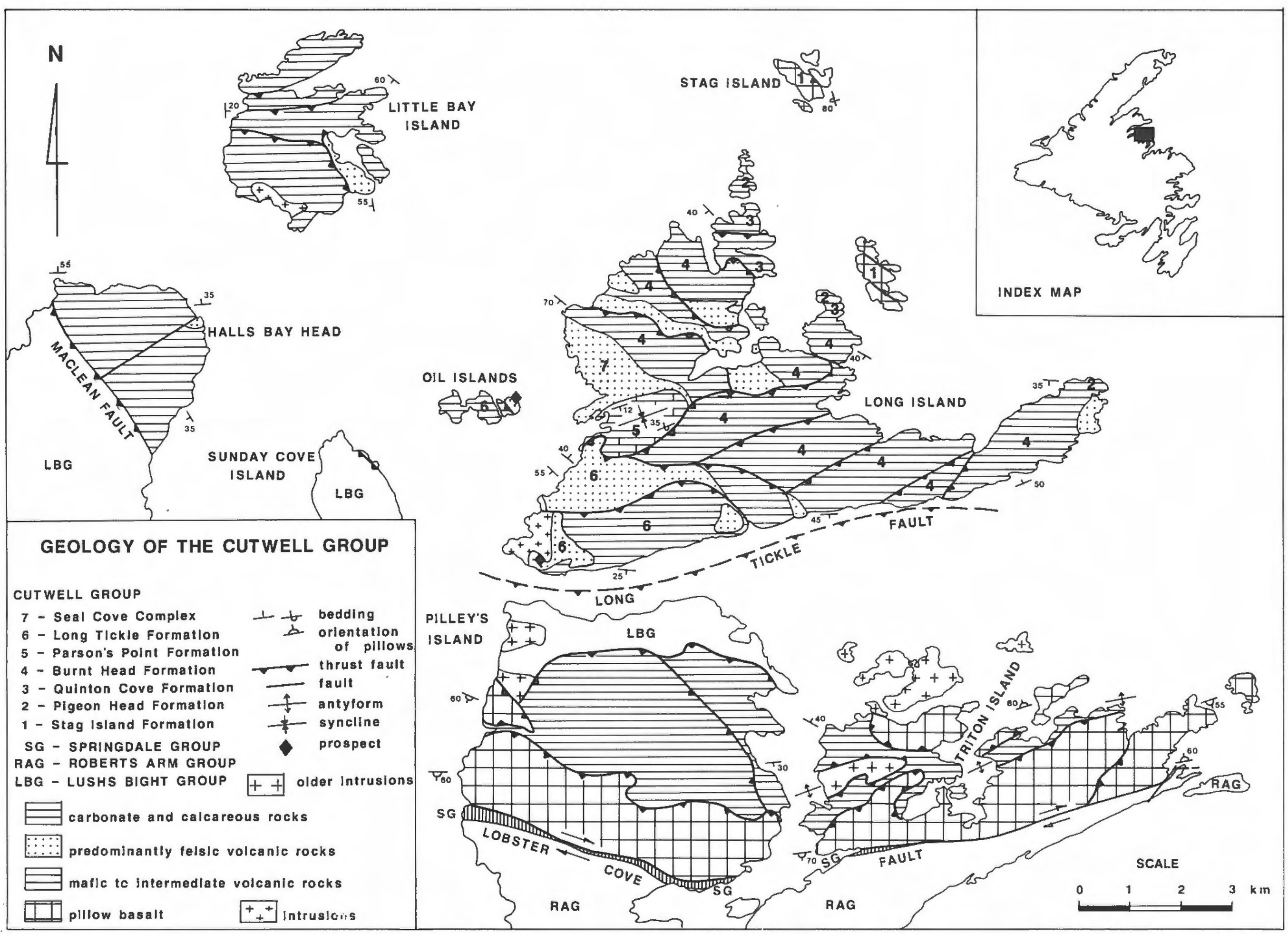

Figure 1. Geology and structural framework of the Cutwell Group. Nomenclature of formations on Long Island after Kean (1973, 1987). Distribution on Pilley's and Triton Islands mostly after Kean (1987).

Another duplex has been recognized on Long Island. It is probably not related to the Triton structure. The southern part of the island resembles a hinterland-dipping duplex. Most of the Long Tickle Formation probably lies above the roof of the duplex, but the floor thrust of this structure has not been identified and may not be exposed.

The MacLean, Long Tickle and Lobster Cove faults have previously been interpreted as south directed, folded thrust faults (Dean and Strong, 1977). The present mapping suggests that the MacLean and Long Tickle faults are indeed moderately south dipping, characterized by well developed schistosity and downdip mineral lineation, thrust zones. However, contrary to the interpretation of Dean and Strong (1977), kinematic indicators along the MacLean Fault imply northeasterly movement and the Lushs Bight Group appears to be thrust over the Cutwell Group. The Long Tickle Fault between Long Island and Pilley's Island is mainly under water. A ductile shear zone, considered by previous workers to be trace of this fault, outcrops on the northeastern tip of Sunday Cove Island where, again, rocks of the Lushs Bight Group lie structurally on rocks assigned to the Cutwell Group. However, the fault in the Pilley's and Long islands area has a somewhat different style, being much steeper than other thrust faults in the belt. Rocks of the Lushs Bight Group on northern Pilley's Island lie in the hangingwall of this steepened thrust fault and form the sole on which rests the Pilley's Island duplex. The thrust fault may be a relatively deep-seated feature, since it brings rocks of ultramafic basement against the Cutwell Group on the southern shore of Long Island (Kean, 1972). The steep attitude of the thrust fault is probably due to tilting of the structure in relation to formation of the Long Island duplex in its footwall.

In general, thrust planes and the axial surfaces of related folds are approximately parallel to the bedding. A regionally developed, southerly dipping, penetrative cleavage (S1) is locally parallel to the axial planes of thrust-related folds. The asymmetry of these minor folds and mesoscopic kinematic indicators imply north-northwest directed movement on Long, Pilley's and Triton islands faults. Kinematic indicators and lineation attitudes indicate north-northeasterly directed movement in the Halls Bay Head and Little Bay Island area faults.

The distribution of movement zones and the geometry of kinematic indicators in these zones varies from area to area. For example, in the northwestern part of Halls Bay Head, fault movement is confined to narrow zones, in which mesoscopic kink folds overprint the S1 cleavage. Asymmetry of the kinks, combined with the lack of deformational features in adjacent rocks, suggests that back slip related to collapse of the thrust stack, has occurred on narrow movement zones. In other areas, for example in northern Triton Island near the hinge area of the Triton antiformal stack, deformation related to thrusting is much more widespread and complex. Rocks in this area are intensively folded and faulted (Fig. 2a, b).

There is some evidence for relative and absolute ages of the deformational events. The Lobster Cove Fault postdates the D_1 deformation because it truncates shear zones related to D_1 thrusting on Triton, Pilley's and Sunday Cove islands. On the basis of kinematic indicators, such as small folds (Fig. 2c), shear bands and lineation, which are indicative of dextral, strike-slip movement, this structure is considered to be a wrench fault. It also involves red sandstone and conglomerate of the Silurian Springdale Group.

The Roberts Arm Group (U-Pb zircon age of 473 $\pm$ 2 Ma, Llanvirnian ?, Dunning et al., 1987) is approximately coeval with the Cutwell Group (Late Arenig-Early Llanvirnian fauna, Stouge, 1980). Reconnaissance work on the structure of the Roberts Arm Group suggests that rocks of both the Cutwell and Roberts Arm groups were deformed by the D_1 event. The upper limit on the age of thrusting in the Cutwell Group is therefore considered to be Late Ordovician, since rocks of the Springdale Group elsewhere in general show no penetrative deformation of any kind. The lower limit on the age of thrusting may be deduced from intrusive relationships within the group (see below).

REMARKS ON STRATIGRAPHY AND LITHOLOGY

The Cutwell Group contains the only known exposures of Caradocian shale north of the Lobster Cove Fault (Dean, 1977, 1978). The stratigraphy developed by Kean (1973) and the presence of the Caradocian shale in the middle of the sequence led Dean (1977, 1978) to suggest that the Long Tickle Formation represented the only example of post-Caradocian volcanic rocks in central Newfoundland.

Present detailed mapping, however, shows that the previously developed lithostratigraphic scheme is in fact a structural order (i.e. the order in which lithological units appear across the Long Island duplex). Identical rock units, representing the same stratigraphic level, reappear in different formations, or, according to the present interpretation, at different structural levels.

A good example of this is the Pigeon Head Formation and the lower Pyroclastic member of the Burnt Head Formation (this and other names of formations and members are after Kean, 1973). Both rock units consist of fine-laminated, cherty black shale interbedded with green reworked tuff (turbidite) and grey chert. In both units, the tuff contains clasts of plagioclase-phyric andesite, and both are intruded by gabbro plugs and diabase sills. They are here interpreted to be stratigraphic equivalents.

A similar situation occurs in the Quinton Cove Formation and the upper Pillow-andesite member of the Burnt Head Formation. Both rock units consist of tuff breccia, predominantly with porphyritic andesite fragments, and pyroxene-bearing tuff. The latter contains andesite flows as well. The Long Tickle Formation, thought previously to represent a post-Caradocian volcanic event, is lithologically very similar to both of the above mentioned formations, and it contains minor fossiliferous limestone associated spatially with red cherty sediment with a Late Arenig-Early Llanvirn fauna. Geochemical data for andesitic pillow lavas from the Burnt Head and Long Tickle formations support their correlation (Fig. 3).

Rocks of the Seal Cove Complex and felsic rocks assigned to the Long Tickle Formation are also lithologically

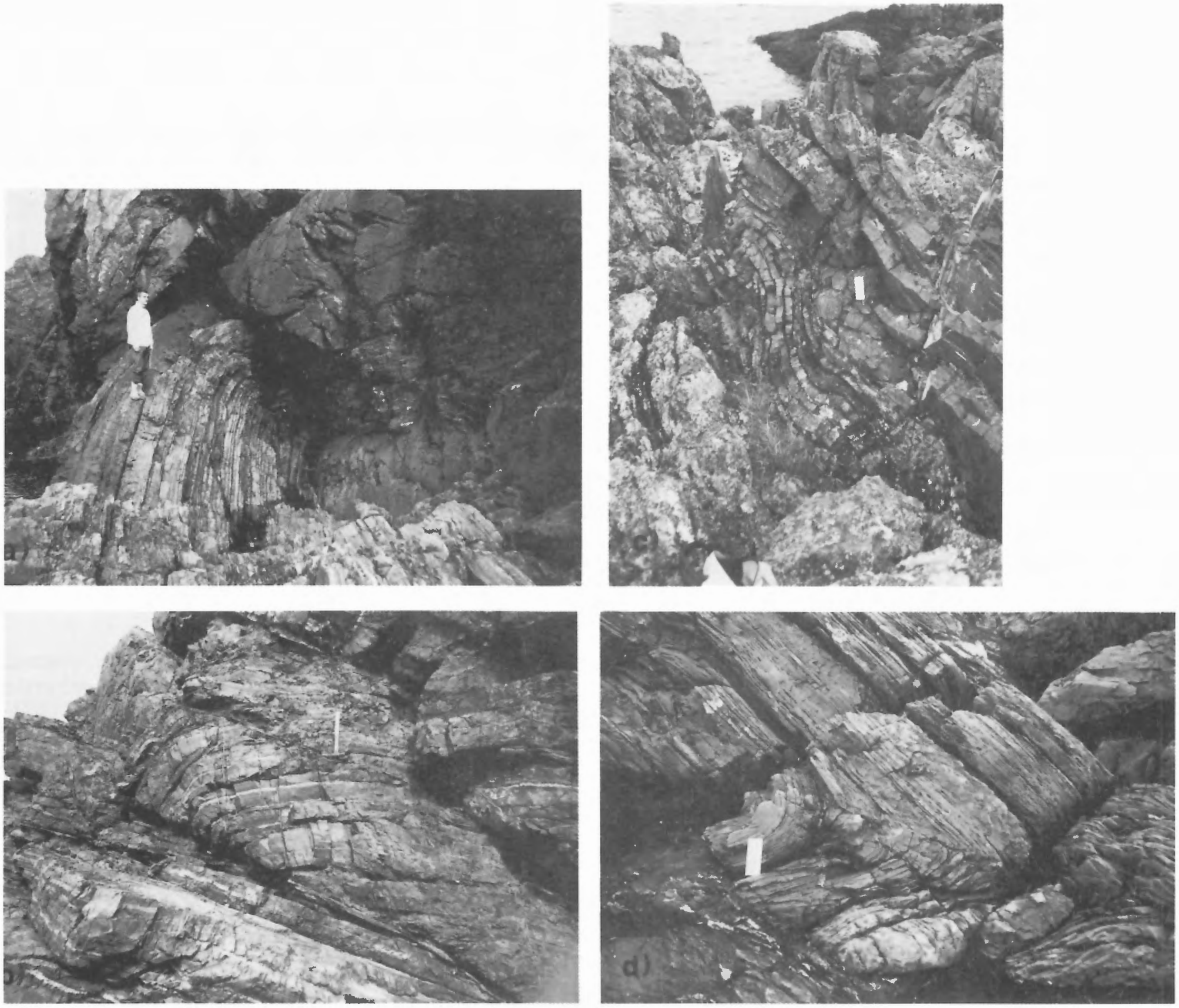

Figure 2. a) Folded strata in the hinge area of Triton antiformal stack (looking east); b) hangingwall ramp on footwall flat in the hinge area of Triton antiformal stack (looking east); c) an asymmetrical fold, verging northward, indicative of north-directed movement on thrust fault, east end of Triton Island (looking east); white bar is 15 cm long; d) asymmetrical, west-closing fold in sandstone and conglomerate of the Springdale Group on Lobster Cove Fault (looking south); white bar is 15 cm long.

similar. This is of considerable economic interest as felsic pyroclastic rocks of the Long Tickle Formation are host to two major mineral occurrences within the Cutwell Group (see below).

The Parson's Point Formation, which contains Caradocian shale at the top, seems to have a relatively wide time span. Blocks of limestone typical of this formation occur at the top of the Pyroclastic member of the Burnt Head Formation and are found throughout the Breccia member of the Stag Island Formation as well (see below). The base of the Parson's Point Formation consists of carbonate breccia with clasts of fossiliferous limestone and minor clasts of plagioclase-phyric andesitic tuff. The limestone contains a faunal assemblage similar to that in limestone beds in pyroclastic rocks of the Long Tickle Formation (Upper Arenig, Strong and Kean, 1972; Stouge, 1980). Carbonate breccia is also interbedded with graptolitic Caradocian shale.

The Stag Island Formation was interpreted by Kean (1973) to be the lowermost unit of the Cutwell Group. Lithologies typical of this unit do not occur in other formations. However, the presence of gabbro, diabase, diorite-granodiorite, minor limestone and porphyritic andesite clasts in the Breccia member of this formation suggests that it may be approximately coeval with andesite volcanism in nearby formations. This is also in agreement with the interpreted order of stacking.

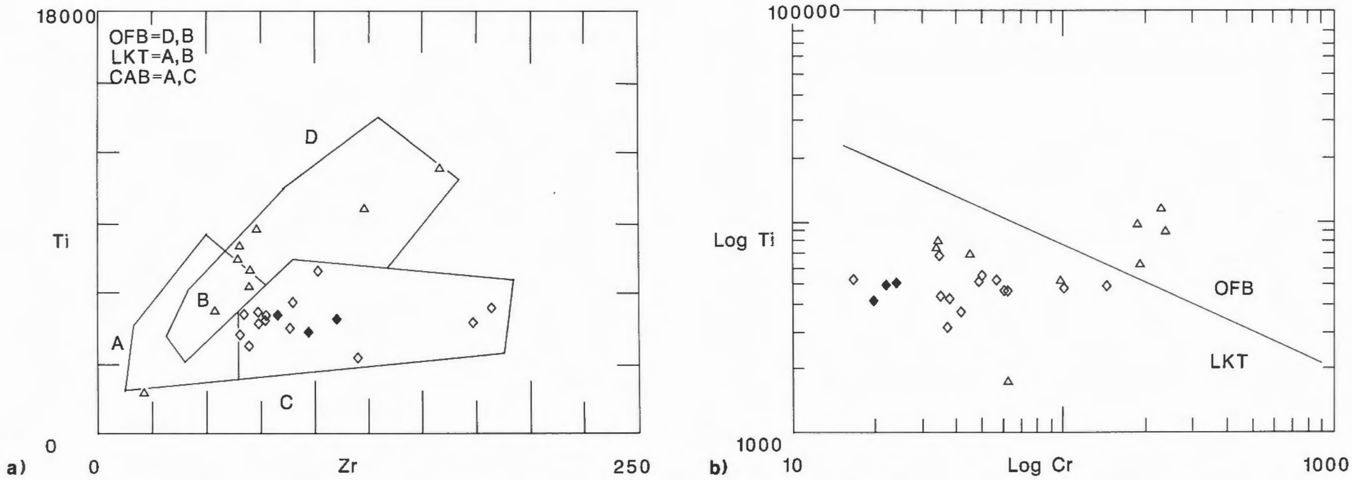

Figure 3. Comparison of the geochemistry of volcanic rocks in the Cutwell Group: a) Ti versus Zr diagram after Pearce and Cann (1973). b) Ti versus Cr diagram after Pearce (1975). Data are from Kean (1973). OFB = ocean floor basalts, LKT = low-K tholeiites, CAB = calc-alkaline basalts. Symbols for both diagrams are: open triangles — Stag Formation, open diamonds — Burnt Head Formation, filled diamonds — Long Tickle Formation. Volcanic rocks in the Long Tickle Formation are geochemically similar to rocks of the Burnt Head Formation, supporting their correlation.

The Cutwell Group is intruded by abundant dykes, sills and small plutons, ranging from gabbro and diabase to quartz-feldspar porphyry. Gabbro, cut by diabase dykes, forms the structural base of the Stag Island Formation. The Breccia member of this formation, the Pigeon Head, Quinton Cove and the Pyroclastic member of the Burnt Head Formation are intruded by diabase dykes and sills, which can be traced into gabbro plugs. The first intrusive phase of the Long Island Pluton (Kean, 1973) is also basic in composition. No mafic dykes or plugs are present in rocks of the Seal Cove Complex or the Parson's Point Formation. Field observations suggest that basic intrusions are pre-D_1, and they are interpreted to be related to andesitic phase of volcanism. This idea will be further tested by geochemistry.

Dioritic to granodioritic plutons are common in the Cutwell Group on Long and Little Bay islands. The Long Island Pluton and some other small dioritic-granodioritic bodies are spatially related to felsic rocks of the Seal Cove Complex, suggesting that they may be related to the centres of felsic volcanism. In places, similar granodiorite intrudes along thrust surfaces and also intrudes rocks of the Seal Cove Complex. Thrusting seems to involve the dioritic (older) phase of the Long Island Pluton, but not granodioritic (younger, main) phase of this intrusion.

Postkinematic quartz-feldspar porphyry dykes cut throughout all rock units of the Cutwell Group, including Caradocian shale of the Parson's Point Formation. There is some evidence in individual thrust sheets for stratigraphic succession in the Cutwell Group as whole. Field observations show that, in all cases, felsic rocks stratigraphically overlie mafic rocks. Intrusions related to the volcanic activity cut mafic volcanic rocks, but not felsic rocks or sedimentary rocks of the Parson's Point Formation.

On this basis, the Cutwell Group is considered to represent a structurally disrupted stratigraphic succession consisting of: 1) a lower, mafic to intermediate, volcanic and pyroclastic member (this includes rocks assigned to the Pigeon Head, Quinton Cove, Burnt Head and Long Tickle formations); and 2) a middle felsic volcanic member related to a dacitic-rhyodacitic volcanic event (rocks presently assigned to the Seal Cove Complex, felsic rocks of the Long Tickle Formation and felsic domes within the Burnt Head Formation); and 3) a sedimentary unit consisting of dominant limestone breccia capped by Caradocian shale (the Parson's Point Formation).

The relationship of the Stag Island Formation to the above units remains uncertain.

IMPLICATIONS FOR MINERAL POTENTIAL

Several volcanogenic sulphide occurences are present in the Cutwell Group, hosted by both mafic and felsic rocks. Sulphide mineralization in mafic rocks of the Halls Bay Head area occurs in shear zones, along which primary volcanogenic mineralization has been remobilized. Mineralization in these zones is mostly pyrite and lesser chalcopyrite. The felsic rocks assigned to the Long Tickle Formation (Kean, 1973; Dean, 1977, 1978), which host the mineralization, lithologically resemble the Seal Cove Complex. Only two showings (the Oil Island and Shamrock prospects) are of economic interest and mineralization in both areas is briefly described below.

Oil Islands Prospect

The Oil Islands deposit occurs as disseminated pyrite with minor chalcopyrite, sphalerite and galena hosted by crystal lithic dacitic tuff. This mineralization and associated alteration is exposed along the north shore of the most easterly of Oil Islands. The extension of the mineralized horizon is covered by beach deposits.

The host rock is strongly foliated and mineralization locally occurs in bands parallel to foliation. Rare clasts of massive sulphide are locally present.

Mineralization occurs as well in mineralized quartz veins, 10-15 cm thick, cutting massive andesite.

The Oil Island deposit has been extensively explored (including drilling) by joint venture of Brinco and Getty Mines Ltd. in early 1980s. Mineralization has proved to be subeconomic.

Shamrock Prospect

Felsic (dacitic to rhyolitic) strongly foliated and altered rocks of Long Tickle Formation host the Shamrock deposit. Mineralization is exposed in several prospect pits, approximately 0.4-0.5 km northeast of the ferry landing on Long Island. It consists predominantly of pyrite with associated sphalerite and galena, occurring as a breccia matrix, lenses parallel to foliation, disseminations and in thin quartz veins. The host rock is predominantly felsic crystal-lithic tuff overlying rocks of predominantly andesitic composition assigned to the Burnt Head Formation. Hydrothermal alteration is present through the host rock in the immediate area of the deposit in the form of seritization and silicification.

CONCLUSIONS

The field work summarized in this report leads to some new conclusions concerning the stratigraphy, structure and economic geology of the Cutwell Group.

1) Two main deformational events involving rocks of the Cutwell Group have been recognized. The D_1 event resulted in extensive, north directed thrusting. D_2 is recorded by major, strike-slip, dextral movement on the Lobster Cove Fault.

2) The Cutwell Group consists of structurally repeated lithological units. The restored stratigraphy is interpreted to comprise a lower horizon of mafic volcanic rocks, an upper felsic unit and a cap of carbonaceous and calcareous sediment. Elements of this succession can be recognized throughout the group.

3) Contrary to previous interpretations, no post-Caradocian volcanic rocks are present within the sequence.

4) Recognition of extensive thrusting and repetition of lithological units suggests that felsic pyroclastic rocks that host volcanogenic sulphide deposits on Long Island are related to felsic units elsewhere within the Cutwell Group. This raises the possibility that more mineral occurrences remain to be discovered and felsic rocks of the Seal Cove Complex may be a promising target for future exploration.

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Preliminary seismostratigraphic and geomorphic interpretations of the Quaternary sediments of Hudson Bay[1]

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Josenhans, H., Balzer, S., Henderson, P., Nielson, E., Thorliefson, H., and Zevenhuizen, J., Preliminary seismostratigraphic and geomorphic interpretations of the Quaternary sediments of Hudson Bay; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 271-286, 1988.

Abstract

A preliminary interpretation of the Quaternary stratigraphy and seafloor morphology of Hudson Bay is presented based on recently collected high resolution single channel (653 cm³) airgun, Huntec D.T.S., 3.5 kHz and sidescan sonar seismic data. Isopach maps of Quaternary sediment thickness and acoustic outcrop were prepared at a scale of 1:1 000 000 and are shown here at reduced scale. Based on seismostratigraphy the Quaternary sequence can be divided into three destinct units overlying bedrock. These are interpreted to represent: glacial till, glaciomarine stratified sediments, and postglacial stratified sediments. Geomorphic evidence obtained from sidescan sonar data shows many unique seafloor types which indicate glacial flow directions as well as the processes active during the final disintigration of the Laurentide Ice Sheet.

Résumé

On présente ici une interprétation préliminaire de la stratigraphie quaternaire et de la morphologie du fond marin de la baie d'Hudson fondée sur des données sismiques récemment recueillies à l'aide d'un sonar à balayage latéral de 3,5 kHz de type Huntec D.T.S., et d'un canon à air (653 cc) à bande unique à haute résolution. Des cartes isopaques des sédiments quaternaires et de l'affleurement acoustique ont été préparées à l'échelle de 1/1 000 000 et elles sont présentées ici à une échelle réduite. En s'appuyant sur la sismostratigraphie, la séquence quaternaire peut se répartir en trois unités distinctes sus-jacentes au substrat rocheux. On considère qu'elles doivent représenter les éléments suivants: du till, des sédiments stratifiés glaciomarins et des sédiments stratifiés postglaciaires. Des preuves géomorphologiques obtenues à partir de données de sonar à balayage latéral montrent de nombreux types exceptionnels de fonds marins qui mettent en évidence des directions d'écoulement glaciaire de même que les processus actifs pendant la disparition de l'inlandsis des Laurentides. Certains exemples de coupes sismiques à haute résolution et de profils de sonar à balayage latéral sont illustrés.

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INTRODUCTION

The results from a regional high resolution seismic survey of Hudson Bay provide insights into the stratigraphy of the area which have not been resolved by previous surveys and lead to new interpretations of the glacial and postglacial history. While Quaternary researchers generally agree that the Hudson Bay region has been covered by thick continental glaciers, much debate persists regarding the age, dynamics, flow direction and configuration of the Pleistocene ice sheets. This report marks the beginning of a systematic, multi-year regional high resolution seismic and sampling survey program of Hudson Bay intended to map the regional stratigraphy and surficial distribution of Quaternary sediments (and bedrock; Grant, and Sanford, 1988) within the bay. The data resulting from this study will, for the first time provide rigorous geological constraints which will help to develop a model for Hudson Bay deglaciation.

PREVIOUS WORK

The first major marine geological and geophysical investigations of Hudson Bay were undertaken in 1961 (Leslie, 1964; Leslie and Pelletier, 1965; Pelletier, 1966) and were followed in 1965 by a more extensive program. The results, summarized in Pelletier et al. (1968), and Pelletier (1986), show the surficial sediment to be thin, averaging 3 m, but varying from zero on bathymetric highs and tidal flats to 30 m in troughs. These earlier studies suggested that marine erosion dominates in many areas of the bay with low postglacial sedimentation. Limited stratigraphic analysis (Leslie, 1963, 1964, 1965) based primarily on foraminiferal abundance records a transition from glacially dominated to present marine conditions at approximately 8000 BP. In 1971, more detailed acoustic studies undertaken by the Geological Survey of Canada in conjunction with Aquitaine Company of Canada Limited (Lewis and Sanford, 1971) showed glacial landforms interpreted as ice-marginal accumulations and relict iceberg scours. The latter features were also described from wellsite surveys conducted for Canadian Occidental Petroleum by Geomarine Associates in water depths ranging from 135 to 185 m (Whittaker et al., 1985).

Leslie and Pelletier (1965), Pelletier et al. (1968), and Pelletier (1966,1969,1986), noted a correlation between seafloor morphology and the underlying geology as interpreted from echograms and seismic data (Grant,1969), and attributed it to a subaerially eroded drainage system developed prior to glaciation. This relationship was also observed by Lewis and Sanford (1971). Pelletier et al. (1968), Nelson (1968), Pelletier (1969, 1986), Bayliss et al. (1970) and Lewis and Sanford (1971) noted that the composition of the sediment reflected the underlying geology. Nelson (1968) used this observation to dismiss sea ice rafting as a major factor in sediment transport as suggested by Leslie (1964) and Pelletier (1969, 1986). Offshore extensions of glacial dispersal trains defined on land were recognized by Shilts (1982, 1986) and Henderson (1983, 1985) and suggests that much of the surficial material is glaciogenic. Adshead (1983a,b,c) showed that fluvial sediment entering southern and western Hudson Bay comprised glaciogenic debris of eastern provenance eroded from thick stratigraphic sequences exposed during isostatic rebound of the land.

METHODS

Approximatley 2800 km of continuous seismic reflection data were collected simultaneously on *CSS Hudson* cruise 86-040 (Fig.1) by: 16 kJ sparker with digital mutichannel (24) array for deep (2 s) bedrock penetration; 653 cm^3 airgun with analog hydrophones for intermediate penetration (1 s) and Huntec deep tow system (DTS) for high resolution (.30 m) and shallow penetration (100 ms). In addition,75 and 50 kHz. sidescan data were collected continuously where water depth was less than 250 m.

Navigation was provided by a combination of global positioning and satellite navigation. Satellite fixing accuracy was substantially improved by use of a doppler sonar to provide accurate speed over the bottom, thereby increasing the accuracy of dead reckoning between satellite passes. Navigational errors were generally less than 20 m. Bottom samples comprising two piston cores,three pilot cores and two grab samples were collected (Fig. 1) when rough weather prevented collection of seismic data.

The base of Quaternary deposits was interpreted from airgun seismic and Huntec DTS data.Subdivision of the Quaternary deposits into acoustic stratigraphic units was based on Huntec DTS data.In addition 3.5 kHz subbottom data collected on earlier cruises (Earth & Ocean Research Limited, internal data report, 1986.) were used to extend these interpretations laterally where the 3.5 kHz data crossed Huntec DTS coverage (Fig.1).

REGIONAL SETTING

Hudson Bay and James Bay cover a total of 1.3 million square kilometres (Martini 1986). Hudson Bay is 950 km at it's widest point and forms a generally shallow (<250m), saucer shaped basin that deepens to 500m at the northeastern margin where it joins Hudson Strait (Fig. 2). The deeper portions are formed by the Winisk trough and Hudson Basin which merge north of the Midbay Bank (Sanford, and Grant, 1976) and trend northeasterly toward Hudson Strait. These basins contain isolated infilled channels in their deepest portions which we interpret to indicate existence of an earlier fluvially cut drainage system. An unconformable surface on top of the fluvial sequence is interpreted to be the result of glacial erosion and subsequent overdeepening along these channels. Seabed relief is subdued but variable with slopes generally less than 2 degrees. Regionally the topography is controlled by glacial overdeepening of pre-existing drainage channels and the erodability of the underlying (Paleozoic) bedrock. On a smaller scale, till ridges up to 15 m high and 300 m wide dominate the relief of the nearshore areas. Ice-keel scour marks with berms up to 3 m high form the dominant relief of deeper portions of Hudson Bay.

RESULTS

Isopach map

The Quaternary sediments isopach map (Fig. 3) is based on the acoustic coverage in Figure 1. The isopach map indicates a generally thin (<5m) sediment veneer for most of Hudson Bay. Maximum observed thicknesses of 21 m and

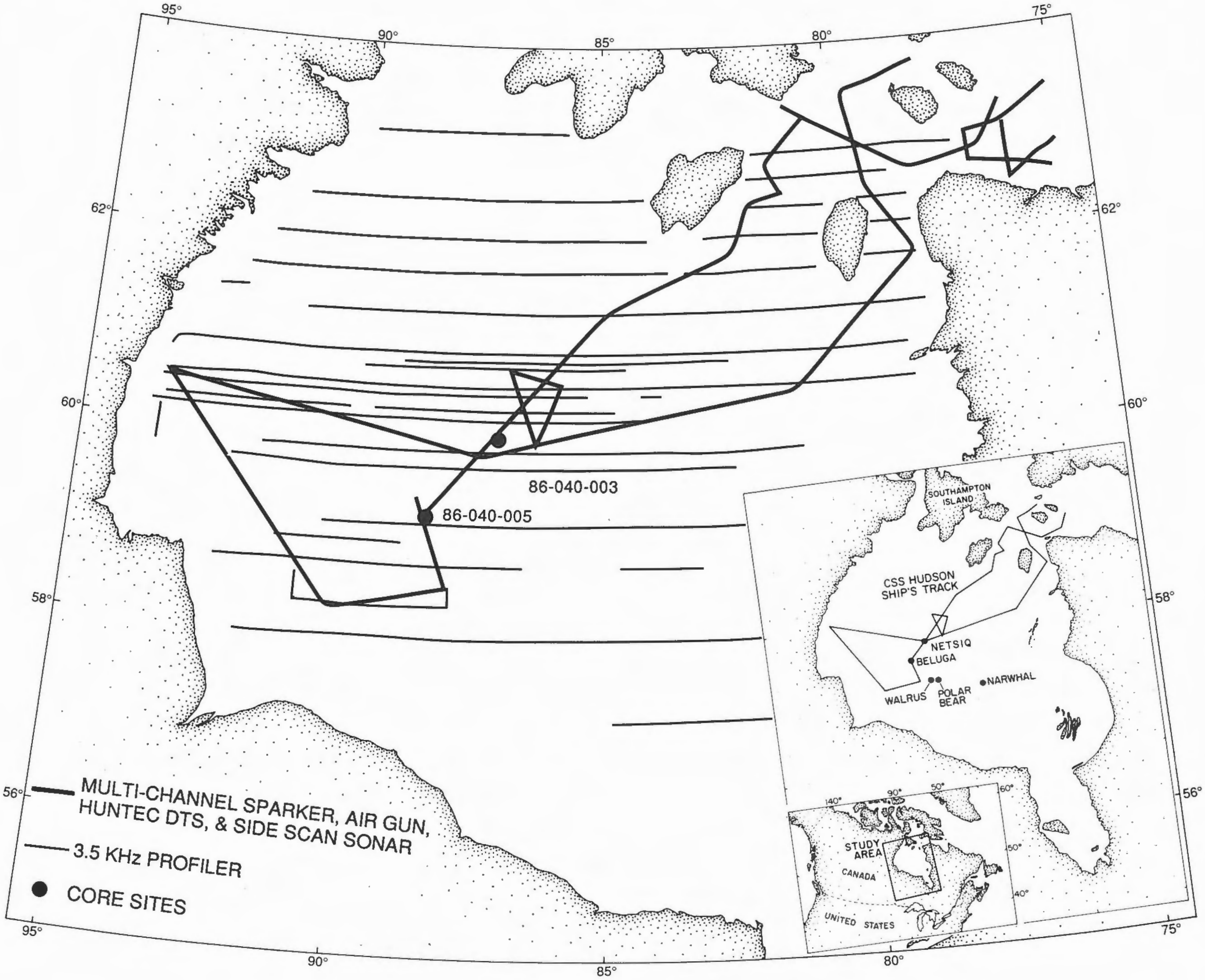


Figure 1. Location map showing seismic and sample control described in text.

15 m were observed in the Winisk trough and on the western margin of Mansel Bank respectivley. In western Hudson Strait the maximum thicknesses of up to 170 m are indicated as spot values in Figure 3.

Acoustic stratigraphic unit map

The Quaternary sediments overlying bedrock are subdivided on the basis of acoustic character as defined by the high resolution Huntec DTS and 3.5 kHz systems (Fig. 4). Acoustic stratigraphic units are chosen on the basis of their similarity to established units from other well-studied areas; ie. Hudson Strait (MacLean et al., 1986), Labrador Shelf (Josenhans et al., 1986) and Scotian Shelf (King and Fader, 1986). Although extensive sampling along acoustic transects is required before definitive correlation of acoustic properties with geological attributes is possible, the similarity in acoustic properties with those of the Scotian Shelf to Hudson Strait warrants correlation with geological features.

Three acoustic stratigraphic units (in various combinations-see Fig. 4) make up the Quaternary sediments of Hudson Bay. These units are designated as: glacial tills (unit 3); glaciomarine stratified sediments (unit 4) and postglacial basinal muds (unit 5). Acoustically, the glacial till unit appears homogenous and unstratified, typically with a smooth base at the bedrock unconformity and an undulating upper surface. The unit varies in thickness from <1-55 m. The glaciomarine unit is well stratified conformably draped deposit mimicking the underlying till surface. It is generally <5m thick although thicker accumulations are present in Hudson Strait and Winisk Trough. The post glacial muds are limited in extent and thickness (<5m) and typically occur near river mouths. They are characterized acoustically by a lack of strong internal reflectors and occur only in localized depressions.

The distribution and acoustic character of these units has been disturbed in places by grounded ice during and after deglaciation.

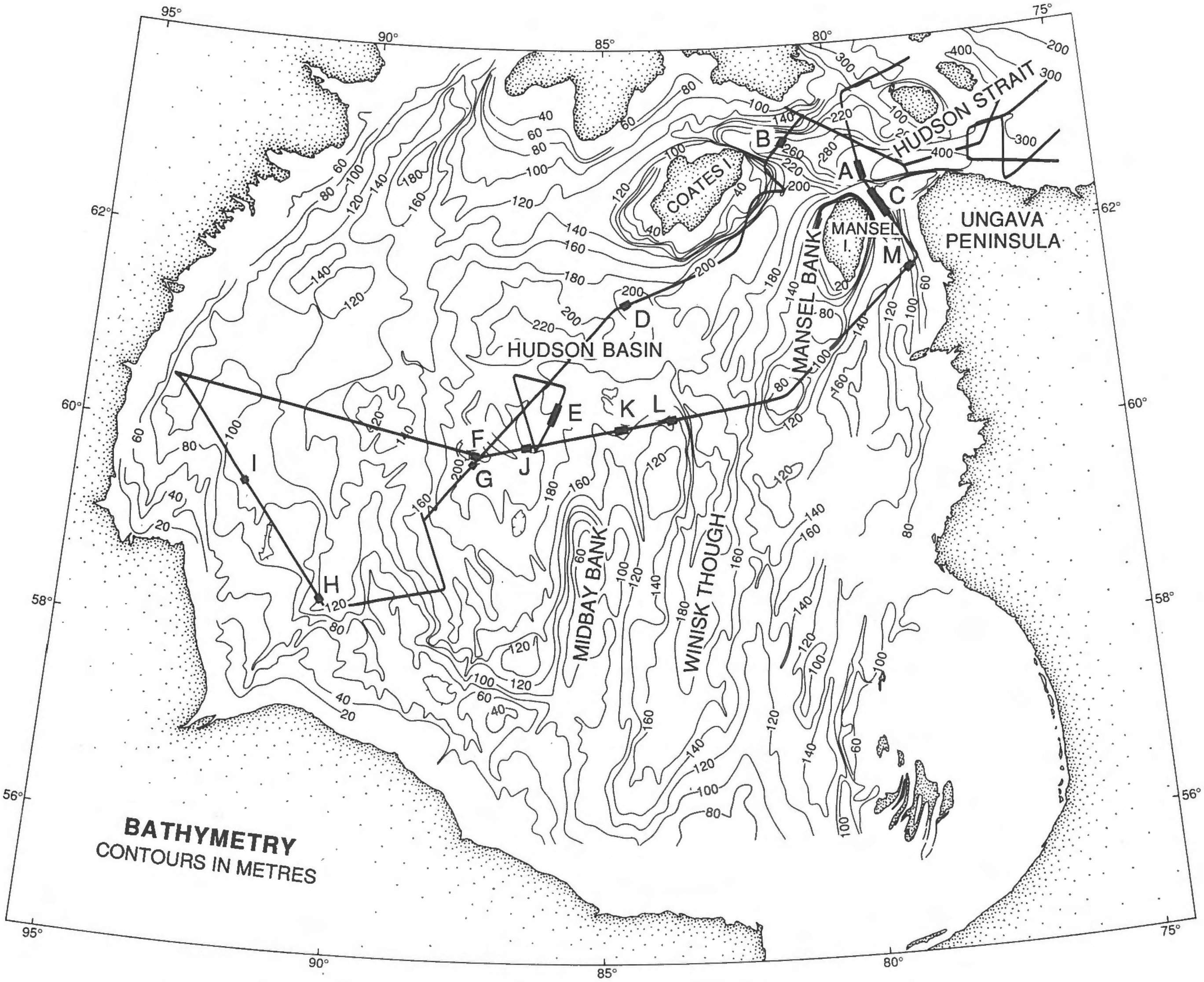


Figure 2. Bathymetry map of Hudson Bay showing the place names and locations of sections described in text.

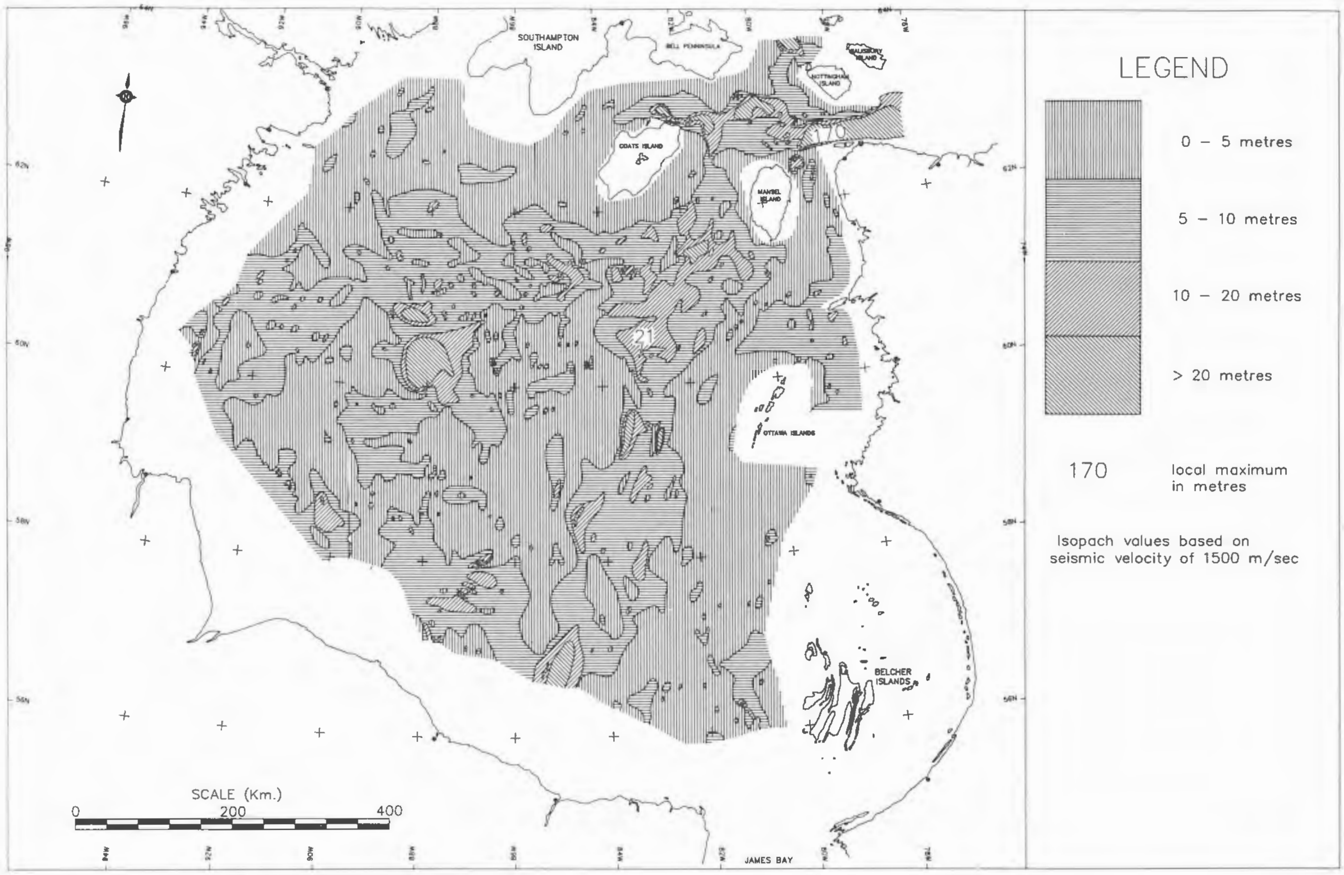
Description of map units

Figure 4 illustrates the outcrop of zones of acoustic stratigraphic units as detected within the resolution of the Huntec DTS (30 cm) and 3.5 kHz (100cm) systems. Designation of the units is preliminary and dictated by acoustic characteristics that are representative of broad areas.

Unit 1 represents areas of Precambrian bedrock which locally includes an undivided veneer of glacial and postglacial sediments. Unit 2 represents outcrops of Paleozoic and younger bedrock. Unit 3 represents areas of till exposed at the seafloor. The unit has a consistently homogenous, unstratified acoustic character from seafloor to the bedrock unconformity. The unit is tentatively interpreted to be basal till. Up to four superimposed tills separated by well developed acoustic horizons can be recognized locally. We infer that the uppermost of these tills was deposited during the last (Late Wisconsinan) glaciation. Unit 4 represents areas of well

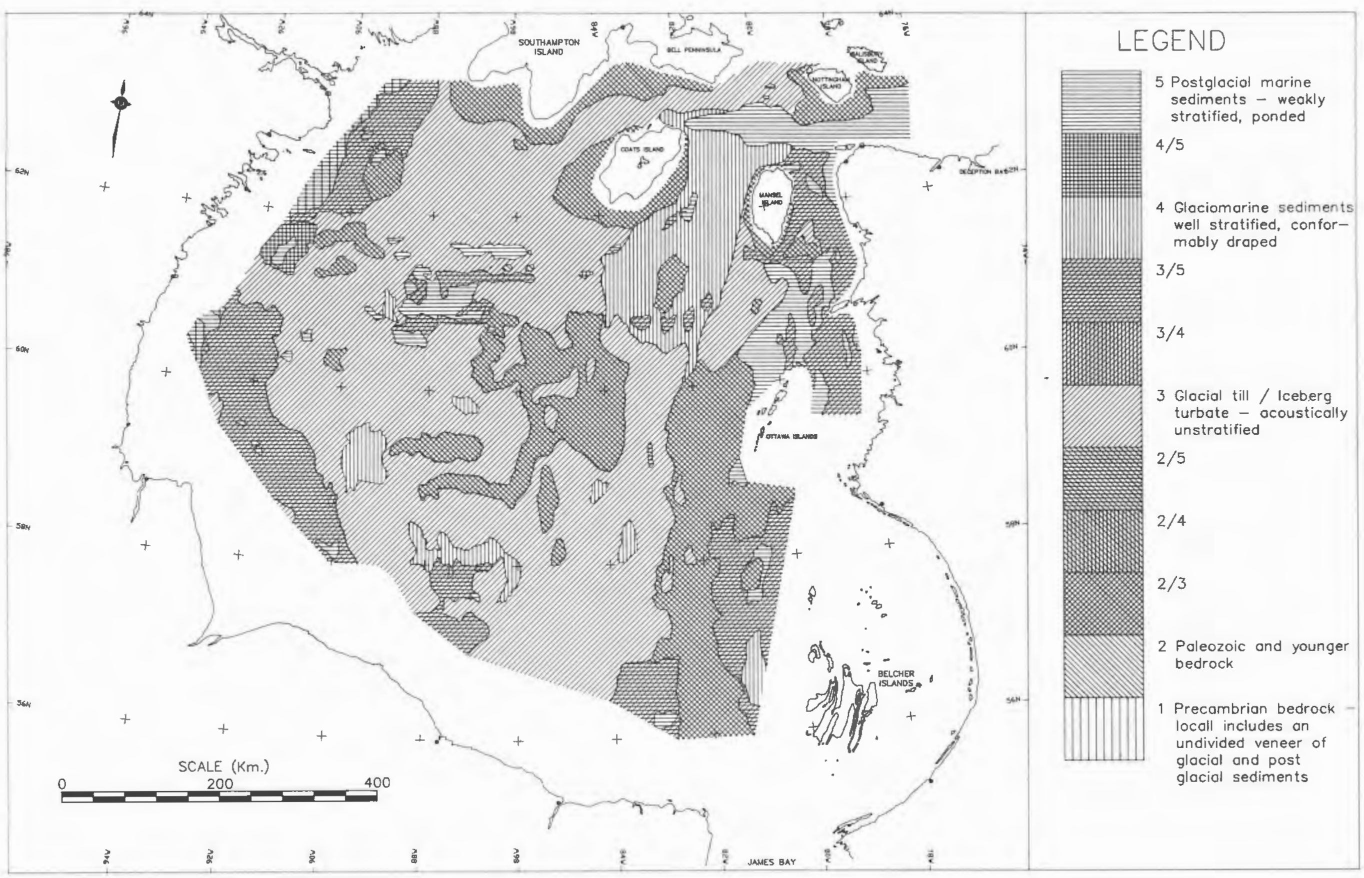
Figure 3. (opposite, above) Isopach map of Hudson Bay Quaternary sediments based on a combination of Airgun and Huntec D.T.S. seismic data obtained on *CSS Hudson* cruises 86-040 and 87-028 and on 3.5kHz data collected in 1977-78 by the Canadian Hydrographic Survey. Numbers indicate spot depths in metres.

Figure 4. (opposite, below) Distribution of surficial acoustic unit based on preliminary interpretation of the Huntec D.T.S. and 3.5kHz seismic data (compiled by S.Balzer). Fractional values (eg.4/5) represent areas where discontinuous patches of postglacial marine sediments (unit 5) overlie glaciomarine sediments (unit 4).



Lambert Conformal Projection–Standard Parallels 56 05'N,62 25'N

HUDSON BAY TOTAL QUATERNARY ISOPACH



Lambert Conformal Projection–Standard Parallels 56 05'N,62 25'N

HUDSON BAY SURFICIAL UNIT OUTCROP

stratified conformably draped sediments which mimic the till surface. In places the unit is heavily incised by ice keel scours. The unit is thought to have been deposited in ice proximal conditions. Unit 5 represents areas of weakly stratified muddy sediments which occur preferentially at river mouths, in localized topographic depressions and within deep basins of Hudson Strait.

Description of seismic sections (see Appendix)

Seismic sections A-K illustrate some of the key stratigraphic and geomorphic features observed during the Hudson 86 cruise. Locations and place names are shown in Figure 2.

Section A

Typical acoustic character of the till, glaciomarine and postglacial sediments at the entrance to Hudson Bay. Note the marked unconformity with the underlying (bedrock) ridge. The acoustic section north of the (bedrock) ridge is typical of the deposits in western Hudson Strait and comprises three units interpreted to represent: 10m of till (unit 3); 6m of conformably draped glaciomarine sediments (unit 4) overlain by 8m of postglacial weakly stratified muds (unit 5). The horizontally stratified ridge may be an erosional remnant that was protected from glacial erosion by the presence of Mansel island.

Section B

The Huntec section is interpreted to represent an ice lift off moraine (King and Fader, 1986) marking the former buoyancy line (at 295m, present water depth) of an ice margin in western Hudson Strait. The acoustically stratified (glaciomarine) sediments seaward (north east) of the till tongue are thinner than in section A and may represent a later ice margin which had less time to accumulate sediment. We suggest this moraine was built at a time (about 8.5ka, J. Stravers, pers. comm.) when the remnants of the Laurentide Ice sheet were pinned by Mansel and Coates islands and an embayment in Hudson Strait had penetrated as far westward as the ice lift off in section B.

Section C

The section is located between Mansel Island and western Ungava peninsula. The morainal deposit in the southeastern part of the section and the acoustically stratified sediments to the northwest may represent an ice contact (till) and proglacial sediment sequence indicative of an ice marginal position. The thick sequence of conformably draped glaciomarine sediments suggests that an ice margin was stationary in this area for a longer time than at section B.

Section D

Ice lift off moraine in central Hudson Bay similar to section B. Note that the buoyancy line (at 220m) is 75m shallower than in section B. The shallower liftoff suggests ice thinning during retreat. The thin sequence of glaciomarine sediments seaward of the till deposit may indicate rapid retreat of the ice margin.

Section E

Interpreted Huntec and air gun section showing the well bedded (Paleozoic) bedrock overlain by weakly stratified cross bedded fluvial deposits (of unknown age) overlain by an unstratified acoustically massive unit designated as till. Note the well developed unconformity between the till and underlying deposits which allows for clear definition of overburden thickness. This till thickness and morphology is typical of >50 % of the bay.

Section F

Composite sidescan and huntec diagram representative of the deep areas in south-central Hudson Bay. The huntec profile

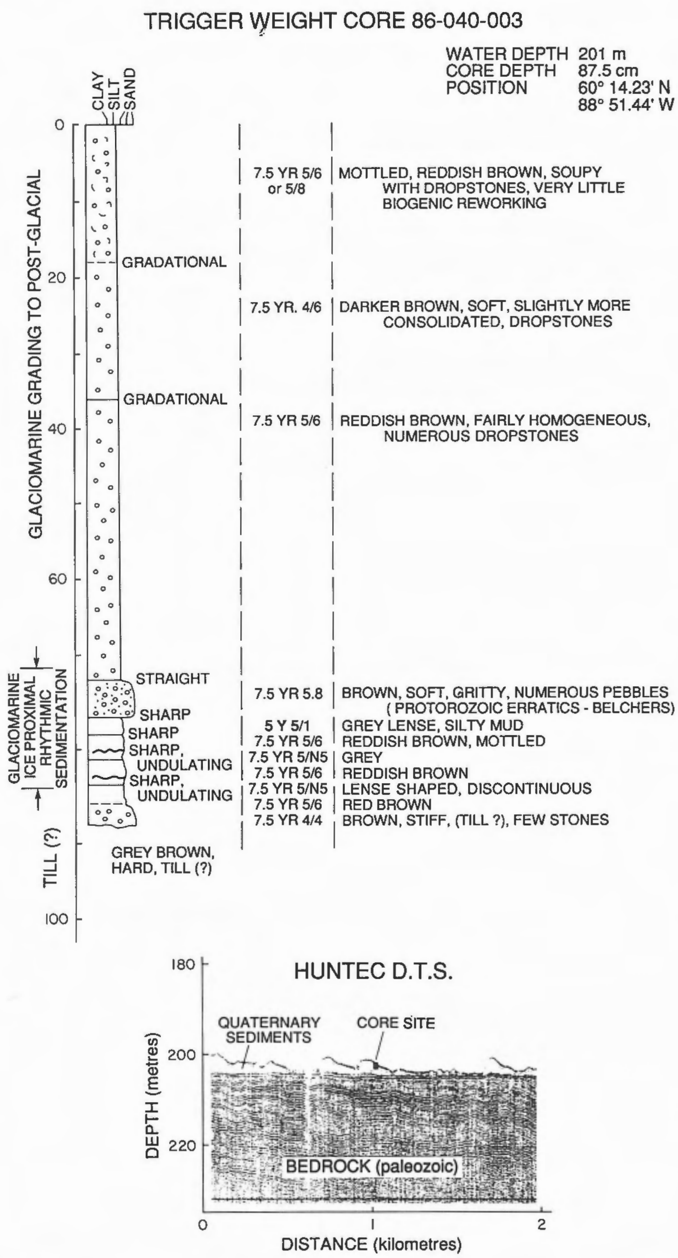


Figure 5. Description of piston core from Hudson Basin with stratigraphic location of core noted in Huntec D.T.S. profile. Location of core is shown in Figure 1.

shows two discontinuous till units up to 10 m thick over bedrock which are in turn overlain by a thin (1-3m) discontinuous reworked veneer of glaciomarine sediments (unit 4). The massive unstratified surficial unit may be representative of till or ice-keel reworked unit 4. The crosscutting sidescan lines show an unusual pattern of wide, elipsoid ice-keel scour marks with a preferred north-south orientation. The shape is thought to be imparted by eliptical tidally driven currents moving large tabular icebergs. The lateral extent of these elipsoid scours is indicated in Figure 6.

Section G

Huntec profile showing two superimposed tills overlain by a thin veneer of reworked glaciomarine sediments or uppermost till. Multiple till sequences as illustrated occur more frequently in the south-central portion of the bay. They may indicate preservation of earlier glacial deposits at the base of the former Laurentide ice sheet.

Section H

Sidescan and Huntec profile showing relief similar to subglacial dead ice topography or stagnation moraine. The surface is developed on a veneer of till over bedrock. The area is thought to be unscoured due to the influence of protective ice cover. The mounds and ridges are similar to the subglacial features reported off Spitsbergen by Solheim and Pfirman (1985) and may have formed at the glacier sole. Alternativley they may be interpreted as dead ice topography similar to that reported by Shilts et al. (1987) from the vicinity of the Keewatin Ice Divide.

Section I

Huntec and sidescan profile of well stratified bedrock overlain by till ridges with postglacial sediments within the depressions. The sidescan profile illustrates the lateral extent and parallel nature of the till ridges. In the southwestern area of the bay, the dominant orientation of the till ridges is 025 degrees. In addition to the large till ridges, smaller ones having an orientation of 080 degress occur in the area north of Churchill (Fig. 6).

Section J

Huntec and sidescan profile showing subglacial and ice marginal flute markings. The Huntec profile indicates a thin (< 5m) veneer of undulating acoustically unstratified material (till) overlying a smooth bedrock unconformity. The sidescan profile illustrates an area 1.5 x 5km which indicates that the surface of the till is moulded into parallel flutes bearing 340 degrees. These flutes are thought to be formed at the base of a moving ice mass flowing toward deeper water in a north westerly direction, although movement in the opposite direction cannot be discounted. Note that the features are mostly parallel throughout and change only slightly in character along track, presumably in response to an eroding ice base. Between these parallel flute ridges are smaller, ribbed features which are generally 45 degrees to the major flute direction. At the eastern and central areas of section J, a series of minor cross cutting ribs are developed on top of one of the fluted ridges. The origin of these minor ribs is not understood but may be due to cyclic lift off and touch down of the ice sheet in response to tidally influenced buoyancy. Alternativley, they may be similar to the rogen moraine developed on top of fluted terrain north of Dubawnt lake reported by Shilts et al. (1987).

Section K

Huntec and sidescan profile from northern Midbay Bank showing dead ice topography similar to section H. The Huntec profile indicates that the features are developed on a thin veneer of unstratified sediments interpreted to represent till. Most areas of Hudson Bay in similar water depth are heavily reworked by ice keel scour marks and the complete absence of scour marks here is interpreted to be due to protection by (late) grounded ice during times when bergs deep enough to scour this water depth (160m) were active in other parts of the bay.

Section L

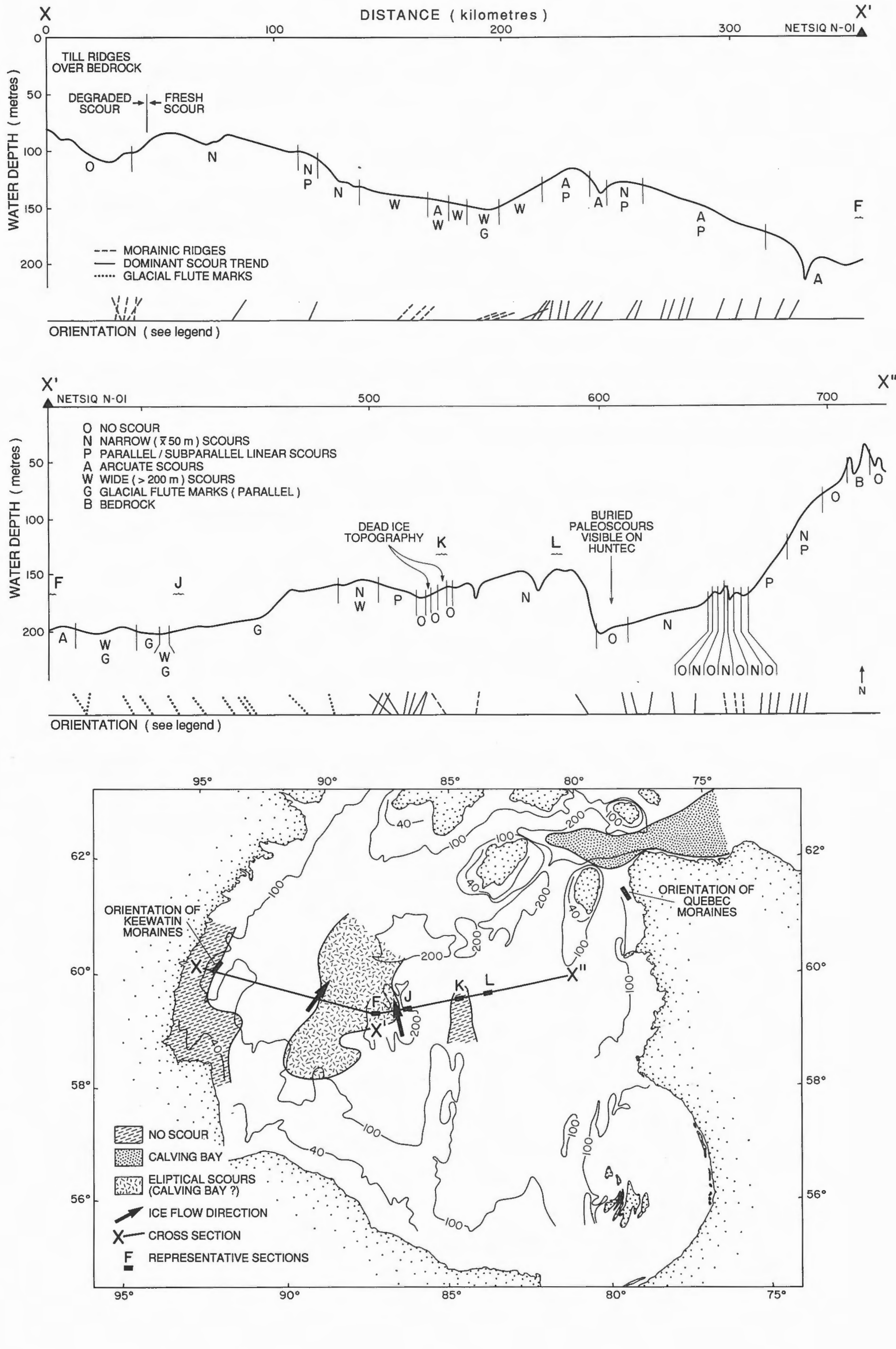
Sidescan Sonar profile showing the typical randomly oriented ice keel scour pattern observed over much of the bay. Note the orientation of this type of scour in comparison to the parallel ridges shown in section J. The random pattern is typical of unrestricted open water conditions in which icebergs drift under the variable influence of local currents and winds.

Section M

Huntec and sidescan profile from northeastern Hudson Bay showing a smooth bedrock surface overlain by till ridges and postglacial sediments ponded in the depressions. The till ridges have an orientation of 320 degrees. The section is similar to section I (from the southwestern side of the bay) and can be correlated with till ridges (DeGeer moraines) occurring on the adjacent Ungava peninsula. There, till ridges having a north-south orientation are defined as Degeer moraines (Prest et al., 1968) and extend to the present coastline. The slight difference in orientation between the onshore and offshore moraines (seperated by 30 km) may reflect local flow of ice along the maximum gradient into the channel between Mansel Island and Ungava peninsula.

BOTTOM SAMPLES

The location of three 10 m piston cores deployed in the area of deep iceberg scour is shown in Figure 1. These bottomed out in very stiff unsorted material interpreted to represent till or ice-keel turbate. Maximum penetration was 0.88 m. The seismostratigraphic setting and lithological interpretation of core 86-040-003 is shown in Figure 5. The bulk of the sediment collected in the core appears to represent a (< 1m) surface veneer not clearly resolved by the acoustic system. The sediment cores consist predominantly of massive to faintly laminated brown to grey silty mud with occasional pebbles. Mottling occurs in the upper units. The basal 17.5 cm of core 86-040-003 consists of distinctly banded sediments varying in texture and composition. Clasts taken from the brown-grey, hard till (?) sediment remaining in the core


X
DISTANCE (kilometres)
X'
NETSIQ N-01
TILL RIDGES OVER BEDROCK
DEGRADED SCOUR
FRESH SCOUR
WATER DEPTH (metres)
MORAINIC RIDGES
DOMINANT SCOUR TREND
GLACIAL FLUTE MARKS
ORIENTATION (see legend)
X'
X"
NETSIQ N-01
O NO SCOUR
N NARROW (x̄ 50 m) SCOURS
P PARALLEL / SUBPARALLEL LINEAR SCOURS
A ARCUATE SCOURS
W WIDE (> 200 m) SCOURS
G GLACIAL FLUTE MARKS (PARALLEL)
B BEDROCK
DEAD ICE TOPOGRAPHY
BURIED PALEOSCOURS VISIBLE ON HUNTEC
WATER DEPTH (metres)
ORIENTATION (see legend)
ORIENTATION OF KEEWATIN MORAINES
ORIENTATION OF QUEBEC MORAINES
NO SCOUR
CALVING BAY
ELIPTICAL SCOURS (CALVING BAY ?)
ICE FLOW DIRECTION
CROSS SECTION
REPRESENTATIVE SECTIONS

catcher of core 86-040-005 were faceted and striated and consisted of predominantly carbonate lithologies although one clast of Belcher Island provenance (crystalline, Precambrian) was recognized. There is a significant change in lithology and composition between the upper part of core 86-040-003 and the hard, dark grey, poorly sorted material at the base. The core is sandier at its base with (3 % gravel, 27 % sand, 50 % silt, 20 % clay) and has a significant proportion (8 %) of greywacke-argillite from Belcher Island Protorozoic rocks as well as Paleozoic carbonates (69 %), Archean crystalline (22 %) and Dubawnt lithologies (1 %). The Belcher island fragments indicate transport from the southeast whereas the Paleozoic fragments can be derived locally within the bay. The Dubawnt fragments indicate a northwesterly source from the District of Keewatin.

Surface sediment collected by Van Veen grab sampler at core site 86-040-003 consists of approximately 5 % sand and gravel, 32 % silt, and 63 % clay. Lithologies of material >2mm reflect predominantly Archean crystalline (59 %) and Paleozoic carbonate (39 %) terranes although minor Dubawnt (?) erratics (2 %) from the District of Keewatin were recognized.

The cores are interpreted to represent an upward transition from glaciomarine to marine sedimentation. The hard, coarse material at the base of the cores has characteristics of till. These include texture and striated and faceted clasts. However, the material has been highly scoured by icebergs and may represent turbated ice proximal glaciomarine sediment.

INTERPRETATION

In western Hudson Strait the combined glacial and postglacial sediments attain thicknesses up to 170 m. Glaciomarine and postglacial stratified sediments account for a maximum of 35 m of this sequence. In Hudson Bay the maximum observed thickness of glaciomarine and postglacial sediments is 15 m in the Winisk Trough and till thickness is generally less than 10 m, although moraines up to 55 m thick have been observed in the southern area. The thicker sediments in Hudson Strait are belived due in part to preferential deposition within the deeper basin. In addition the moraines illustrated in sections C and B may indicate an ice marginal position at the entrance to Hudson Bay (Fig. 6) which was stable for sufficient time to deposit these thick sediments. Extrapolation of the glaciomarine sediments along seismic horizons to nearby dated (marine) deposits on land (Lauriol ang Gray, 1987; Laymon, 1986), suggests that a restricted open water embayment, in which the stratified glaciomarine sediments were deposited existed as early as approximatley 10 ka. Dyke and Prest (1987) suggested this embayment developed later, at approximatley 8.4 ka, but the relatively thick sediment sequence in western Hudson Strait may support our interpretation of the earlier opening. Pinning of the ice sheet by Mansel and Coates islands may account for the lengthy still stand of the ice margin in western Hudson Strait, and the thick accumulation of glaciomarine sediments. In contrast the thin glaciomarine deposits illustrated in section D from Hudson basin suggest rapid retreat once the ice was free of the islands.

Figure 6. Interpreted cross section through Hudson Bay showing the major geomorphic zones and a summary of glacial landforms. In areas with superimposed geomorphic stipple patterns we interpret the upper one to have formed last. The inset map shows a summary of the interpreted ice flow directions and the lateral extent of eliptical scours.

If marine conditions in western Hudson Strait existed as early as 10 ka and the earliest marine sediments in southern Hudson Bay are approximately 8 ka (Dyke and Prest, 1987), then deglaciation of the bay must have occurred in 2 ka.

In view of the thin postglacial sediment accumulation in Hudson Bay, we suggest that the ice margin in western Hudson Strait persisted for much of the 2 ka and that once ice was free of the islands, retreat was very rapid.

In central Hudson Bay (section E, Fig. 3) the generally thin veneer of Quaternary deposits comprises primarily till. Toward the southern areas of the bay (section G) and particularly in the paleochannels thicker sections are preserved which we interpret to represent multiple tills. Troughout most of the bay it appears that each glaciation eroded earlier deposits, particularly in the central area. Alternatively they were never deposited. The smooth unconformity between the tills and the apparent absence of interglacial sediments indicates erosion during glacial episodes.

A unique elipsoid pattern of ice keel scours is preserved in southern Hudson Basin in water depth ranging from 160-200 m (section F). We suggest these scours formed in a calving bay dominated by tidal circulation. The lateral extent of these unique features is mapped (Fig. 6) on the basis of elipsoid scours observed on sidescan coverage. Some of these elipsoid patterns cover an area of about 2 km^2 and must have been formed by large icebergs with flat bases that impacted large areas of the seafloor. If as we suggest, the features were formed during deglaciation (i.e. 8.5ka), another 300 m must be added to the (paleo) water depth to account for postglacial sealevel lowering due to isostatic recovery. Thus we suggest the seafloor in southern Hudson Basin was intensely reworked during deglaciation by large tabular icebergs calved from an ice margin in up to 500 m depth.

Geomorphic (sidescan sonar) evidence from Midbay Bank and the shallow (120m) areas west of the Hudson Basin indicate well developed, parallel fluted terrain interpreted to have formed under moving ice.West of Hudson Basin we observe a single dominant ice flow direction bearing 025°. On western Midbay Bank the same (025°) trend is visible (section J) but truncated by overlying (and younger) flutes bearing 340 degrees. These flutes display a number of complicated ribbed patterns (section J) on the ridges which are not fully understood but may be due to vertical movement of the ice sheet in response to tidal variations. The fluted terrain strongly suggests that ice flowed parallel to the flutes. We suggest that topographic control caused the ice to flow downslope in a northerly direction toward the deep areas of the Hudson Basin. Because this is opposite to directions recorded on land south of the bay we suggest that this pattern reflects the last and local effects of the disintegration of the Laurentide Ice Sheet. Consequently the dominant dispersal direction of thicker more vigorous glaciations may not be recorded in the modern geomorphology of the bay.

Moraine ridges were observed nearshore in the western and eastern areas of the bay. In the western areas near Churchill, sidescan evidence (Section I) indicates dominant orientations of 025° cut by smaller ridges trending approximatley 080°. Seaward of the Ungava Peninsula moraine ridges trend 035° and these are correlated with DeGeer Moraines mapped onshore which have a north-south orientation (Prest et al., 1968). The difference in orientation is attributed to topographic steerage of the ice into the channel between Mansel Island and Ungava Peninsula. Ice flow patterns trending toward the maximum depth were observed throughout Hudson Bay.

An inventory of iceberg scours and subglacial morphologies recorded on an east-west transect across Hudson Bay are shown in Figure 6. Note that some areas of the bay are unscoured and preserve the original subglacial morphology even though nearby areas of similar water depth are completely reworked by scouring. We interpret this to indicate protection of the seafloor by late remnants of the Laurentide Ice Sheet which prevented icebergs from impacting the area. The diagram also illustrates areas where subglacial markings are slightly crosscut and modified by iceberg scour marks but the original surface is still clearly visible. This is interpreted to indicate rapid deglaciation of the bay with insufficient time for icebergs to completely obliterate the original glacial morphology.

On the northern margin of Midbay bank and in the southwestern area of Hudson Bay (Fig. 6.) sidescan data (sections H, k,) indicate evidence of sub-ice or dead ice topography. These areas are not iceberg scoured and may have been formed by late ice surges similar to those described off Spitsbergen by Solheim and Pfirman (1985). There, sidescan evidence obtained adjacent to modern surging glaciers in Svalbard shows a distinctive undulating rhombohedral pattern which is interpreted to reflect the morphology ofthe crevasse pattern at the base of the ice sheet once it had come to rest after surging. The terminal position of the Svalbard advance is marked by a well developed moraine ridge and numerous iceberg scours occur seaward of the moraine, but up-ice the seafloor is entirely formed of the rhombohedral ridges thought to have formed at the glacier sole. Disintegration of the ice sheet after the surge did not rework the rombohedral pattern, and likely occurred through calving of small fragments which had insufficient draft to impact the seafloor. The geomorphic pattern shown in sections H and K is similar to the subglacial relief but could also be interpreted as an ablation moraine resulting from vertical meltout of englacial debris.

The available data base indicates that glaciomarine sediments in Hudson Bay are thin and postglacial sediments are restricted to river mouths and localized depressions. On the basis of sedimentary patterns seafloor currents appear to be limited and seafloor disturbance by grounded ice has been restricted to the nearshore areas since deglaciation. Consequently the seafloor morphology preserved in Hudson Bay is a pristine indicator of the glacial and deglacial processes which occurred during the final disintegration of the Laurentide ice mass.

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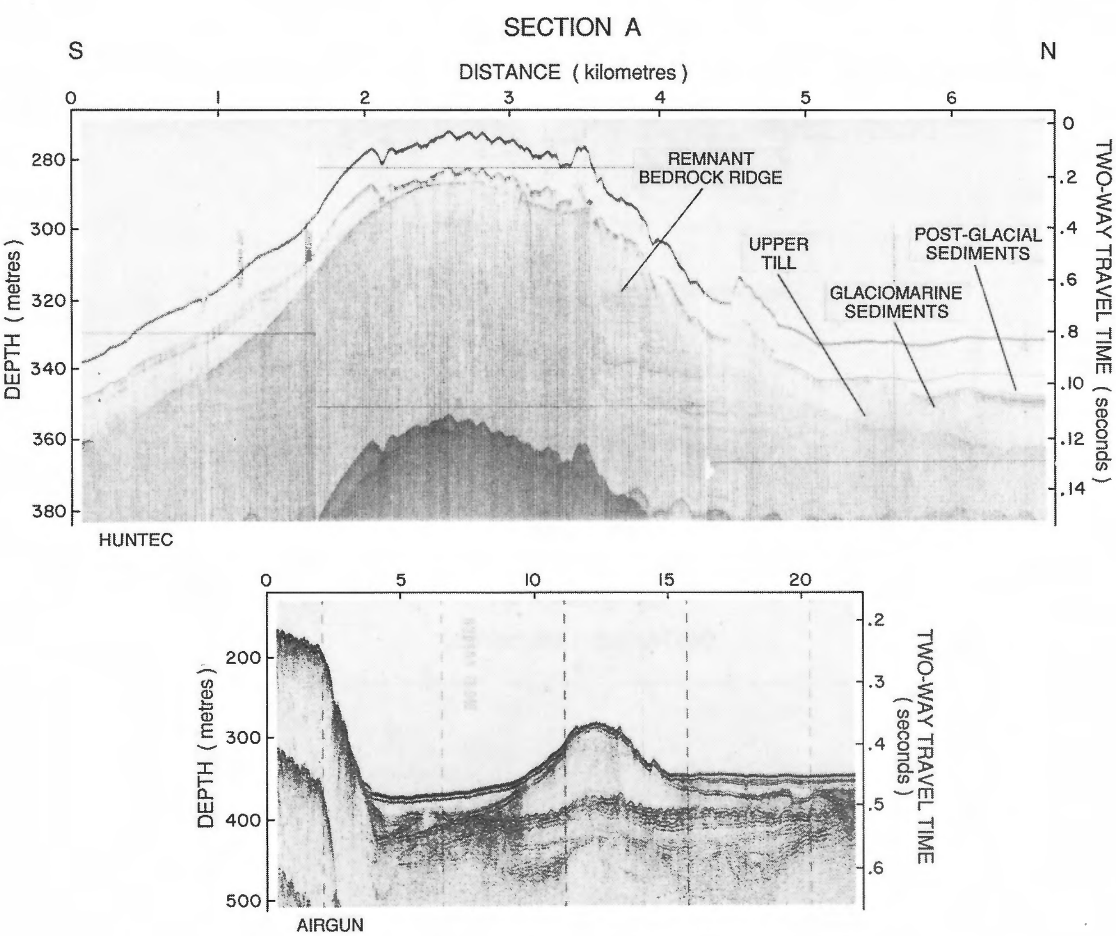
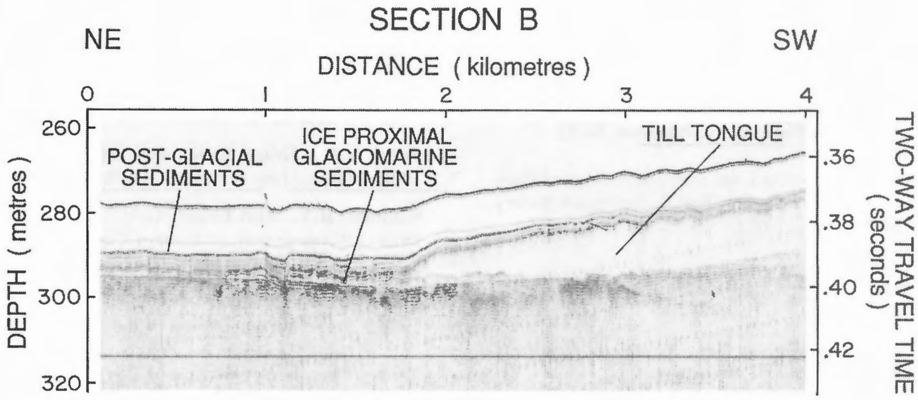
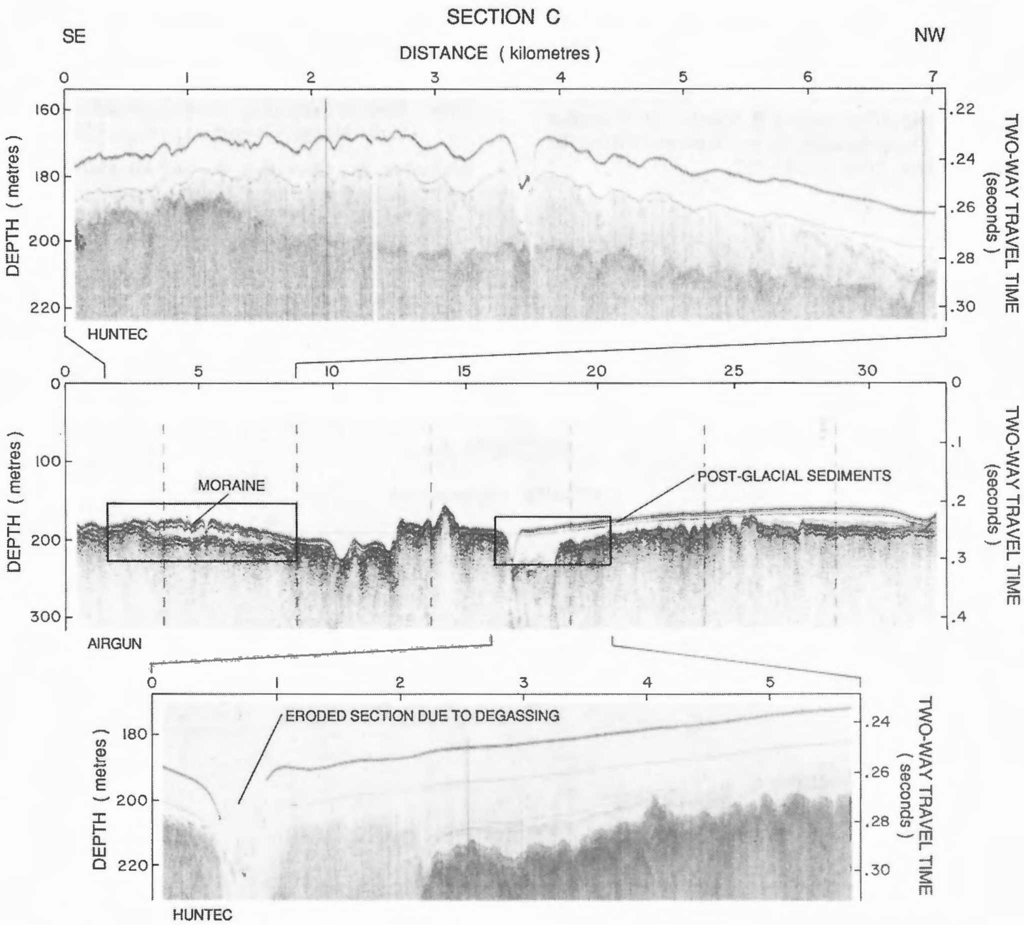
Appendix: Sections A-M. See text for descriptions.




SECTION B
NE
SW
DISTANCE (kilometres)
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)
POST-GLACIAL SEDIMENTS
ICE PROXIMAL GLACIOMARINE SEDIMENTS
TILL TONGUE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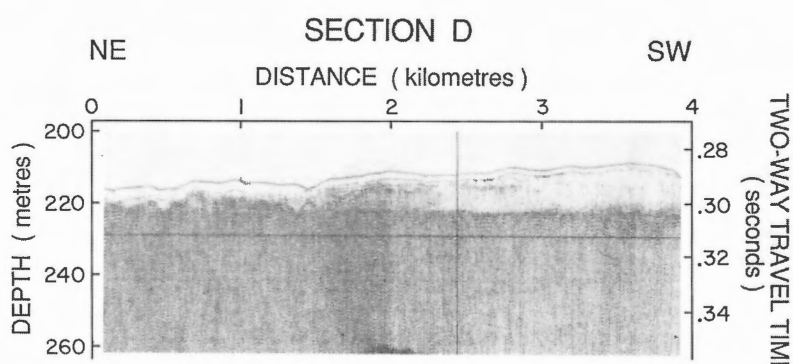
SE
NW
DISTANCE (kilometres)
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)
HUNTEC
MORAINE
POST-GLACIAL SEDIMENTS
AIRGUN
ERODED SECTION DUE TO DEGASSING
HUNTEC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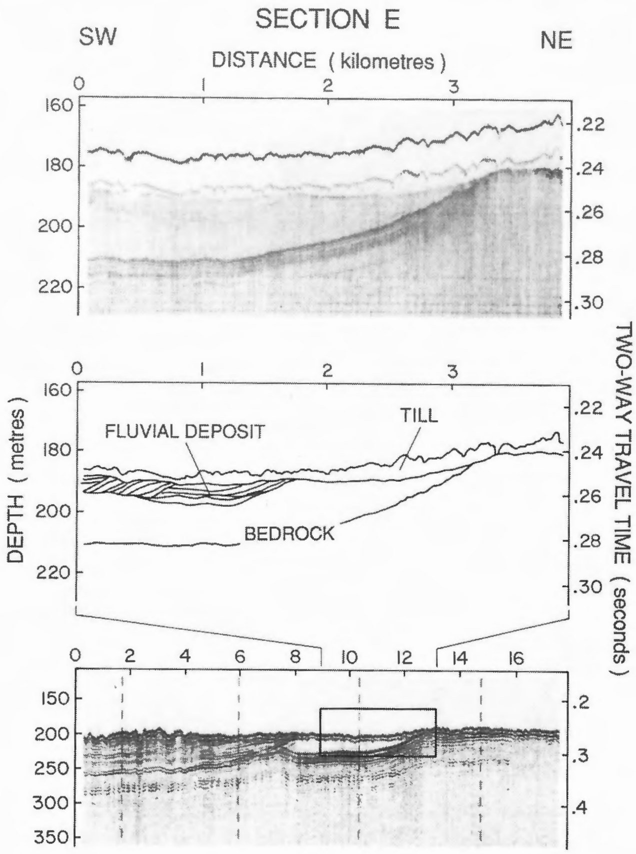
NE
SW
DISTANCE (kilometres)
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)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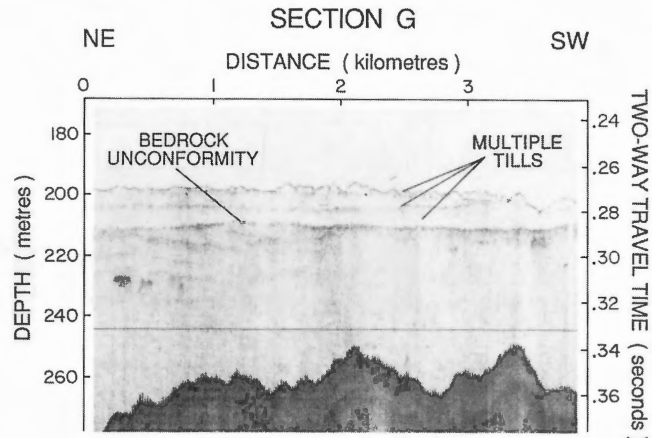
SW
NE
DISTANCE (kilometres)
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)
FLUVIAL DEPOSIT
TILL
BEDROCK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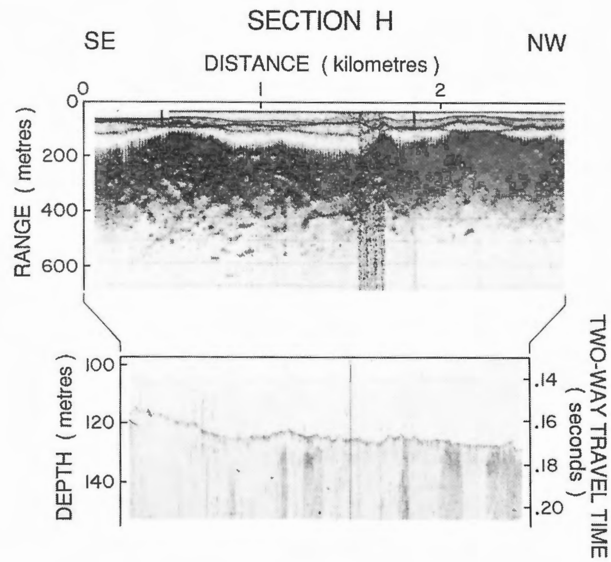
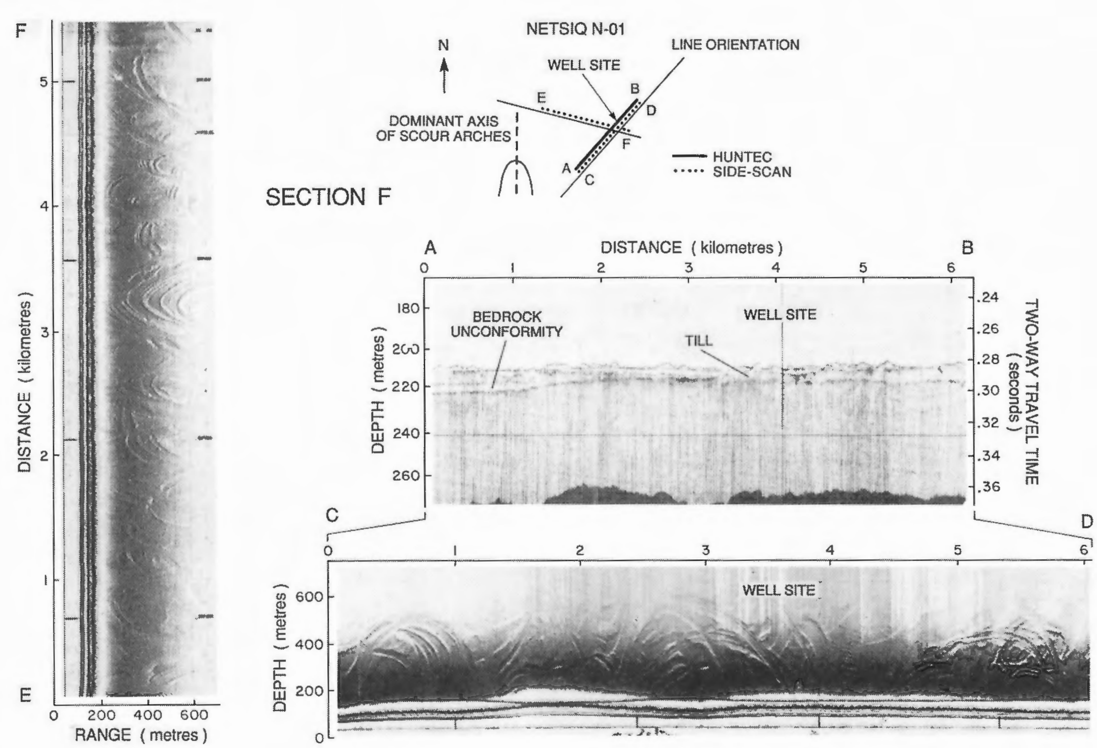
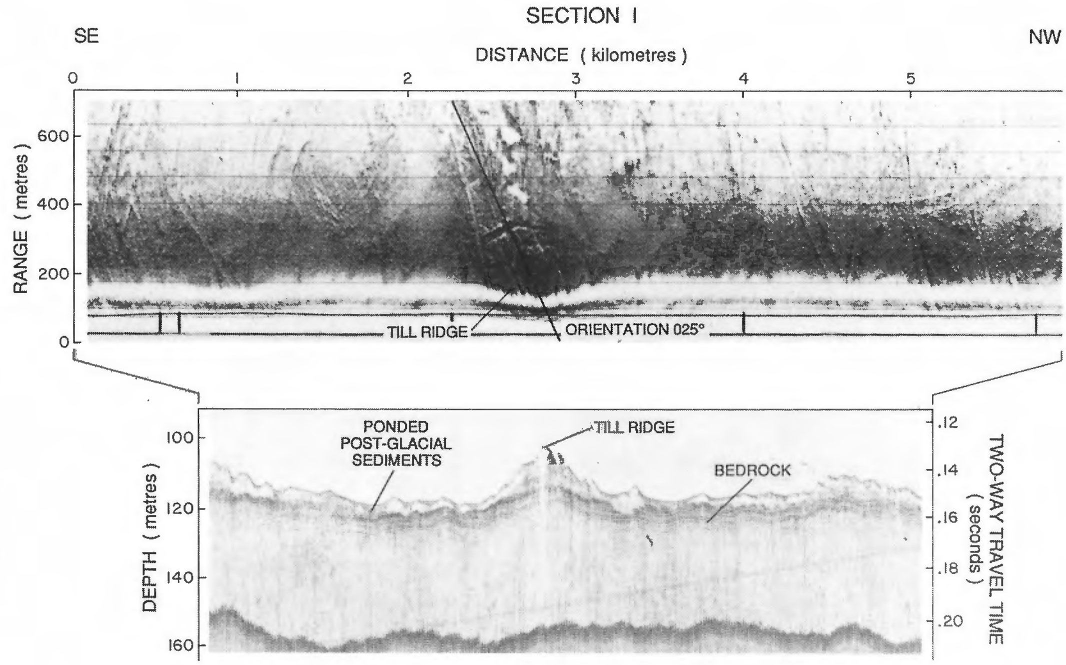
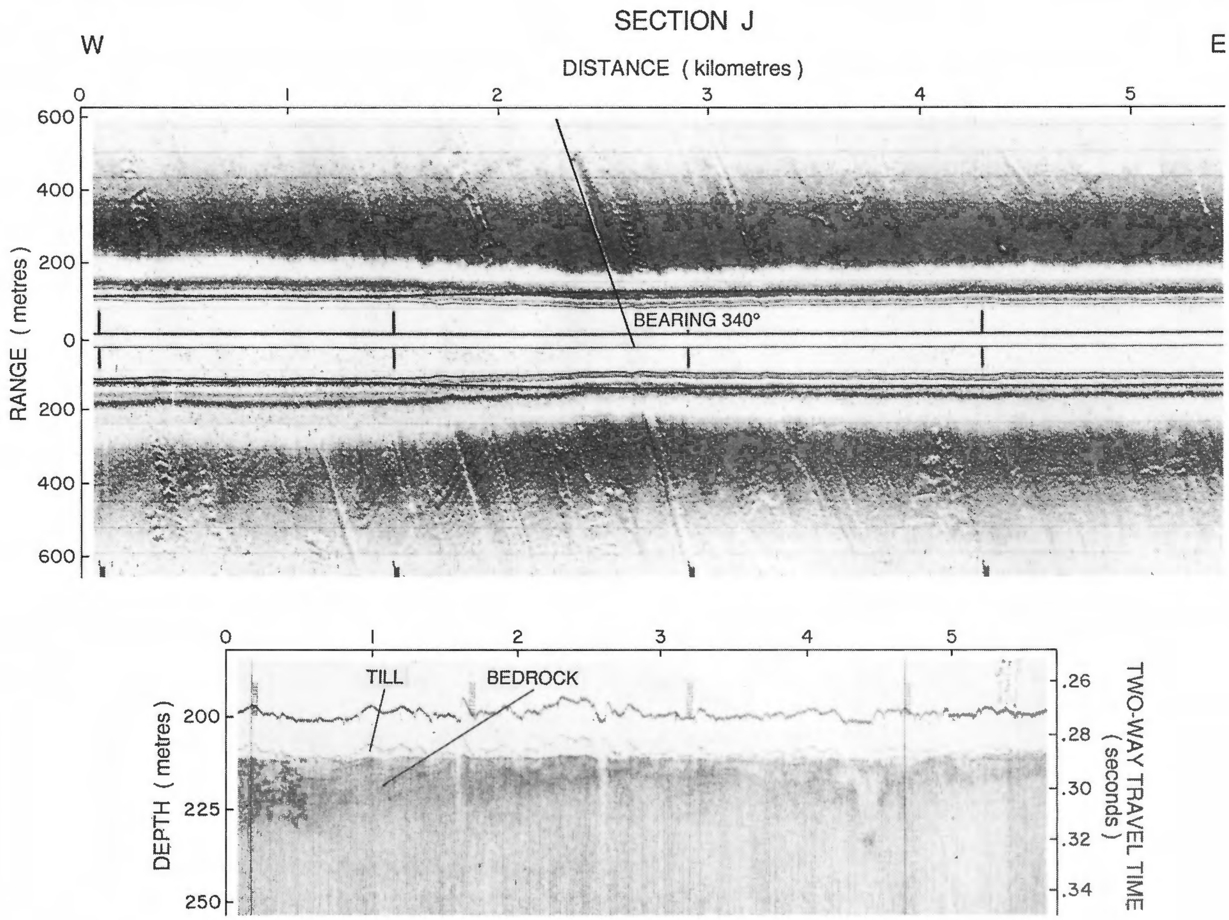
NE
SW
DISTANCE (kilometres)
BEDROCK UNCONFORMITY
MULTIPLE TILLS
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)

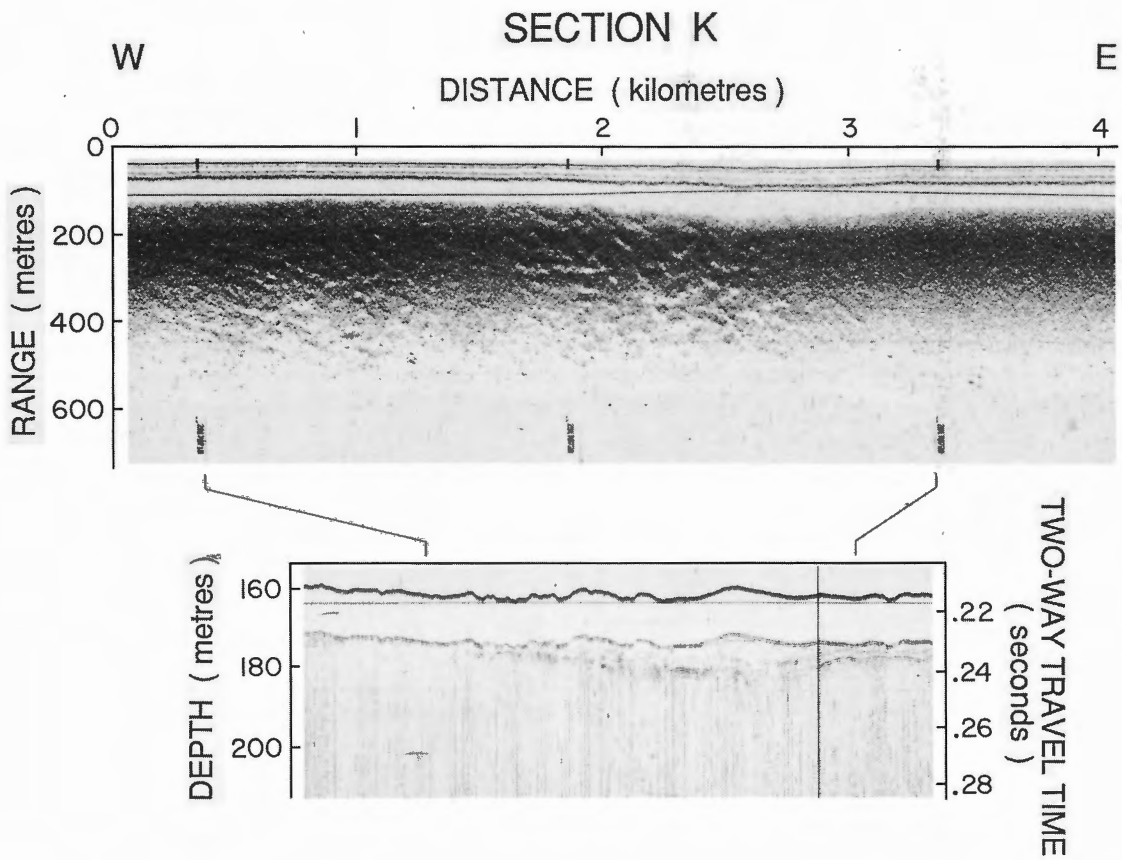

SECTION H
SE
NW
DISTANCE (kilometres)
RANGE (metres)
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)

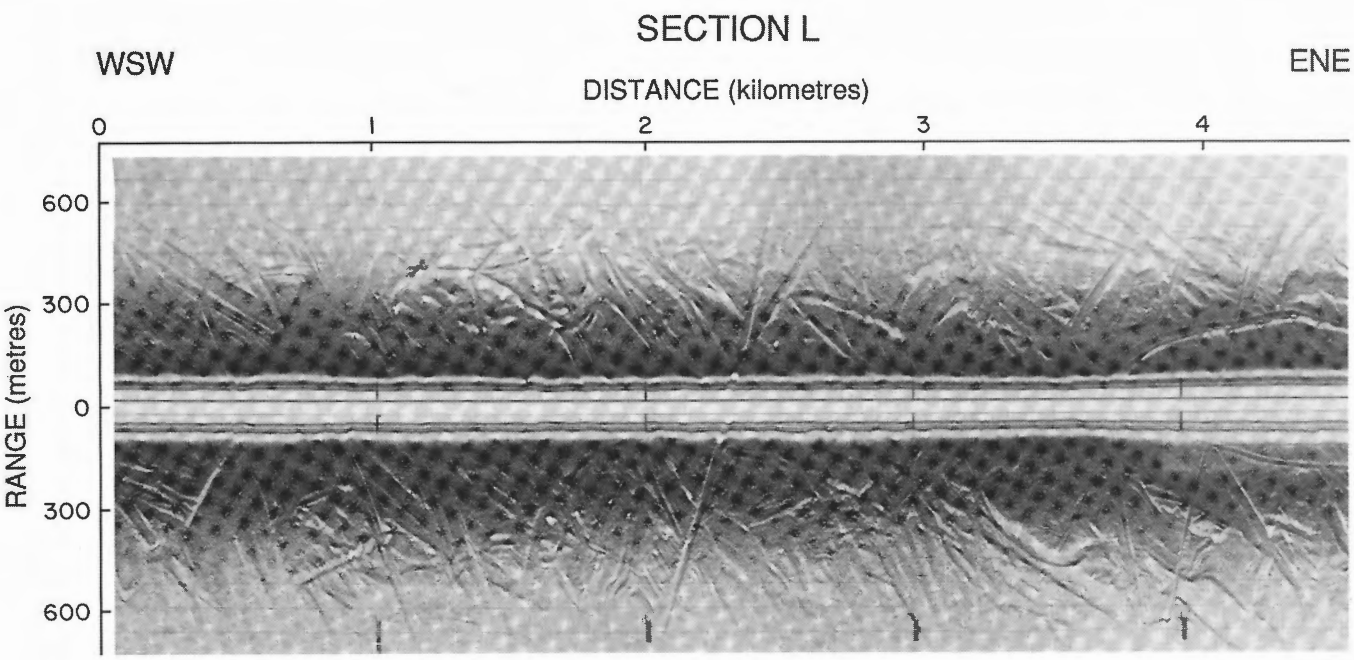
SECTION F


F
E
DISTANCE (kilometres)
RANGE (metres)
NETSIQ N-01
N
WELL SITE
LINE ORIENTATION
DOMINANT AXIS OF SCOUR ARCHES
HUNTEC
SIDE-SCAN
SECTION F
A
B
DISTANCE (kilometres)
BEDROCK UNCONFORMITY
WELL SITE
TILL
DEPTH (metres)
TWO-WAY TRAVEL TIME (seconds)
C
D
WELL SITE
DEPTH (metres)


SECTION I
SE
NW
DISTANCE (kilometres)
0
1
2
3
4
5
RANGE (metres)
600
400
200
0
TILL RIDGE
ORIENTATION 025°
PONDED
POST-GLACIAL
SEDIMENTS
TILL RIDGE
BEDROCK
DEPTH (metres)
100
120
140
160
TWO-WAY TRAVEL TIME (seconds)
.12
.14
.16
.18
.20


SECTION J
W
E
DISTANCE (kilometres)
0
1
2
3
4
5
RANGE (metres)
600
400
200
0
200
400
600
BEARING 340°
TILL
BEDROCK
DEPTH (metres)
200
225
250
TWO-WAY TRAVEL TIME (seconds)
.26
.28
.30
.32
.34


SECTION K
W
E
DISTANCE (kilometres)
0
1
2
3
4
0
200
400
600
RANGE (metres)
DEPTH (metres)
160
180
200
TWO-WAY TRAVEL TIME (seconds)
.22
.24
.26
.28


SECTION L
WSW
ENE
DISTANCE (kilometres)
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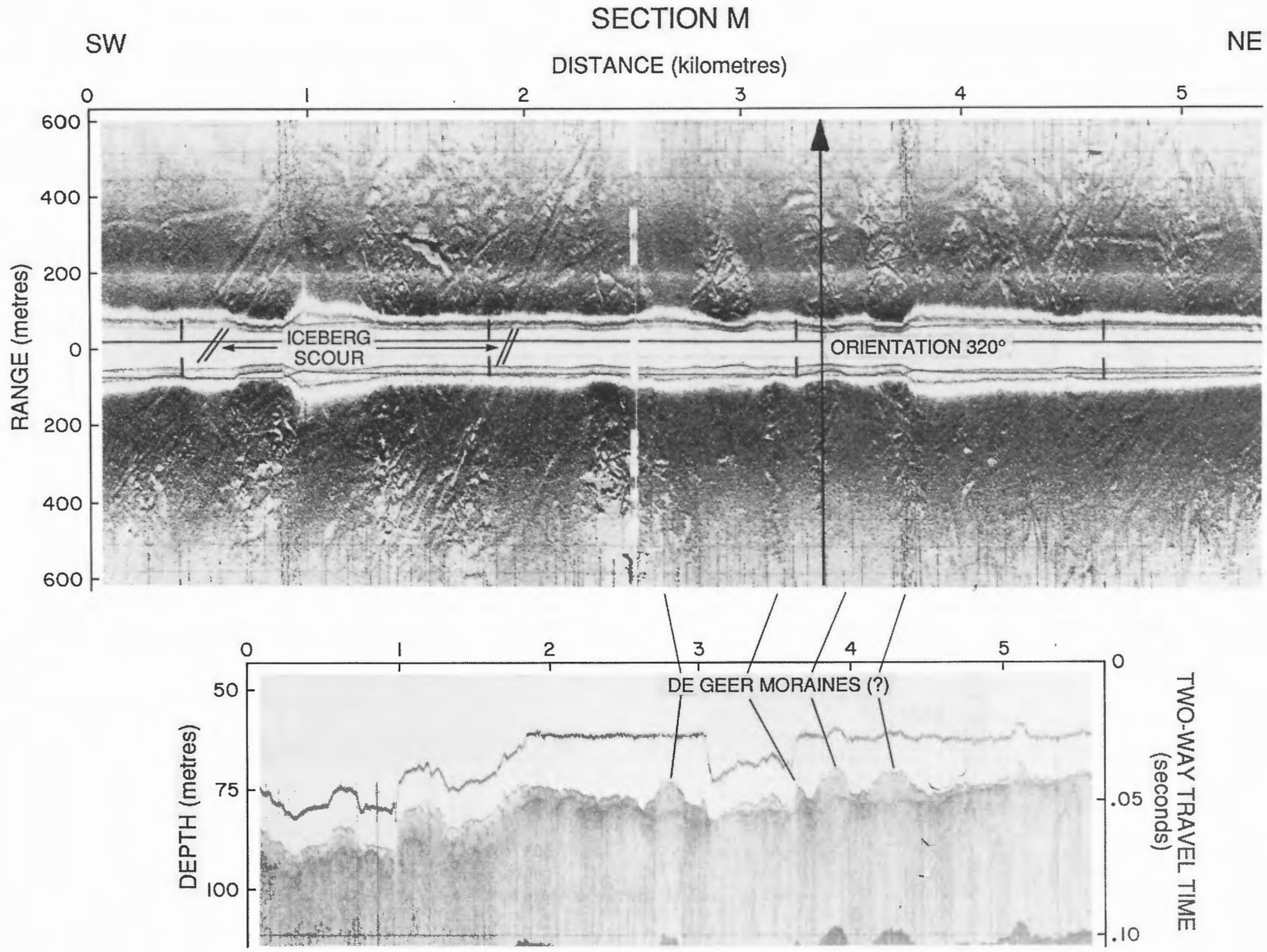
1
2
3
4
600
300
0
300
600
RANGE (metres)


SECTION M
SW
NE
DISTANCE (kilometres)
0
1
2
3
4
5
600
400
200
0
200
400
600
RANGE (metres)
ICEBERG
SCOUR
ORIENTATION 320°
DE GEER MORAINES (?)
50
75
100
DEPTH (metres)
0
.05
.10
TWO-WAY TRAVEL TIME
(seconds)

Bedrock geological mapping and basin studies in the Hudson Bay region[1]

A.C. Grant[2] and B.V. Sanford[3]

Grant, A.C. and Sanford, B.V., Bedrock geological mapping and basin studies in the Hudson Bay region; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 287-296, 1988.

Abstract

High resolution reflection seismic profiles obtained in 1986 and 1987 in Hudson Bay and adjacent channels to the north provide improved definition of subsurface structure and subsea boundaries of Paleozoic bedrock units. The new data provide particularly good resolution of Lower Silurian and Middle Devonian reefal units and of collapse structures related to dissolution of Silurian and Devonian evaporites. In fault-bounded subbasins beneath Foxe Channel and Evans Strait, the Phanerozoic succession may be 2 km or more thick roughly comparable to the thickness recorded by borehole drilling in the deepest part of the much larger Hudson basin to the south.

Fault-bounded blocks, salt dissolution structures and reefal facies that have wide occurrence in the Hudson Platform have many characteristics in common with structures that contain oil and gas in the Michigan Basin.

Résumé

Les profils obtenus par levés de sismique-réflexion à haute résolution, en 1986 et 1987 dans la baie d'Hudson et dans les détroits adjacents au nord, nous ont permis de mieux définir la structure de subsurface et les limites sous-marines des unités du socle paléozoïque. Les nouvelles données permettent une résolution particulièrement poussée des unités récifales du Silurien inférieur et du Dovénien moyen, ainsi que des déformations par glissement résultant de la dissolution des évaporites siluriennes et dévoniennes. Dans les bassins secondaires limités par des failles, qui se trouvent sous le détroit de Foxe et le détroit d'Evans, la succession phanérozoïque peut atteindre une puissance de 2 km ou davantage, à peu près comparable à celle indiquée par les trous de sondage forés dans la partie la plus profonde du bassin de la baie d'Hudson, beaucoup plus vaste, située au sud.

Les blocs faillés, les structures créées par la dissolution du sel et les faciès récifaux qui apparaissent fréquemment sur la plate-forme d'Hudson, présentent de nombreuses caractéristiques communes avec les structures pétrolifères et gazéifères du bassin du Michigan.

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[1] A contribution to the Frontier Geoscience Program
[2] Atlantic Geoscience Centre, Dartmouth
[3] Lithosphere and Canadian Shield Division, Ottawa

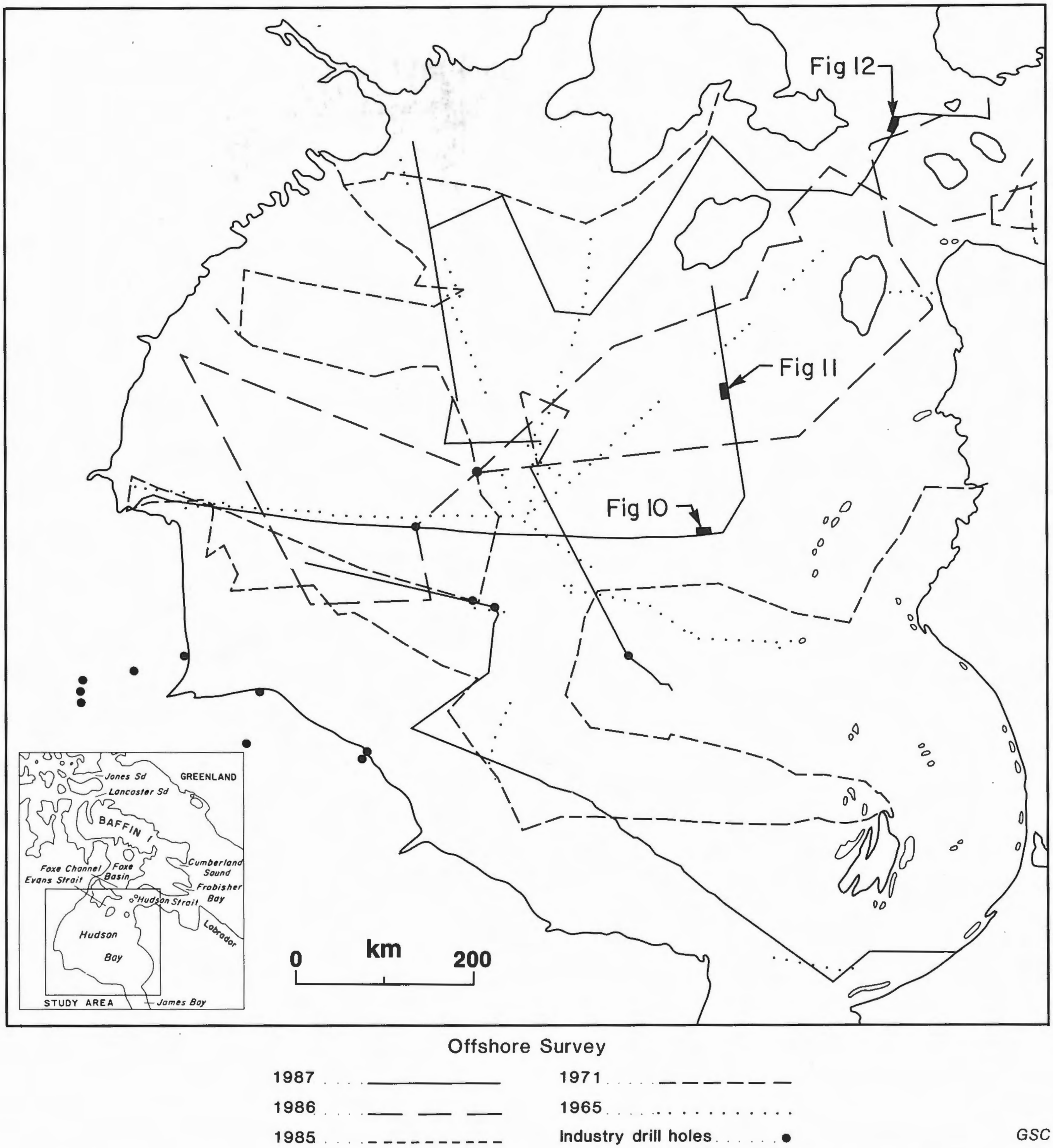


Figure 1. Location map showing lines of seismic coverage by the Geological survey of Canada and sites of exploratory wells by industry. Thick-line segments are locations of seismic profiles shown in Figures 10 to 12.

INTRODUCTION

Objectives of the study

In 1986 and 1987 the Canadian research vessel C.S.S. HUDSON carried out multidisciplinary seismic surveys in Hudson Bay on behalf of the Geological Survey of Canada to collect data on bedrock and surficial geology (Fig. 1). The purpose of renewed investigation of Hudson Bay is to upgrade the geological map, identify "exploration plays" that may have been overlooked by past exploration programs, and to estimate the potential hydrocarbon reserves of the region. The two cruises carried out to date were funded by the Frontier Geoscience Program of the Department of Energy, Mines and Resources. The information from these surveys, integrated with industrial seismic data from Hudson Bay, is being used in the construction of suites of maps encompassing bathymetry, bedrock geology, structure maps of basement and selected subsurface Phanerozoic horizons, relevant isopach and lithofacies maps, and structure sections. Structure and isopach maps will be converted from reflection time to depth (metres).

This paper reports on some of the highlights of cruise results related to bedrock geology; studies of Quaternary surficial deposits are reported elsewhere in this volume (Josenhans et al., 1988).

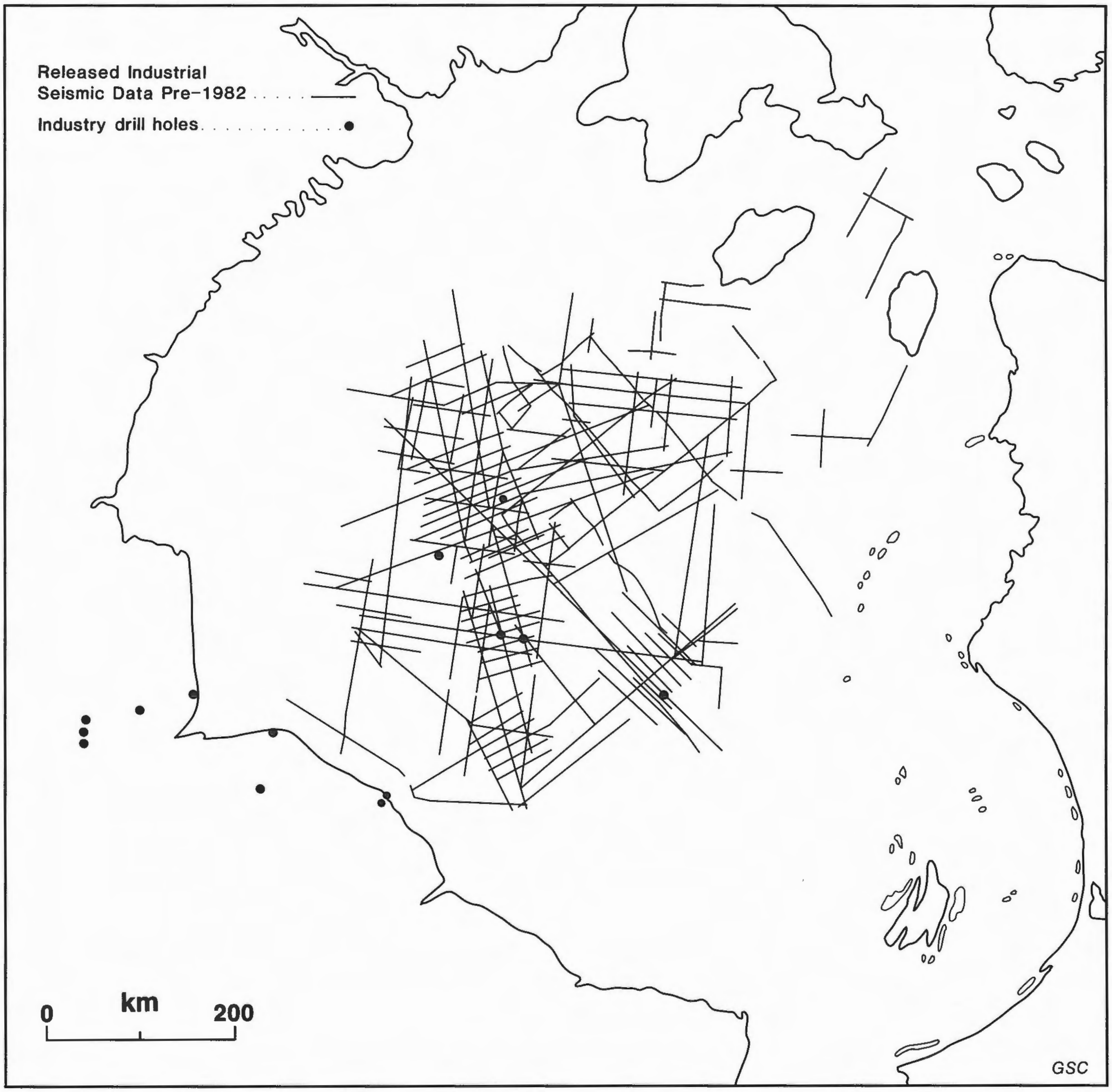


Figure 2. Location map showing lines of industrial multichannel seismic data released to open file by the Canada Oil and Gas Lands Administration, Ottawa, and sites of exploratory wells.

The data base

In 1987 (cruises Hu87-028 and -031, Aug. 3-28) approximately 3700 km of single channel seismic profiling was completed using 655 cm3 air gun energy sources, and a Huntec D.T.S. deep-tow ("boomer") profiler and B.I.O. 75 kHz sidescan sonar systems (Fig. 1). In addition, bedrock cores were recovered at a number of localities using an underwater electric drill contracted from Nordco Ltd. of Newfoundland.

In 1986 (Cruise Hu86-040, Oct. 4-21) a high resolution multichannel (24) reflection seismic system with a 16 000 joule sparker array energy source was contracted from McElhanney Services Ltd. for deep penetration bedrock studies. More than 2500 km of data were collected with this system (Fig. 1), which was operated concurrently with a single channel profiler (655 cm3 air gun energy source) and a Huntec D.T.S. deep-tow ("boomer") profiler equipped with a 50 kHz sidescan sonar system. Approximately 1500 km of the multichannel seismic data were processed by geophysical Service Incorporated, Calgary.

Data from previous programs in Hudson Bay by the Geological Survey of Canada (Fig. 1) include approximately 4400 km of sidescan sonar and single channel seismic (164 cm3 air gun) profiling by the M.V. HUDSON HANDLER in 1971 (Sanford, 1974). Bedrock cores were collected on

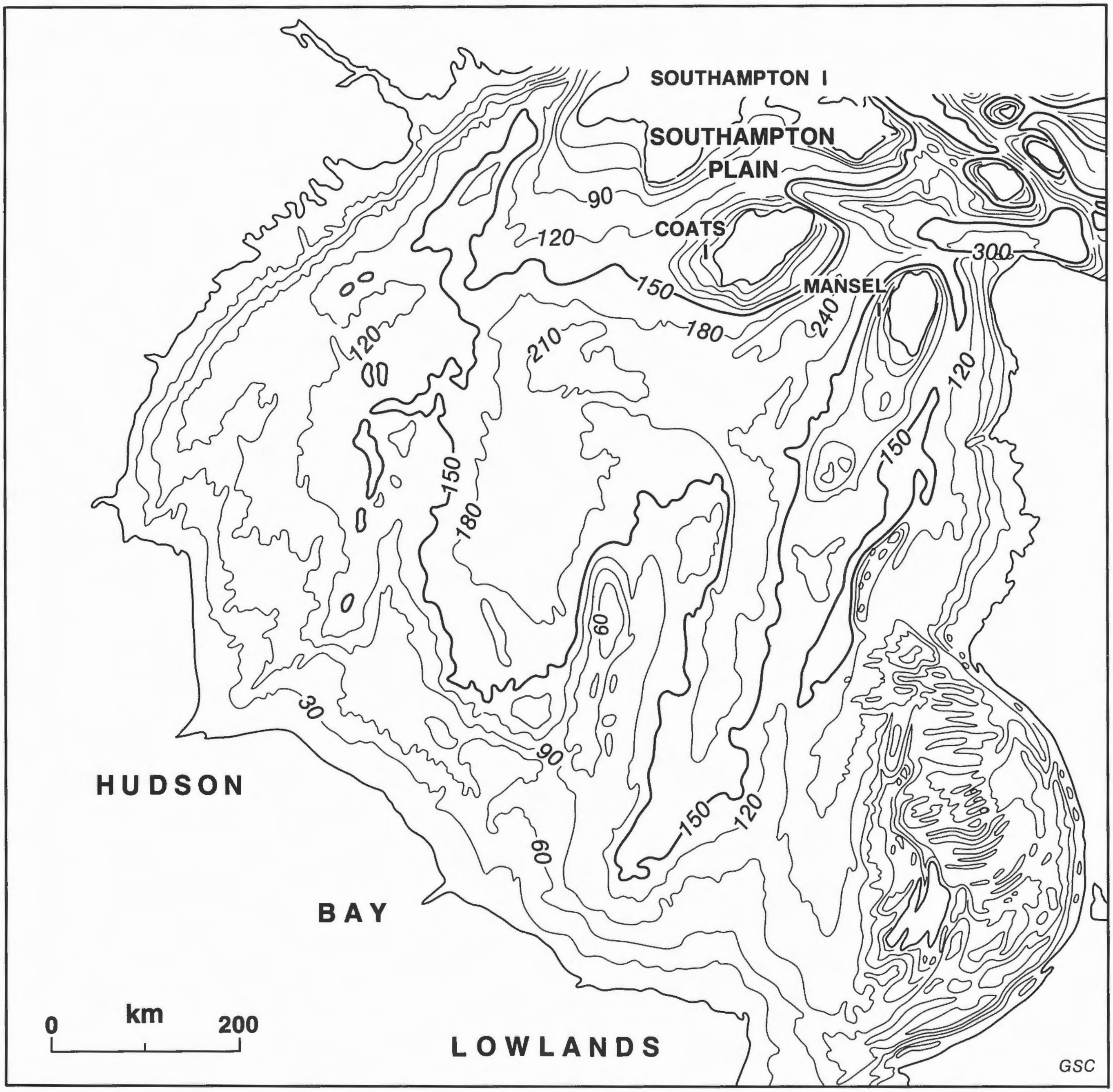


Figure 3. Generalized bathymetric map of Hudson Bay compiled from all available sources (contour interval 30 m).

this survey using an underwater diamond drill, and seafloor observations and sampling were conducted with the PISCES III submersible. The seismic data from this cruise are of limited resolution in many areas. In 1965 the C.S.S. HUDSON carried out a multidisclipinary survey in Hudson Bay that recovered approximately 1800 km of single channel (sparker) profiler coverage (Grant, 1969).

Data from the above field programs are being interpreted in conjunction with multichannel seismic and exploratory drilling data collected in Hudson Bay by oil companies since the mid-1960s. Industrial seismic coverage exceeds 35 000 line km (Fig. 2); five wells have been drilled offshore, five along or near the southwest soast of Hudson Bay in Ontario and Manitoba, and four boreholes have been drilled farther onshore in Manitoba for stratigraphic purposes and mineral exploration. The single channel profiles are an important complement to the multichannel coverage, because they resolve shallow details not detected by the lower frequency, deeper penetration industry data. Compilation of bathymetric data from all sources is a key element of this interpretation (Fig. 3), because seafloor physiography accurately reflects bedrock lithology and structure. Concurrent acquisition of bedrock and surficial geological data has been a standard procedure of all programs in the region carried out by the Geological Survey of Canada.

REGIONAL GEOLOGY

The Paleozoic rocks of the Hudson Platform that blanket Archean and Proterozoic terranes are erosional remnants of a much broader cratonic cover that once connected with Lower Paleozoic platformal areas to the north and south (Fig. 4). The Hudson Platform is now separated from the latter by Precambrian basement arches that were rejuvenated at various times in the Early Paleozoic and later (Sanford, in press).

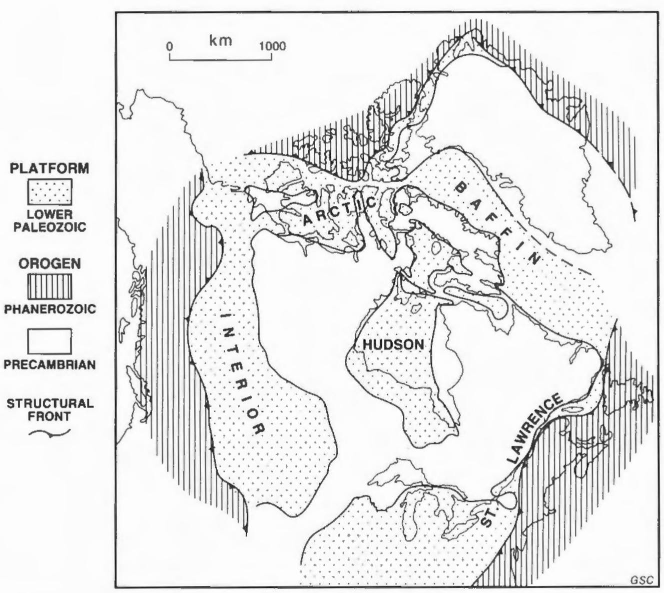


Figure 4. Lower Paleozoic platforms of northern North America (after Sanford, in press).

The Paleozoic rocks of the Hudson Platform range in age from Cambrian to Devonian (Fig. 5), and are analogous in many aspects of age, lithology and structure to the rocks of the Michigan Basin to the south (Fig. 4).

The inception and evolution of the Hudson Platform undoubtedly was linked to Proterozoic geosynclines that formed along the margins of the stable Archean platforms in Aphebian time. An attempt is being made in this study to piece together the surface and subsurface distribution and structural framework of the Proterozoic supracrustal rocks along the south and east sides of the Hudson Bay Basin from seismic data, bathymetry and potential field data. Some tentative interpretations of the distribution of the various fold belts that border the Paleozoic along the east side of Hudson Bay are shown in Figure 6.

The principal tectonic element of the Hudson Platform (Fig. 4) is the Hudson Bay Basin that lies mainly offshore, extending into northern Ontario and Manitoba to the south and including major parts of Southampton, Coats and Mansel islands in the north. An important segment of the platform to the south is the Moose River Basin underlying the western part of James Bay and adjacent onshore areas of Ontario and Quebec. To the north and northwest Paleozoic rocks of the Hudson Platform are confined to Foxe Basin and to the complex graben structures beneath Foxe Channel, Evans Strait and Hudson Strait.

Hudson Bay Basin, the main target of this study, contains a variable succession of Paleozoic rocks that are more than 2 km thick beneath the central part of the bay (Fig. 7). Although the strata express a relatively symmetrical saucer-shaped basin at surface (Fig. 6), the rocks in the subsurface

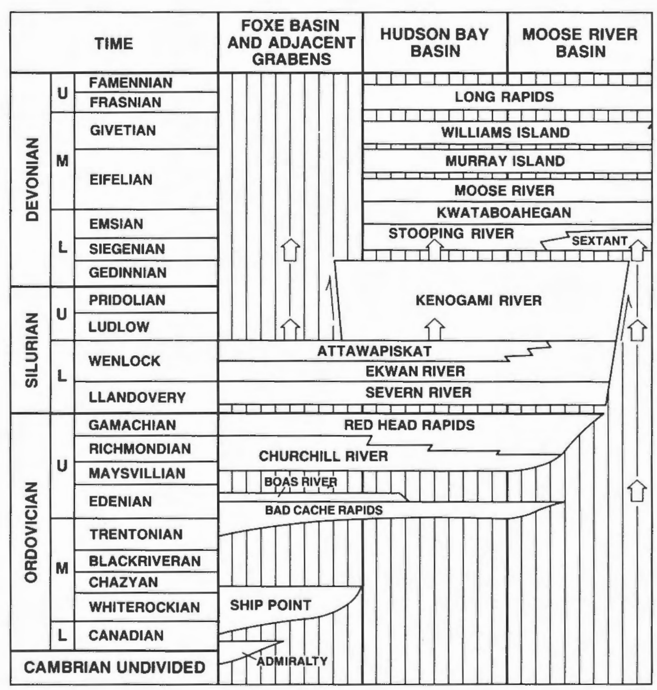
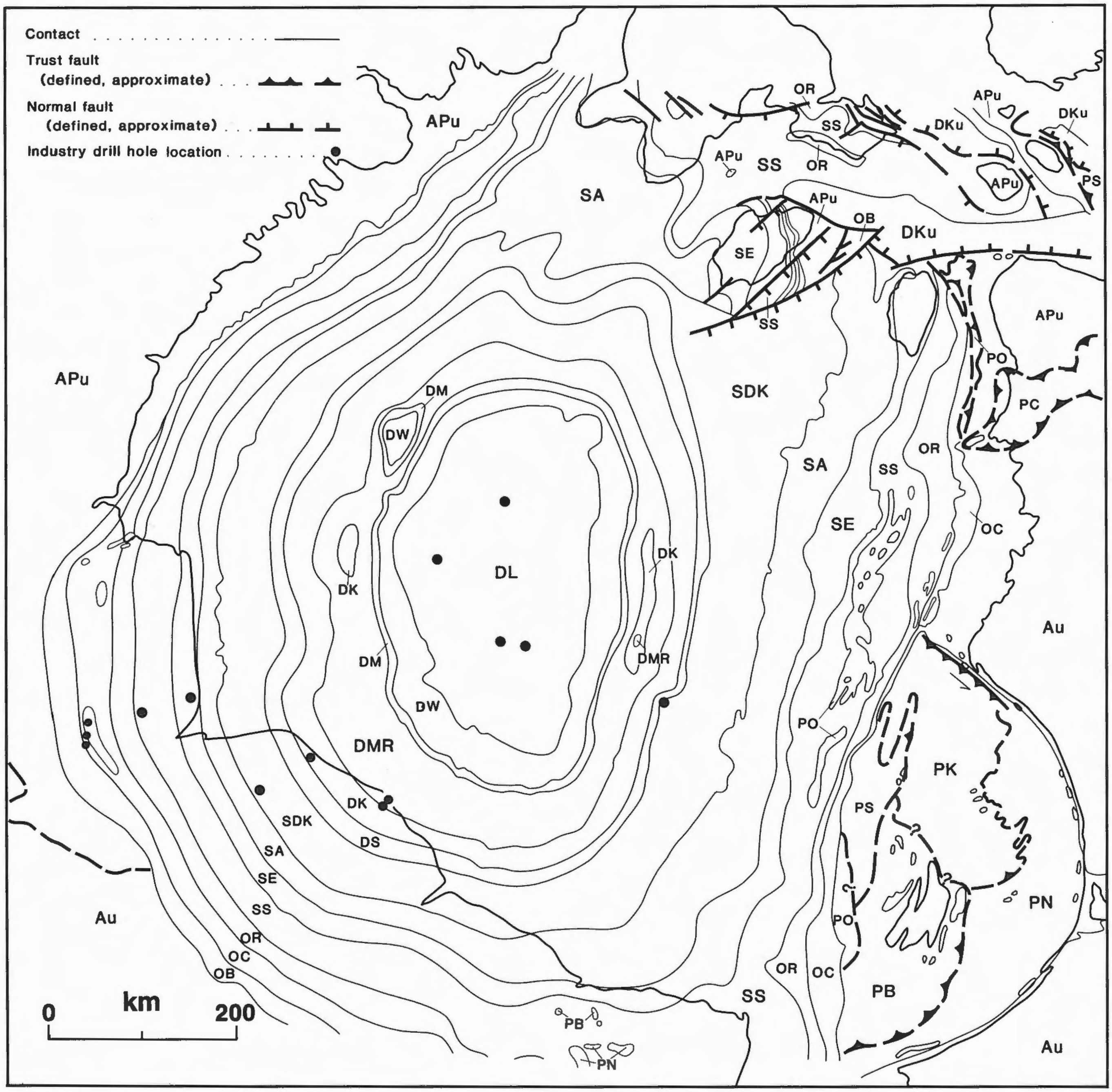


Figure 5. Stratigraphic framework of the Hudson Bay region (after Sanford, in press).



DEVONIAN TO CRETACEOUS ?

DKu – Sand/Shale/evaporite sequence undivided

DEVONIAN

DL – Long Rapids

DW – Williams Island

DM – Murray Island

DMR – Moose River

DK – Kwataboahegan

DS – Stooping River

SILURIAN/DEVONIAN

SDK – Kenogami River

SILURIAN

SA – Attawapiskat

SE – Ekwan River

SS – Severn River

ORDOVICIAN

OR – Red Head Rapids

OC – Churchill River

OB – Bad Cache Rapids

ARCHEAN/PROTEROZOIC

APU – ARCHEAN AND PROTEROZOIC UNDIVIDED

Proterozoic folded belts

PB – Belcher

PS – Sleeper

PO – Ottawa

PC – Cape Smith

PK – King George

PS – Salisbury

Proterozoic platform

PN – Nastapoka

ARCHEAN

AU – ARCHEAN UNDIVIDED

GSC

Figure 6. Geological map of the Hudson Bay region, and sites of exploratory wells by industry.

in the centre of the basin are interrupted by a north- and northwest-trending basement high (Fig. 8), and by northwest-oriented faults along the Boothia-Bell Arch trend (Sanford, in press) in the vicinity of Southampton, Coats and Mansel islands.

The oldest Paleozoic rocks contained in the Hudson Bay Basin are Late Ordovician (Fig. 9), and these are succeeded by strata of Early Silurian age. The rocks consist mainly of carbonates and minor evaporites that have broadly similar lithological and biostratigraphical character to coeval strata in the Arctic, Interior and St. Lawrence platforms (Fig. 4). Following their deposition, major faulting possibly triggered by Caledonian Orogeny activated the basement arch systems bordering and underlying Hudson Bay Basin. Uplift of the fault-bounded blocks in central Hudson Bay occurred in the Early Silurian late Llandovery to Wenlock time (Fig. 8). Displacement across the faults bounding the blocks amounts to more than one kilometre. Three of the exploratory wells drilled in Hudson Bay are located on prominent structures associated with this early Silurian uplift. The succeeding Upper Silurian and lower Devonian shales consist of redbed siltstones, shales and mudstones with thick interbeds of

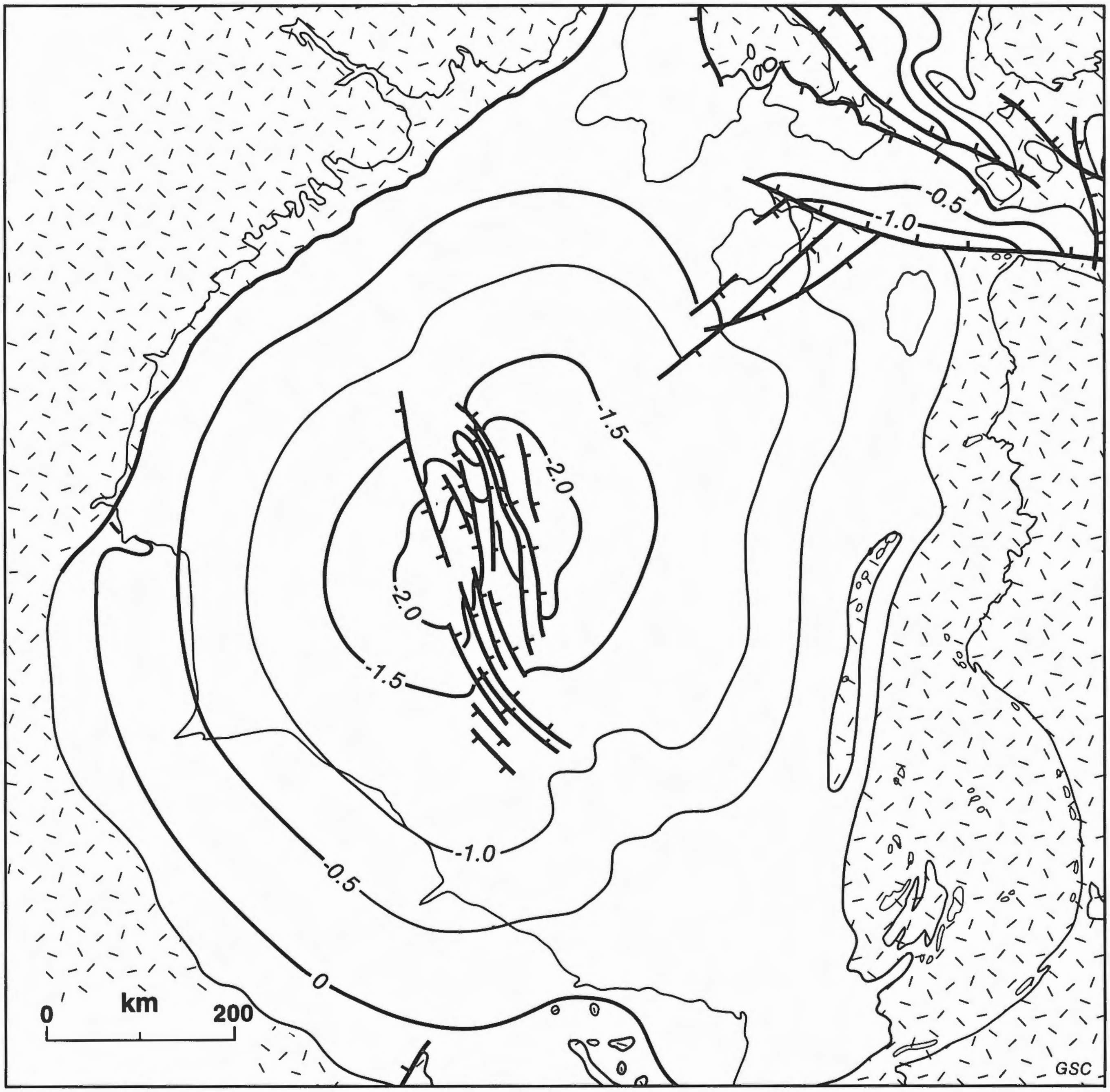


Figure 7. Structure contours on Precambrian basement (after Sanford, in press). Note that this map does not incorporate data from 1986 and 1987 cruises.

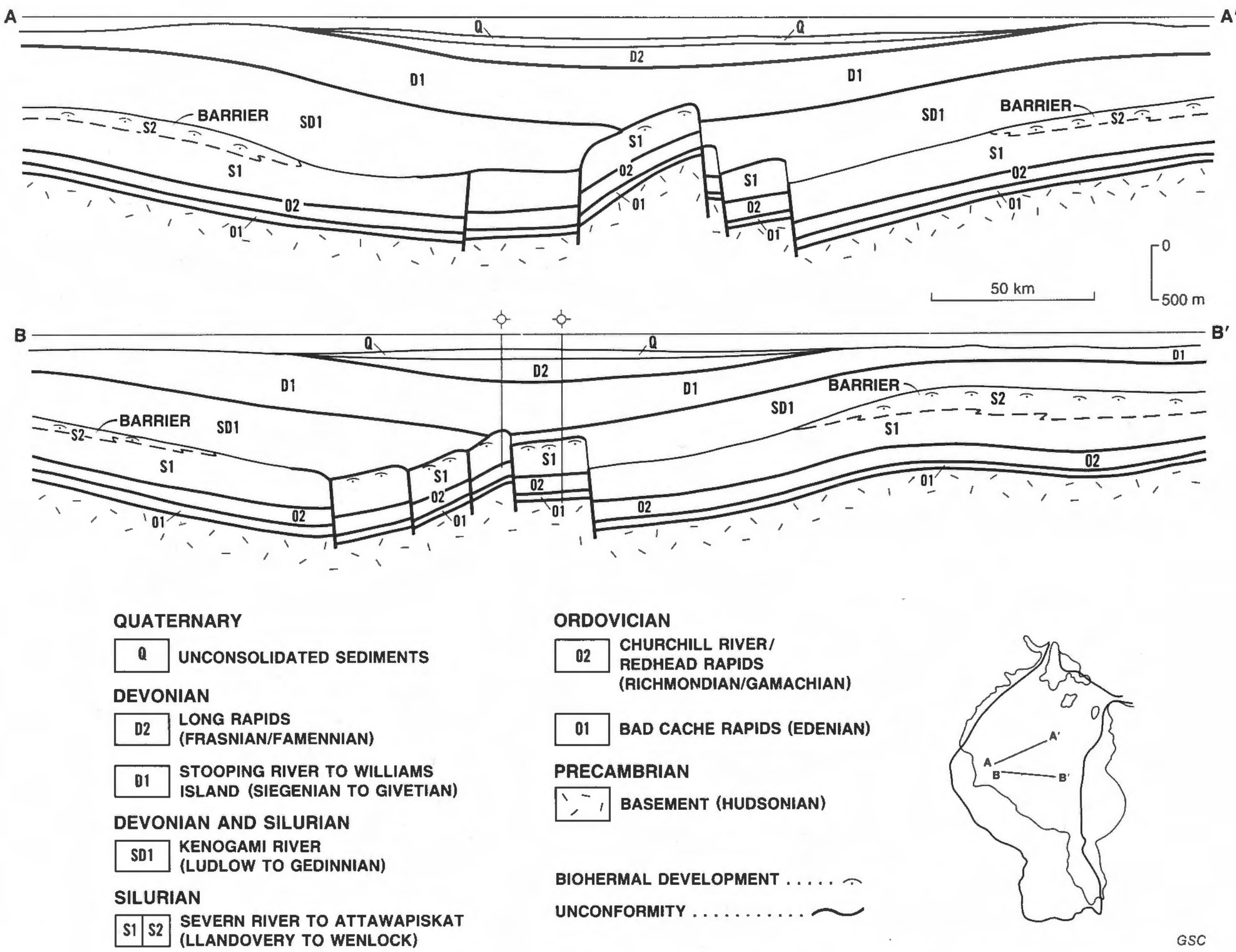


Figure 8. Structural cross sections of the central Hudson Bay Basin (after Sanford, in press).

evaporites and carbonates of the Kenogami River Formation (Fig. 9). Halite beds, an important component of the Kenogami River, are more than 700 m thick in a borehole in the subbasin west of the Central Hudson Bay Arch (Fig. 7). Siliclastics interbedded with carbonates and evaporites are an important component also of the Middle Devonian, and are dominant in the Upper Devonian. Reef-bearing zones occur at several intervals of the Paleozoic succession — in the Upper Ordovician Red Head Rapids Formation, Lower Silurian Attawapiskat Formation and in the Middle Devonian Kwataboahegan and Williams Island formations (Fig. 9).

Regional papers on the structure and stratigraphy of the Hudson Bay region have been written by Nelson and Johnson (1966), Sanford and Norris (1973) and Sanford (in press).

RESULTS

The bedrock geology of Hudson Bay has been substantially revised to incorporate preliminary interpretation of the seismic data collected in the past two field seasons (Fig. 1, 6). Further revision of the geological map (Fig. 6) will likely be necessary when these data have been thoroughly analysed and integrated with industrial multichannel seismic data. The rock unit boundaries are based on analysis of seafloor physiography, deep and shallow seismic data, and stratigraphic control from exploratory wells and bedrock sampling.

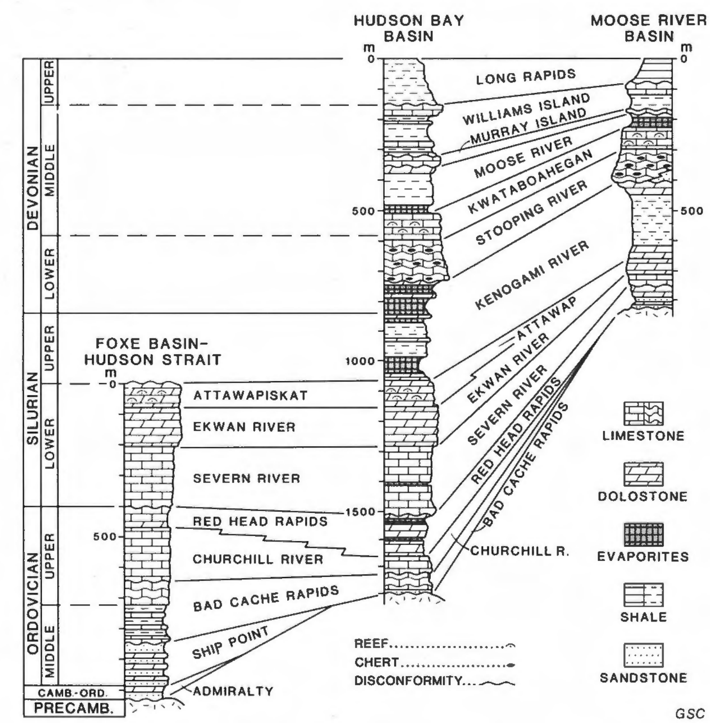


Figure 9. Composite stratigraphic column for the Hudson Bay region (after Sanford, in press).

Examples of seismic records illustrating structural aspects of bedrock units are presented in Figures 10 to 12.

An example of inferred salt collapse in the Kenogami River Formation is illustrated in the seismic record of Figure 10 where it crosses one of the deeper segments of the Winisk Trough (Sanford and Grant, 1976). The zone of greatest collapse is enclosed by a 375-m isobath that is approximately 200 m deeper than the surrounding seafloor. Glacial deposits up to 30 m thick occur in adjacent segments of the Winisk trough both to the north and south (H. Josenhans, personal communication, 1987), and this anamolous depression would appear to be a natural trap for the accumulation of similarly thick Quaternary deposits. The relative absence of Quaternary sediments in the depression illustrated in Figure 10 may suggest that the salt collapse is of quite recent origin.

The zone of reefal structure illustrated by Figure 11, occurring in the Atawapiskat Formation beneath the draped reflectors of the Kenogami River Formation, can be traced around the outer margins of the Hudson Bay Basin (Fig. 9). Figures 10 and 11 illustrate the type of salt collapse and reefal structures that may form traps for hydrocarbons.

The seismic profile (Fig. 12) from Foxe Channel (Fig. 1) shows sedimentary strata dipping into a fault against presumed Precambrian basement. Projecting observed reflectors and applying conservative velocity estimates indicates the section attains a thickness of approximately 2 km. Similar structural relationships are observed in Evans Strait. The section in these basins appears to reach thicknesses similar to the maximum in the Hudson Bay Basin to the south, and in Hudson Strait to the east (Grant and Manchester, 1970; MacLean et al., 1986a). The age of the strata beneath Foxe Channel and Evans Strait is uncertain because no bedrock samples have been recovered in these fault-bounded subbasins. These basins are part of the Hudson Strait structural trend; the Phanerozoic basins in Hudson Strait appear to be predominantly half graben (MacLean et al., 1986a). It has been speculated that rocks as young as Cretaceous may occur locally in these basins (Grant and Manchester, 1970; MacLean et al., 1986a). Regionally, the structural style of Hudson Strait may be compared with that of the fault-bounded basins underlying Frobisher Bay, Cumberland Sound, and Lancaster and Jones sounds to the North (Fig. 1). Cretaceous rocks have been sampled in Cumberland Sound (MacLean et al, 1986b), and thick sections of Cretaceous sediments are suspected in Lancaster Sound and Jones Sound (Hea et al., 1980).

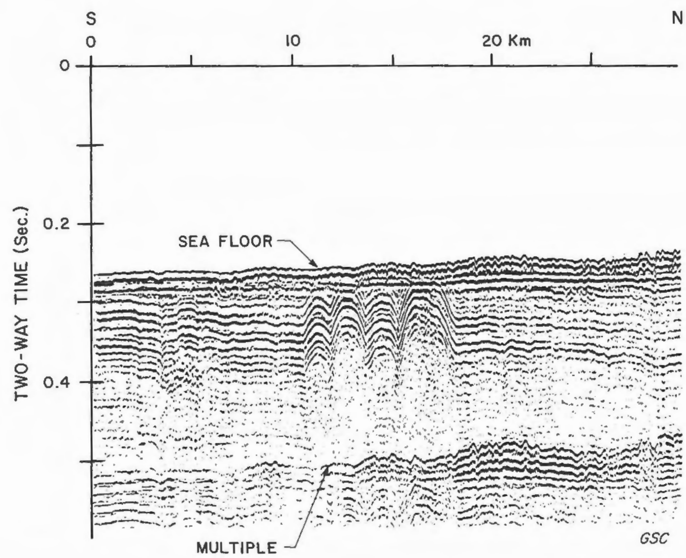


Figure 11. Seismic reflection profile showing reefal structure (see Fig. 1 for section location).

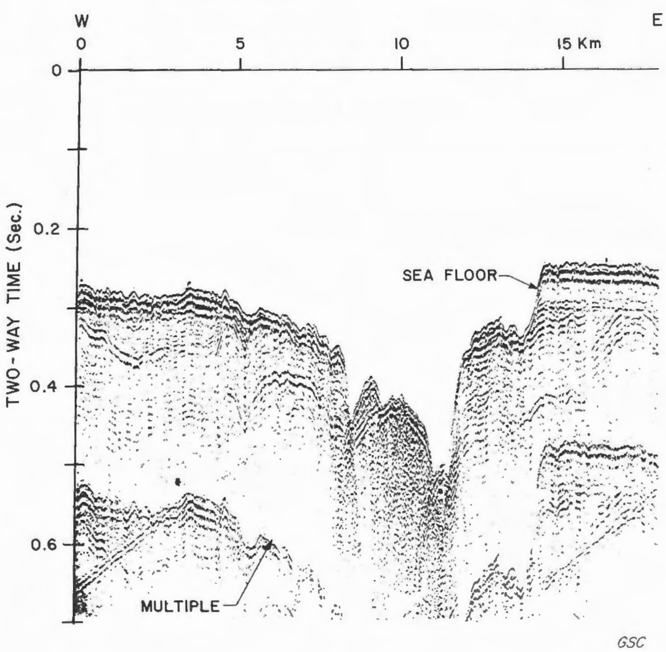


Figure 10 Seismic reflection profile showing salt collapse structure (see Fig. 1 for section location).

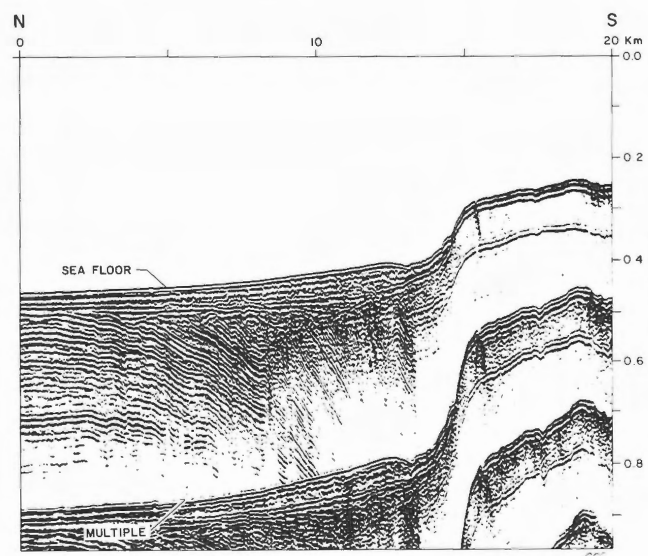


Figure 12. Seismic reflection profile showing thick section in Foxe Channel (see Fig. 1 for section location).

OIL AND GAS POSSIBILITIES

The Phanerozoic rocks in the Hudson Bay region have many characteristics in common with rocks of comparable age that yield oil and gas in the Michigan and Williston basins. It would thus be surprising if the rocks were found to be completely devoid of these commodities in similar settings in the Hudson Bay Basin. The Geological Survey of Canada (Proctor et al., 1984) has estimated hydrocarbon resources of the Hudson Platform in the approximate order of 3 billion barrels of oil and 14 tcf of gas. These estimates are based on a minimum of geological information and are subject to revision on the basis of the present study.

Extensive fault-bounded structures along with biohermal facies and salt collapse structures have now been mapped in the Hudson Bay region, proving widespread occurrence of potential trapping mechanisms for hydrocarbons. Reservoir rocks may occur in zones of clastic, fracture and solution porosity associated with both reefal and salt dissolution structures. The Boas River oil shale of Ordovician age that was long thought to be restricted to Southampton Island has now been identified by the writers in many of the boreholes completed in the Hudson Bay Basin. The Boas River shale unit may be best developed in the regionally and locally deeper parts of the basin. The deeper basinal areas are generally not the settings that have been tested by exploratory drilling in Hudson Bay to date. Most of the drilling targets in fact were located on the largest and highest structural anomalies in the basin.

The fault-bounded basins beneath Foxe Channel and Evans Strait, in the northeast part of the platform (Fig. 1), locally contain thick sedimetary sections. Source rocks (Boas River and Sixteen Mile Brook oil shales) occur nearby on Southampton Island (Dwing et al., 1987). If these basins occupy zones with long histories of depositional restriction, then the possibility for occurrence of source rocks may be enhanced. They are not large basins, but their restricted settings may contain favourable source, reservoir and trapping conditions.

In summary, hydrocarbon accumulations in the Hudson Bay region could conceivably favour some of the smaller basins and structures. There is no reason why these regions could not be productive, and, because the potential traps occur at shallow depth, they may be economically exploitable. Relative to the Arctic and East Coast offshore regions of Canada where thick pack ice and icebergs hamper exploration and development, Hudson Bay region is subject only to relatively thin seasonal ice.

ACKNOWLEDGMENT

The writers wish to thank Captains F.W. Mauger and L.A. Strum and the officers, crew, and scientific staff of the C.S.S. HUDSON for dedicated and cheerful support during the 1986 and 1987 cruises in Hudson Bay. Grateful acknowledgment is extended to H.W. Josenhans, the chief scientist on both cruises, who was responsible for the logistical planning and day to day direction of the work. Thanks are also extended to H.W. Josenhans and B. MacLean for reviewing the manuscript.

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Acoustic tests of seabottom core in Hudson Bay

P.J. Kurfurst[1] and B.V. Sanford[2]

Kurfurst, P.J. and Sanford, B.V., Acoustic tests of seabottom core in Hudson Bay; in Current Research, Part B, Geological Survey of Canada, Paper 88-1B, p. 297-299, 1988.

Abstract

Acoustic wave velocities were measured on core samples drilled from the Paleozoic bedrock at various locations of the Hudson Bay during the scientific cruise 87-028. Brief geology of the bedrock encountered and range of dynamic elastic constants derived from the field measurements are summarized for all drill stations.

Résumé

On a mesuré la vitesse des ondes acoustiques sur des échantillons prélevés par carottage dans le socle paléozoïque en divers endroits de la baie d'Hudson durant la croisière scientifique 87-028. On donne un bref résumé géologique sur la nature de la roche de fond rencontrée, et l'on indique la gamme de constantes élastiques dynamiques dérivées des mesures réalisées in situ, pour toutes les stations de forage.

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Cruise 87-028 was a geological survey of Hudson Bay using seismic techniques to map the subsea distribution of Paleozoic formations, along with the overlying unconsolidated deposits of Quaternary and recent age. To validate the geophysical data, piston cores of Quaternary deposits and diamond drill cores of the Paleozoic bedrock were taken at various localities within the bay.

In this region, rocks of Ordovician, Silurian and Devonian ages, that are more than 2 km thick beneath the central part of the bay, are contained within a major cratonic sedimentary basin (Hudson Bay Basin). The oldest rocks of the Ordovician System lie beneath the seafloor and adjacent onshore areas along the outer margins of the basin, and these are succeeded by younger rocks of the Silurian and Devonian systems towards the central part of Hudson Bay (Grant and Sanford, 1988).

Attempts to core the Paleozoic formations were made at fourteen locations during Cruise 87-028. Seven were successful in recovering material thought to be representative of bedrock formations beneath the seafloor. Of the seven, only five cores (stations 9, 10, 18, 23 and 37 — see Fig. 1), were of sufficient length for measurements of the seismic velocity of the rock types to be made. The location and highly generalized description of the cores encountered at the above stations are as follows:

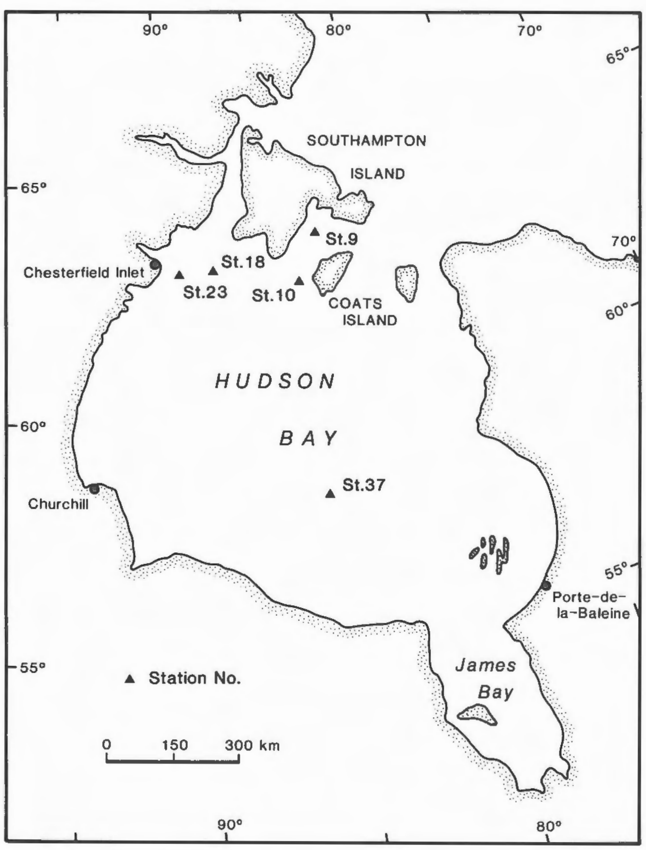

Figure 1. Location map of drill stations.

Station 9, located at 63°30.21'N; 83°16.43'W, penetrated 15.24 cm of strata classified as Severn River Formation of early Silurian age. The rocks consist of uniformly bedded, light brown coloured, relatively nonporous micritic to finely crystalline limestone.

Station 10, located at 62°22.47'N; 84°30.27'W, penetrated 116.64 cm of carbonate of the Attawapiskat Formation of early Silurian age. The rock consists of massive reefal dolostones with medium crystalline texture that are light tan in colour. Intercrystalline and vuggy porosity are common throughout.

Station 18, located at 62°41.42'N; 87°44.68'W, penetrated 5.08 cm of the Attawapiskat Formation of early Silurian age. The rock consists of cream coloured, highly porous medium crystalline dolostone.

Station 23, located at 62°58.51'N; 89°48.20'W, penetrated 5.08 cm of carbonate of either the Bad Cache Rapids or Churchill River group of late Ordovician age. It consists of medium greyish-brown, argillaceous limestone that contains thin shale partings and crinoid ossicles throughout.

Station 37, located at 58°42.20'N; 84°50.40'W, penetrated 17.78 cm of carbonate that resembles the Stooping River and Kwataboahegan Formations of early and middle Devonian age respectively. The upper 12.70 cm consist of medium to dark brown bituminous limestone containing many fossils including brachiopods and corals. This lithology closely resembles the Kwataboahegan Formation as identified in industry boreholes drilled elsewhere in the Bay. The lower 5.28 cm, consisting of light tan, fine to medium grained dolomitic limestone, is more closely related to the older Stooping River Formation. All of the material recovered at this locality is assumed to have been from boulders detached from the bedrock (Kwataboahegan?) surface.

Acoustic wave velocity measurements were carried out on all core samples obtained during the cruise aboard the ship, immediately after recovery, and then were repeated later in the Ottawa laboratory. An OYO Sonic viewer 5217A was used to measure compressional and shear wave velocities independently. 100 kHz and 200 kHz frequency transducers respectively were used for measurements of the first arrival of the shear and compressional waves. Pulse travel times were read directly from the display screen and were simultaneously recorded and printed. The details of the technique used are described by Kurfurst (1977).

The sample lengths(L) were measured and their bulk density (ρ) was determined in the laboratory. Using these parameters and the calculated compressional (V_p) and shear (V_s) wave velocities, acoustic elastic constants such as Young's modulus (E), Poisson's ratio (υ), shear modulus/modulus of rigidity (G) and bulk modulus (K) were calculated using formulas recommended by the American Society for Testing and Materials (ASTM, 1984). The results are reported in Table 1.

Table 1 — Acoustic properties

Sample	L (cm)	ρ (g/cm^3)	Vp (m/s)	Vs (m/s)	E (GPa)	υ	G (GPa)	K (GPa)
9-1	2.94	2.68	5888	3062	66.1	.31	25.1	59.4
9-2	5.42	2.82	5420	3011	65.3	.28	25.6	48.8
10-1	3.53	2.59	4645	2942	52.2	.17	22.4	26.0
10-2	3.53	2.62	4770	2942	54.1	.19	22.7	29.4
18-1	4.79	2.47	4989	2721	47.1	.29	18.3	37.1
23-1	6.60	2.65	4459	2853	49.8	.15	21.6	23.9
23-2	10.45	2.68	4283	2724	46.2	.16	19.9	22.6
37-1	6.43	2.70	4296	2724	46.6	.16	20.0	23.1

In general, all results of acoustic wave velocity measurements are very consistent with variations of q10 % from the mean with the exception of samples from Station No. 9. Detailed study of the petrography of the samples and their origin will be required for interpretation of their acoustic properties.

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