GEOLOGICAL SURVEY OF CANADA
BULLETIN 427

GEOLOGY OF THE CORNER BROOK-GLOVER ISLAND REGION, NEWFOUNDLAND

P.A. Cawood and J.A.M. van Gool

1998

Natural Resources Canada   Ressources naturelles Canada

Canada

COOPERATION

COOPERATION AGREEMENT ON MINERAL DEVELOPMENT

ENTENTE DE COOPÉRATION SUR L'EXPLOITATION MINÉRALE

Contribution to Canada-Newfoundland Cooperation Agreement on Mineral Development (1990-1994), a subsidiary agreement under the Economic and Regional Development Agreement.

Contribution à l'Entente de coopération Canada-Terre-Neuve sur l'exploitation minérale (1990-1994), entente auxiliaire négociée en vertu de l'Entente Canada-Nouveau-Brunswick de développement économique et régional.

Canada

Newfoundland
Terre-Neuve

GEOLOGICAL SURVEY OF CANADA
BULLETIN 427

GEOLOGY OF THE CORNER BROOK-GLOVER ISLAND REGION, NEWFOUNDLAND

P.A. Cawood and J.A.M. van Gool

1998

Catalogue No. M42-427E
ISBN 0-660-17109-0

Available in Canada from
Geological Survey of Canada offices:

601 Booth Street
Ottawa, Ontario K1A 0E8

3303-33rd Street N.W.
Calgary, Alberta T2L 2A7

101-605 Robson Street
Vancouver, B.C. V6B 5J3

or from

Canadian Government Publishing
Public Works and Government Services Canada
Ottawa, Ontario K1A 0S9

A deposit copy of this publication is also available for reference in selected public libraries across Canada

Price subject to change without notice

Cover illustration

View of Humber Arm and the city of Corner Brook from Eastern Lake. The outcrops in the foreground are psammitic schist of the Upper Precambrian South Brook Formation. GSC 1995-149DDDD

Critical reviewer

K.L. Currie

Authors' addresses

P.A. Cawood
Tectonics Special Research Centre
School of Applied Geology,
Curtin University of Technology,
G.P.O. Box U1987, Perth 6001,
Western Australia, Australia

J.A.M. van Gool
Danish Lithosphere Centre
Øster Voldgade 10,
1350 Copenhagen K,
Denmark

Original manuscript submitted: 1994-01
Final version approved for publication: 1995-09

Preface

The Corner Brook Lake-Glover Island region has been recognized for some years as a key to understanding both the tectonic development of the eastern margin of North America in early Paleozoic time, and the relation of economic mineralization to this development. This study shows conclusively that the bulk of deformation, metamorphism, intrusion, and mineralization is of Silurian age, not middle Ordovician as previously supposed. The careful deciphering of structure and metamorphism by Cawood and van Gool presents for the first time a clear description of how these events unfolded.

This important study was conducted by contract under the Canada-Newfoundland Mineral Development Agreement, insuring both that the most knowledgeable local experts conducted the study, and that the study was made with the full co-operation of the Geological Survey of Canada, the Geological Surveys Branch of the province of Newfoundland, and the mineral industry.

M.D. Everell
Assistant Deputy Minister
Earth Sciences Sector

Préface

La région du lac Corner Brook et de l'île Glover est reconnue depuis quelques années comme étant d'une importance capitale pour comprendre, d'une part, l'évolution tectonique de la marge orientale de l'Amérique du Nord au début du Paléozoïque et, d'autre part, la relation entre cette évolution et une éventuelle minéralisation exploitable. La présente étude démontre que la déformation, le métamorphisme, les épisodes intrusifs et la minéralisation datent du Silurien, et non de l'Ordovicien moyen comme on le croyait auparavant. L'analyse détaillée du style structural et du métamorphisme de Cawood et van Gool indique clairement et pour la première fois comment ces événements se sont produits.

Cette importante étude a été menée à contrat dans le cadre de l'Entente Canada – Terre-Neuve sur l'exploitation minérale; elle a donc été confiée aux plus grands spécialistes de la région et a été réalisée en collaboration avec la Commission géologique du Canada, la *Geological Surveys Branch* de la province de Terre-Neuve et l'industrie minière.

M.D. Everell
Sous-ministre adjoint
Secteur des sciences de la Terre

CONTENTS

Tables

Illustrations

GEOLOGY OF THE CORNER BROOK-GLOVER ISLAND REGION, NEWFOUNDLAND

Abstract

The Corner Brook Lake-Glover Island region includes carbonate shelf rocks and parts of the Humber Arm Allochthon (externides of the Humber Zone) as well as slices of variously metamorphosed rift-facies siliciclastic rocks resting on Precambrian basement and intruded by latest Precambrian igneous rocks (internides of the Humber Zone). U-Pb zircon dating suggests that the rift-drift transition, marking opening of the Iapetus Ocean, occurred about 555 Ma. Ophiolitic rocks of the Dunnage Zone, thrust westward over the Humber Zone, are locally intercalated with the siliciclastic rocks. Original emplacement of the Humber Arm Allochthon and ophiolites on Glover Island may have been a mid-Ordovician (Taconic) event. However, most of the deformation is Silurian (as dated by U-Pb zircon, rutile, and monazite, and Ar-Ar ages). Deformation was accompanied by thermal activity which produced P-T conditions as high as 7-9 kilobars at about 650°C in a metamorphic culmination just west of the Humber-Dunnage boundary, and extensive granitoid plutonism in the Dunnage Zone. Following this climactic event, relaxation produced extensional and transtensional faults culminating in development of the Carboniferous Deer Lake Basin, a pull-apart basin along the Cabot Fault system which is filled with little-metamorphosed clastic rocks. Important gold prospects occur along the Humber-Dunnage boundary on Glover Island, zinc prospects occur in carbonates of the Humber Zone, and garnet and staurolite have been examined as industrial minerals in the metamorphic rocks. All the mineralization is Silurian or younger.

Résumé

La région du lac Corner Brook et de l'île Glover renferme des roches de plate-forme carbonatée et englobe des parties de l'Allochtone de Humber Arm (division externe de la Zone de Humber); on y observe aussi des lambeaux de roches silicoclastiques (faciès de rift) diversement métamorphisées, reposant sur un socle précambrien et entrecoupées des roches ignées précambriennes les plus tardives (division interne de la Zone de Humber). Les données de datation U/Pb sur zircon indiquent que la transition rift-dérive, marquant l'ouverture de l'océan Iapetus, s'est produite il y a environ 555 Ma. Les roches ophiolitiques de la Zone de Dunnage, charriées vers l'ouest par-dessus la Zone de Humber, sont intercalées par endroits avec les roches silicoclastiques. Sur l'île Glover, il se peut que la mise en place initiale de l'Allochtone de Humber Arm et des ophiolites soit un événement de l'Ordovicien moyen (orogenése taconique). Toutefois, la déformation remonte essentiellement au Silurien (datation U/Pb sur zircon, rutile et monazite, ainsi que datation Ar/Ar). La déformation a été accompagnée d'une activité thermique qui est à l'origine de pressions atteignant 7 à 9 kilobars et de températures d'environ 650 °C (correspondant à une apothéose métamorphique juste à l'ouest de la limite entre les zones de Humber et de Dunnage), mais aussi d'un plutonisme granitoïde généralisé dans la Zone de Dunnage. Après cet événement, le relâchement des contraintes a produit des failles d'extension et de transtension, culminant avec la naissance du bassin de Deer Lake du Carbonifère; cette dépression est un bassin d'extension qui s'observe le long du système de failles de Cabot et qui est rempli de roches clastiques peu métamorphisées. D'importantes zones d'intérêt à minéralisation en or bordent la limite entre les zones de Humber et de Dunnage sur l'île Glover; d'autres à minéralisation en zinc se trouvent dans les roches carbonatées de la Zone de Humber; il y aurait aussi des minéraux industriels (du grenat et de la staurotide) dans les roches métamorphiques. L'ensemble de la minéralisation remonte au Silurien ou à une époque plus récente.

SUMMARY

The Corner Brook-Glover Island region straddles the eastern Humber and western Dunnage zones of the Newfoundland Appalachians. The region is divided into five lithotectonic belts: Humber Arm Allochthon and Carbonate Belt within the external domain of the Humber Zone, the Corner Brook Lake Belt within the internal domain of the Humber Zone, the Glover Island Belt of the Dunnage Zone, and the Deer Lake Basin.

SOMMAIRE

La région du lac Corner Brook et de l'île Glover chevauche la partie orientale de la Zone de Humber et la partie occidentale de la Zone de Dunnage, dans les Appalaches terre-neuviennes. La région est divisée en cinq secteurs lithotectoniques; ce sont les suivants : l'Allochtone de Humber Arm et la ceinture de roches carbonatées (division externe de la Zone de Humber ou Zone de Humber externe), la ceinture de Corner Brook Lake (division interne de la Zone de Humber ou Zone de Humber interne), la ceinture de Glover

The Carboniferous Deer Lake Basin lies unconformably on the older lithotectonic units; all boundaries between these other units are structural.

Precambrian banded and massive granitoid gneiss and amphibolite of the Corner Brook Lake Complex within the Corner Brook Lake Belt form the oldest lithostratigraphic unit within the Humber Zone and are the crystalline basement to a late Precambrian to middle Ordovician stratigraphic sequence. Zircon from the banded gneiss yielded a U-Pb age of 1510 ± 6 Ma. Silicic and mafic igneous rocks of the Lady Slipper Pluton, dated by U-Pb zircon at 555 +3/-5 Ma lie at the base of the stratigraphic sequence in the Corner Brook Lake Belt, overlain by the siliciclastic South Brook Formation of the Mount Musgrave Group, succeeded by the carbonate-dominated Breeches Pond Formation and overlying undifferentiated massive carbonates. In the Humber Arm Allochthon the basal siliciclastic Summerside Formation is overlain by the siliciclastic Irishtown and mixed carbonate and siliciclastic Pinchgut Formation. The Carbonate Belt consists of carbonate-dominated sequences with clastic rocks of the Reluctant Head Formation overlain by massive carbonate sequences which include the Port au Port, St. George, and Table Head groups.

The change from the lower siliciclastic-dominated sequences, along with associated igneous activity in the Corner Brook Lake Belt, to the upper carbonate-dominated sequence is interpreted to correspond with the rift-drift transition of modern continental margins.

Dunnage Zone units include plutonic ultramafic to mafic rocks of the Grand Lake Complex, dated by U-Pb zircon from trondhjemite at 490 ± 4 Ma, volcanic and epiclastic rocks of the Glover Island Formation, and the Matthews Brook Serpentinite. The serpentinite unit is restricted to fault slivers within the Humber Zone sequence, whereas the other two units lie east of the Cabot Fault system. Lithological character and ordering of rock units within the Grand Lake Complex suggest it represents a disrupted ophiolite, although the upper segments of a normal ophiolite sequence, including the sheeted dyke complex and overlying pillow basalts are absent, presumably due to removal during subsequent faulting at the top of the complex along the Kettle Pond shear zone. The range in composition of the igneous rocks of the Glover Formation from mafic to felsic, the low-K tholeiite composition of these rocks, and the evidence for at least some explosive igneous activity, indicated by the presence of tuffs and associated volcaniclastic sedimentary rocks, suggest the formation accumulated in an island-arc setting.

Units postdating the Humber and Dunnage tectonostratigraphic zones include the Glover Island Granodiorite, dated by U-Pb zircon and titanite at 440 ± 2 Ma, and associated plutons (Little Paddle Point, Island Pond, and Red Indian Brook plutons), and an unnamed pegmatite and granitoid intrusive into internal Humber Zone lithologies, dated by U-Pb zircon at 434 +2/-3 Ma.

Island (Zone de Dunnage) et le bassin de Deer Lake. Le bassin de Deer Lake du Carbonifère repose en discordance sur les unités lithotectoniques plus anciennes, lesquelles sont délimitées par des contacts structuraux.

Dans le complexe de Corner Brook Lake de la ceinture du même nom, les gneiss et les amphibolites granitoïdes du Précambrien, à texture massive et rubanée, forment la plus ancienne unité lithostratigraphique de la Zone de Humber; ils constituent le socle cristallin d'une séquence stratigraphique datant du Précambrien tardif à l'Ordovicien moyen. Une datation U/Pb sur zircon du gneiss rubané permet de lui assigner un âge de 1 510 ± 6 Ma. Les roches ignées siliceuses et mafiques du pluton de Lady Slipper (555 + 3/- 5 Ma, datation U/Pb sur zircon) s'observent à la base de la séquence stratigraphique de la ceinture de Corner Brook Lake; suivent ensuite, vers le haut, la formation silicoclastique de South Brook (associée au Groupe de Mount Musgrave) et la Formation de Breeches Pond (dominée par des roches carbonatées). Le tout est surmonté de roches carbonatées massives non différenciées. Dans l'Allochtone de Humber Arm, la formation silicoclastique de Summerside, à la base, est recouverte des formations d'Irishtown (roches silicoclastiques) et de Pinchgut (mélange de roches carbonatées et silicoclastiques). La ceinture de roches carbonatées est constituée de séquences dominées par ce type de roches, les clastites de la Formation de Reluctant Head étant surmontées de séquences de roches carbonatées massives dont font partie les groupes de Port au Port, de St. George et de Table Head.

Le passage, d'une part, des séquences inférieures principalement silicoclastiques (accompagnées de signes d'activité ignée dans la ceinture de Corner Brook Lake) à, d'autre part, la séquence supérieure dominée par les roches carbonatées correspondrait à la transition rift-dérive des marges continentales modernes.

Les unités de la Zone de Dunnage comprennent des roches plutoniques (ultramafiques à mafiques) du complexe de Grand Lake (490 ± 4 Ma, datation U/Pb sur grains de zircon extraits de trondhjémites), des roches volcaniques et épiclastiques de la Formation de Glover Island, de même que la serpentinite de Matthews Brook. L'unité de serpentinite est confinée à des segments de faille observés dans la séquence de la Zone de Humber, tandis que les deux autres unités se trouvent à l'est du système de failles de Cabot. Le caractère lithologique et la disposition des unités du complexe de Grand Lake indiquent qu'elles constituent une ophiolite disloquée, malgré l'absence des parties supérieures d'une séquence ophiolitique normale, en l'occurrence le complexe filonien et les basaltes en coussins sus-jacents; ces unités ont présumément disparu pendant le jeu ultérieur de failles au sommet du complexe, le long de la zone de cisaillement de Kettle Pond. La composition mafique à felsique des roches ignées de la Formation de Glover, leur composition tholéiitique pauvre en potassium ainsi que la présence de tufs et de roches sédimentaires volcanoclastiques (qui prouve qu'il y a eu une certaine activité ignée explosive) indiquent que la formation a été déposée dans un milieu d'arc insulaire.

Parmi les unités ultérieures aux zones tectonostratigraphiques de Humber et de Dunnage, il y a la granodiorite de Glover Island (440 ± 2 Ma, datation U/Pb sur zircon et titanite) et les plutons associés (ceux de Little Paddle Point, d'Island Pond et de Red Indian Brook), ainsi qu'un intrusif sans nom de pegmatite et de granitoïde recoupant les lithologies de la Zone de Humber interne (434 +2/ -3 Ma, datation U/Pb sur zircon).

Red to olive green siliciclastic rocks of the Carboniferous Anguille and Deer Lake groups were deposited in the Deer Lake Basin, which represents a pull-apart basin formed during Carboniferous strike-slip movement on the Cabot Fault system.

Each of the lithotectonic belts has a distinct structural history indicating that they were in different tectonic positions during at least part of their development, but sufficient similarity exists between generations of structures in the different belts to enable correlation. The overall style and sequence of deformation in the Humber Arm Allochthon and the Carbonate Belt are similar, with early west-vergent folds related to thrusting, followed by east-vergent folding and thrusting and the development of west-vergent F_3 crenulation and fracture cleavage. These two belts were juxtaposed by late extensional faulting on the Hughes Brook fault which also caused extensional faulting throughout the belts. Pre-D_1 brittle thrusting and the pre-D_1 foliation development in the Corner Brook Lake Belt are possibly related to a single event. The west-directed thrusting is either correlated with the west-directed thrusting of the Humber Arm Allochthon, and possibly a Taconian event, or it is the start of the Silurian tectonic development, and the onset of the build-up of a thrust wedge which subsequently underwent ductile deformation. The D_1 deformation in the Corner Brook Lake Belt is mainly a result of west-directed movement. It is tentatively correlated with the D_2 event in the two western belts because of the similarity of the style of deformation. D_2 and D_3 in the Corner Brook Lake Belt are progressive developments of the overall west-vergent deformation, and can be correlated with the west-vergent D_3 deformation in the Humber Arm Allochthon and Carbonate Belt. The last deformation in the Corner Brook Lake Belt is D_4 folding, which resulted in lateral extension. Because of the extensional nature of this deformation, it is tentatively correlated with the extensional faulting in the two western belts, and is also related to the extensional Humber River Fault. D_4 folding subsequently deformed this fault.

The earliest deformational features in the Glover Island Belt are limited to foliated xenoliths in the Glover Island Granodiorite. This foliation predates the 440 Ma intrusive age of the granodiorite and may be related to the Taconian Orogeny. The main shear zones and foliation in the Glover Island Belt postdate the granodiorite, and may be similar to the 430 Ma deformation in the Corner Brook Lake Belt. The D_1 deformation in these two belts is correlated, based on age data and the apparent similarity of the S_1 foliation in the slice of Humber Zone rocks on Glover Island and the surrounding Glover Island Belt rocks. Consequently, the two belts were juxtaposed by the end of the D_1 deformation, and so F_2 and F_3 folding are also correlated. Final juxtaposition of the two belts to their present positions occurred during movement on the Cabot Fault. This Carboniferous faulting, which caused the deposition and deformation of the sediments in the Deer Lake Basin, is presumed to be responsible for the

Les roches silicoclastiques rouge à olive des groupes d'Anguille et de Deer Lake du Carbonifère ont été déposées dans le bassin de Deer Lake. Cette dépression est un bassin d'extension, résultant du mouvement de coulissage du système de failles de Cabot au Carbonifère.

Chacun des secteurs lithotectoniques a une évolution structurale distincte, ce qui indique qu'ils se trouvaient dans des positions tectoniques différentes pendant au moins une partie de leur formation, mais il y a suffisamment de rapprochements entre les générations de structures de chaque secteurs pour permettre d'établir une corrélation. Dans l'ensemble, le style et la séquence des déformations dans l'Allochtone de Humber Arm et la ceinture de roches carbonatées sont semblables; il y a d'abord eu formation précoce de plis à vergence ouest par charriage, suivie de mouvements de plissement et de charriage à vergence est, puis de la formation de clivages de crénulation P_3 (à vergence ouest) et de fracture. Ces deux secteurs ont été juxtaposés par les mouvements de la faille d'extension tardive de Hughes Brook, lesquels ont eu des répercussions sur l'ensemble des secteurs. Dans la ceinture de Corner Brook Lake, le charriage (déformation cassante) et la formation de la foliation sont antérieurs à D_1 et sont peut-être reliés au même événement. Ou bien le charriage de direction ouest a un lien avec le charriage de même direction de l'Allochtone de Humber Arm (vraisemblablement associé à l'orogenèse taconique), ou bien il marque le début de l'évolution tectonique du Silurien et de la formation d'un biseau de charriage, qui a ensuite subi une déformation ductile. Dans la ceinture de Corner Brook Lake, la déformation D_1 découle surtout d'un mouvement de direction ouest. On a tenté de la relier à la déformation D_2 dans les deux secteurs occidentaux à cause de la similarité des styles de déformation. Toujours dans ce secteur, D_2 et D_3 sont des étapes successives de la déformation globale à vergence ouest et peuvent être corrélées avec la déformation D_3 de même vergence dans l'Allochtone de Humber Lake et la ceinture de roches carbonatées. La dernière déformation dans la ceinture de Corner Brook Lake est le plissement D_4, qui est le résultat d'une extension latérale. Comme cette déformation s'est produite par extension, on a tenté de la corréler avec la formation des failles d'extension dans les deux secteurs occidentaux, mais aussi de la relier à la faille d'extension de Humber River qui a été ensuite déformée par le plissement D_4.

Dans la ceinture de Glover Island, les plus anciens éléments de déformation sont limités à des xénolites foliés dans la granodiorite du même nom. Cette foliation est antérieure à l'intrusion de la granodiorite, il y a 440 Ma, et pourrait être reliée à l'orogenèse taconique. Toujours dans la ceinture de Glover Island, les principales zones de cisaillement et structures de foliation sont postérieures à la mise en place de la granodiorite, et pourraient être similaires à la déformation survenue il y a 430 Ma dans la ceinture de Corner Brook Lake. La déformation D_1 dans ces deux secteurs est corrélée sur considération de datations, mais aussi de l'apparente similarité entre la foliation S_1 dans le lambeau de roches de la Zone de Humber sur l'île Glover et dans les roches environnantes de la ceinture de Glover Island. Par conséquent, les deux ceintures ont été juxtaposées à la fin de la déformation D_1, et les plissements P_2 et P_3 sont aussi corrélés. La juxtaposition finale des deux ceintures dans leurs positions actuelles s'est produite pendant les mouvements le long de la faille de Cabot. Ces mouvements remontent au Carbonifère; ils ont entraîné le dépôt et la déformation des sédiments dans le bassin de Deer Lake, et seraient responsables des failles cassantes les plus latest brittle faults in all lithotectonic units, with the exception of the Humber Arm Allochthon, where Carboniferous deformational features are not recognized.

The Upper Proterozoic to lower Paleozoic rocks in the map area have been affected by one regional metamorphic event, which in the Humber Zone ranged from lower greenschist facies in the west to intermediate pressure amphibolite facies east and south of Corner Brook Lake. The rocks in the Glover Island Belt are in greenschist facies (Knapp, 1982). The thermal imprint on the rocks of the Deer Lake Basin was of subgreenschist grade and reflects depth of burial (Hyde et al., 1988).

Based on distribution of mineral assemblages in pelite, semipelite, and psammite, five metamorphic zones were defined in the Humber Zone. These are from west to east, in order of increasing metamorphic grade: 1) muscovite-chlorite; 2) biotite; 3) garnet; 4) staurolite; 5) biotite-garnet-kyanite. Analysis of the distribution of these zones on a petrogenetic grid allows a semiquantitative estimation of their P-T conditions. Zone 1, representing the Humber Arm Allochthon and the Carbonate Belt, experienced pressures and temperatures that were lower than about 3 to 4 kbar and 440°C, presumably decreasing to the west. The western margin of the Corner Brook Lake Belt forms the biotite zone and the P and T were above 3-4 kbar and 440°C, but probably no higher than about 4 kbar and 480°C as determined from the garnet isograd. Most of the Corner Brook Lake Belt is in the garnet zone, and the maximum temperature is defined by the staurolite-in isograd, which in the field represents conditions of about 5 to 7 kbar at 570° to 580°C. Maximum temperatures within the staurolite zone were probably around 650°C, requiring pressure estimates of up to 7 to 9 kbar. These also provide minimum estimates of P-T conditions for Zone 5 assemblages.

Peak metamorphic conditions were attained between the D_1 and D_2 deformation. Metamorphism is interpreted to result from crustal thickening by thrust stacking (pre-D_1 and D_1), but the isotherms were deformed during ensuing thrust movement (D_2 and D_3). U-Pb and Ar-Ar isotopic age data from the Corner Brook Lake region indicate that regional deformation and peak amphibolite facies metamorphism in the eastern Humber Zone is Early Silurian. A lower limit on deformation is provided by the U-Pb zircon age of 434 +2/-3 Ma for a pegmatite affected by regional foliation and is interpreted to be syntectonic. Monazite and rutile from a garnet-kyanite-staurolite schist, which records peak metamorphic conditions and in which porphyroblasts overgrow the regional foliation, gave U-Pb ages of 430 ± 2 Ma and 437 ± 6 Ma, respectively. Ar-Ar cooling ages for hornblende from amphibolites and muscovite from psammitic and pelitic schists range from 430 to 420 Ma.

Metamorphic mineral assemblages and textural data indicate that the rocks of the Dunnage Zone were affected by a single regional metamorphic event of

tardives dans toutes les unités lithotectoniques, sauf dans l'Allochtone de Humber Arm où aucune structure de déformation du Carbonifère n'a été observée.

Les roches du Protérozoïque supérieur au Paléozoïque inférieur de la région à l'étude ont connu un épisode de métamorphisme régional qui, dans la Zone de Humber, variait du faciès des schistes verts inférieur, dans l'ouest, au faciès des amphibolites (pression intermédiaire), à l'est et au sud du lac de Corner Brook. Les roches de la ceinture de Glover Island sont métamorphisées au faciès des schistes verts (Knapp, 1982). L'empreinte thermique laissée sur les roches du bassin de Deer Lake indique des conditions de métamorphisme inférieurs à celles du faciès des schistes verts et témoigne de la profondeur d'enfouissement (Hyde et coll., 1988).

D'après la répartition des associations de minéraux dans la pélite, la semipélite et la psammite, la Zone de Humber a été divisée en cinq zones de métamorphisme croissant (d'ouest en est); ce sont les suivantes : (1) muscovite-chlorite; (2) biotite; (3) grenat; (4) staurotide; (5) biotite-grenat-kyanite. L'analyse de la répartition de ces zones sur une grille pétrogénétique permet une estimation semi-quantitative des conditions de pression et de température de chacune d'elles. La zone 1, représentant l'Allochtone de Humber Arm et la ceinture de roches carbonatées, a été le siège de pressions et de températures inférieures à 3 ou 4 kbar et à 440 °C environ, lesquelles diminueraient vers l'ouest. La marge occidentale de la ceinture de Corner Brook Lake forme la zone à biotite, et les conditions de pression et de température étaient d'au moins 3 à 4 kbar et 440 °C, sans vraisemblablement dépasser environ 4 kbar et 480 °C comme l'indique l'isograde du grenat. La plus grande partie de la ceinture de Corner Brook Lake se trouve dans la zone à grenat, et la température maximale est donnée par l'isograde interne de la staurotide, qui témoigne de conditions de pression et de température d'environ 5 à 7 kbar et de 570 à 580 °C. Les températures maximales dans la zone à staurotide se situaient probablement aux environs de 650 °C, de sorte que les pressions devaient atteindre de 7 à 9 kbar. Cela permet d'établir aussi des estimations de la pression et de la température minimales pour les assemblages de la zone 5.

Le métamorphisme a été à son maximum entre les déformations D_1 et D_2. Il aurait résulté d'un épaississement de la croûte découlant d'un empilement par chevauchement (antérieur à D_1 et contemporain de D_1); les courbes isothermes montrent cependant une déformation pendant l'épisode de charriage ultérieur (D_2 et D_3). Les données de datation isotopique U/Pb et Ar/Ar de roches provenant de la région du lac Corner Brook indiquent que la déformation régionale et le métamorphisme maximale au faciès des amphibolites dans la partie orientale de la Zone de Humber remontent au Silurien précoce. Une datation U/Pb sur zircon d'une pegmatite marquée par la foliation régionale a permis d'établir une limite inférieure de déformation à 434 + 2/- 3 Ma, laquelle serait syntectonique. La monazite et le rutile d'un schiste à grenat-kyanite-staurotide, qui représente les conditions de métamorphisme maximal et dans lequel des porphyroblastes recoupent la foliation régionale, remonteraient à 430 ± 2 et 437 ± 6 Ma respectivement (datation U/Pb). Les âges Ar/Ar de refroidissement de la hornblende des amphibolites et de la muscovite des schistes psammitiques et pélitiques varient entre 430 et 420 Ma. Les assemblages de minéraux métamorphiques et les données texturales indiquent que les roches de la Zone de Dunnage n'ont connu qu'un seul épisode de métamorphisme régional, essentiellement au faciès des schistes verts. Le métamorphisme aurait eu lieu pendant la déformation D_1 de la largely greenschist grade. Metamorphism is interpreted to have taken place during D_1 deformation of the Glover Island Belt. A thermal signature on the Deer Lake Basin due to diagenetic burial metamorphism suggests that rocks of the Anguille Group experienced maximum temperatures of 200°C whereas the Deer Lake Group did not exceed 100°C (Hyde et al., 1988).

ceinture de Glover Island. Dans le bassin de Deer Lake, la signature thermique résultant d'un métamorphisme d'enfouissement diagénétique indique que les roches du Groupe d'Anguille ont été soumises à des températures maximales de 200 °C, tandis que dans le Groupe de Deer Lake, celles-ci n'ont pas dépassé 100 °C (Hyde et coll., 1988).

INTRODUCTION

The Corner Brook-Glover Island region lies within the western part of the Appalachian Orogen in Newfoundland. The region straddles the boundary between the ancient continental margin, or miogeocline, of North America (Laurentia) and tectonostratigraphic terranes accreted to the margin during collisional orogenesis (Fig. 1A). This boundary forms a fundamental structure in the northern Appalachians and is termed the Baie Verte Line (Fig. 1B; Williams and St-Julien, 1978, 1982). Rocks west of the line lie within the Humber Zone (Fig. 1B) of Williams (1978, 1979) and consist of upper Precambrian to middle Ordovician carbonates and siliciclastic strata with minor igneous rocks, unconformably overlying Mesoproterozoic gneissic basement. Immediately east of the line the Dunnage Zone consists of Cambrian to middle Ordovician ultramafic to mafic and silicic igneous rocks and associated epiclastic strata interpreted to have formed in a supra-subduction zone, intra-oceanic setting (Williams, 1979; Swinden et al., 1990; Dunning et al., 1991). Initial juxtaposition of the Humber and Dunnage zones is inferred to have taken place during the Ordovician Taconian Orogeny, with the Baie Verte Line interpreted as the root zone for klippen of Dunnage Zone preserved in the Taconian allochthons which structurally overlie the miogeocline (Fig. 1A; Williams and St-Julien, 1978, 1982). In the Corner Brook-Glover Island region, this original boundary has been largely overprinted by later deformation, and the Humber and Dunnage zones are now separated along the western arm of Grand Lake by the Carboniferous Cabot Fault system (Fig. 1B). Williams and St-Julien (1982; Knapp, 1982) considered that original Ordovician relations between these two zones within the map area are preserved on Glover Island. Carboniferous sedimentary rocks of the Deer Lake Basin are developed along, and adjacent to, the Cabot Fault system in the northeastern corner of the Corner Brook-Glover Island region.

The Humber Zone is divided into a western external domain and an eastern internal domain on the basis of increasing deformation and metamorphism towards the eastern orogenic hinterland of the Appalachian Orogen (Fig. 2; Williams, in press). In the Corner Brook-Glover Island region, the external domain consists of the Humber Arm Allochthon and a carbonate dominated cover succession. Transported rocks of the Humber Arm Allochthon are exposed along the northwest margin of the map area within, and west of, the town of Corner Brook. Carbonates and minor siliciclastics of the cover sequence form a north-trending belt from east of the town towards Corner Brook Lake. Polydeformed, greenschist- to amphibolite-facies sedimentary and igneous rocks of the internal domain of the Humber Zone, which lie unconformably on Grenville basement, form the dominant rock assemblage of the map area (Fig. 2). This sequence extends from east of Pinchgut Lake to Grand Lake in the east and from the southwestern arm of Grand Lake north to the Northern Harbour road and the Trans Canada Highway. The assemblage is correlated with the Fleur de Lys Supergroup on the Baie Verte Peninsula (Williams and St-Julien, 1982; Hibbard, 1983a, b). The internal and external domains of the Humber Zone are separated by a structural front (Williams and Cawood, 1989), termed the Humber River Fault.

Igneous and sedimentary rocks of the Dunnage Zone are exposed on Glover Island in Grand Lake, and to the east of the lake. All lie within the Notre Dame subzone of Williams et al. (1988). Minor occurrences of this unit are also found in fault slices along the western shore of the lake.

Late Ordovician to Devonian silicic igneous rocks are locally intrusive into the mid-Ordovician and older lithologies of the Humber and Dunnage zones. These later intrusives form part of a suite of igneous and sedimentary rocks which are wide-spread throughout the central segment of the Newfoundland Appalachians (Fig. 1B; Williams et al., 1989a,b; Colman-Sadd et al., 1990).

Carboniferous sedimentary rocks of the Deer Lake Basin unconformably overlie the juxtaposed segments of the Humber and Dunnage zones in the northeast portion of the map area (Fig. 2; Hyde, 1982). The distribution of Late Ordovician and younger rock units of the Newfoundland Appalachians is largely independent of the tectonostratigraphic zonal subdivision of the orogen which is based on differences in mid-Ordovician and older rocks units, predating the juxtaposition of the zones (Williams, 1979; in press).

This report details stratigraphic, structural, and metamorphic relationships within the Corner Brook map sheet (12A/13) and adjoining segments of the Little Grand Lake and Harrys River map sheets (12A/12 and 12B/9). It provides a detailed description of the rock units, the first integrated structural and metamorphic analysis, and presents the first reliable radiometric data on the rock units and events within the area. These data allow development of a new model for the timing and nature of interaction between the Humber and Dunnage zones and have significant implications for the timing of shear-hosted gold mineralization in the region.

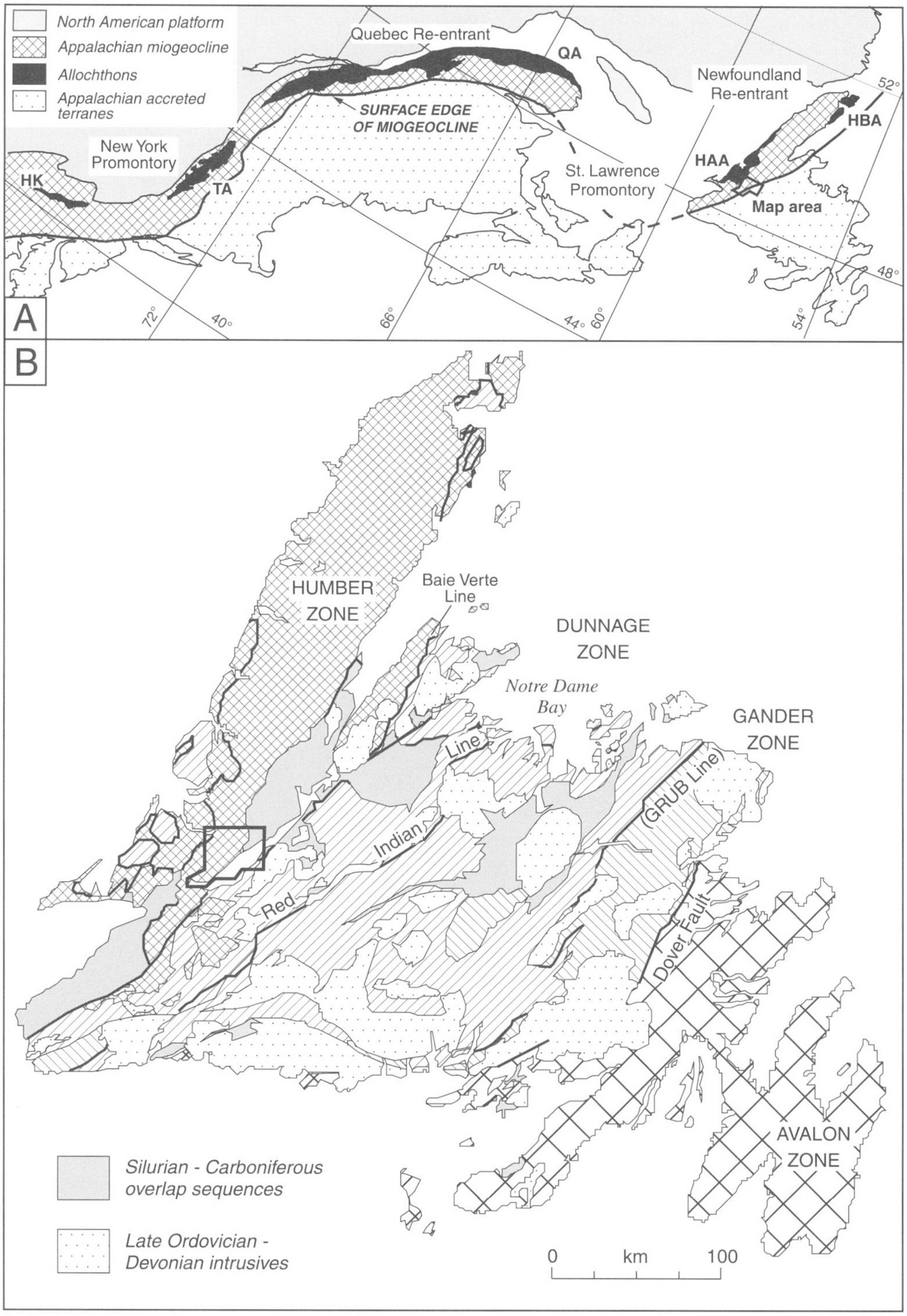


*Figure 1. Position of Corner Brook region in the western Newfoundland Appalachians. Box outlines location of Corner Brook-Glover Island region in both maps. **A**) Distribution of promontories and re-entrants of Appalachian miogeocline and accreted terranes. Abbreviations: HK – Hamburg Klippe; TA –Taconic Allochthon; QA – Quebec Allochthon; HAA – Humber Arm Allochthon; HBA – Hare Bay Allochthon. **B**) Geological map of the Newfoundland Appalachians showing distribution of tectonostratigraphic zones and younger igneous and sedimentary overlap successions. Tectonostratigraphic zones delineated by cross hatches and diagonal lines.*

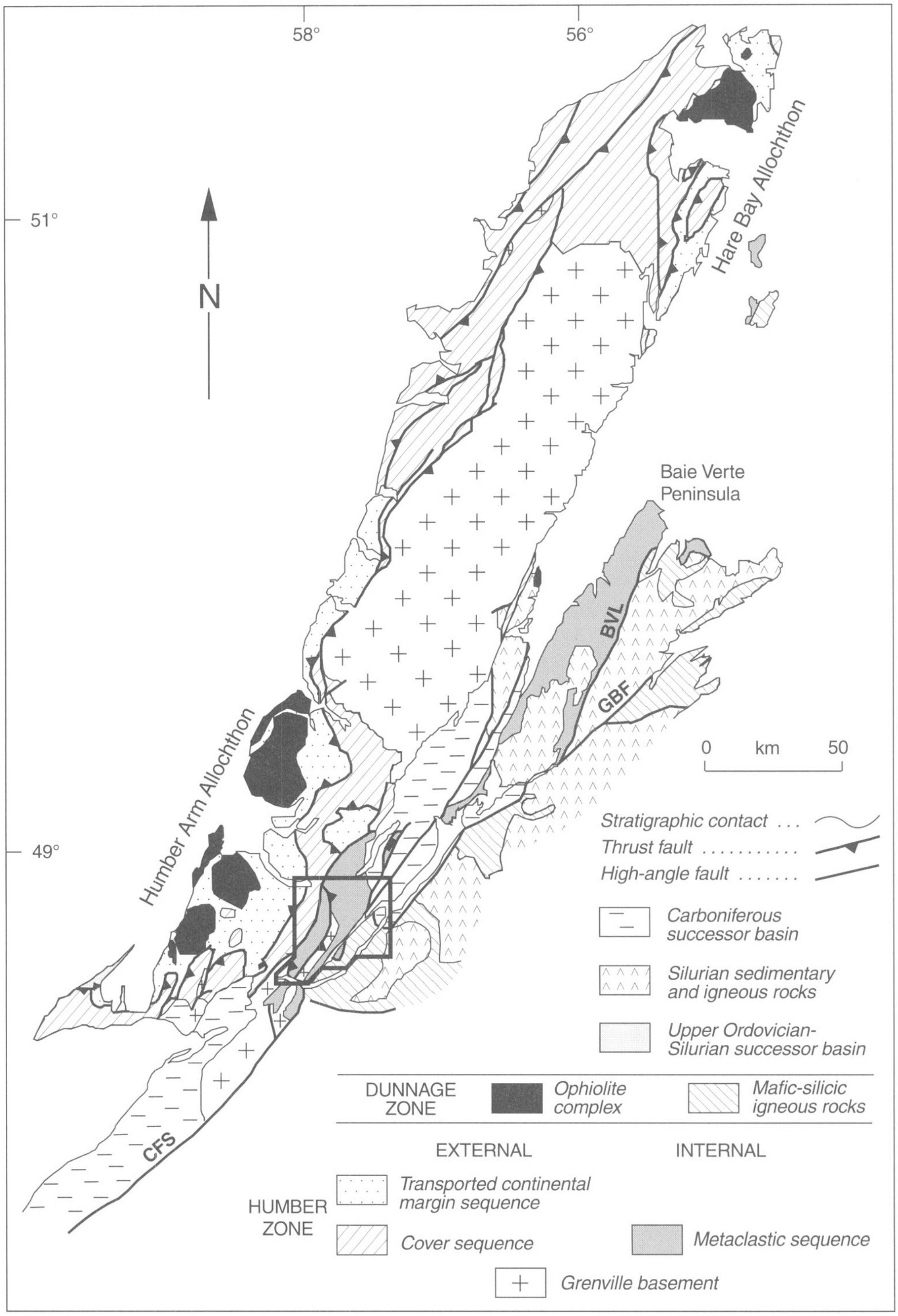


Figure 2. *Simplified geological map of the Humber and western Dunnage zones in west Newfoundland. Box outlines location of the Corner Brook-Glover Island region. Abbreviations: BVL – Baie Verte Line; GBF – Green Bay Fault; CFS – Cabot Fault System.*

Previous work

Previous work within the Corner Brook-Glover Island region focused on the eastern margin of the Humber Arm Allochthon and its relationship with the carbonate succession (Walthier, 1949; Williams and Cawood, 1986, 1989, and references therein), and on the Carboniferous successor basin sequence (Hyde, 1982; Hyde et al., 1988). Mapping within the internal domain of the Humber Zone and the adjoining Dunnage Zone has been restricted to reconnaissance scale work (Fig. 3; Riley, 1957; Baird, 1959; Lilly, 1963; McKillop, 1963; Kennedy, 1980, 1981; Knapp, 1982; Knapp et al., 1979; Martineau, 1980; Hibbard, 1983b; Williams et al., 1982, 1983, 1985).

Kennedy (1981) delineated the extent of the metaclastic sequence of the internal domain between Deer Lake and Grand Lake and erected an informal stratigraphic division of the sequence. Williams et al. (1982, 1983, 1985) mapped the metaclastic sequence in the Pasadena map sheet (12H/4) northwest of the Corner Brook sheet and formalized the stratigraphic divisions initially recognized by Lilly (1963) and McKillop (1963), grouping the metaclastic lithologies into the Mount Musgrave Group, consisting of the Little North Pond Formation and the overlying South Brook Formation. In addition, they established that granitoid rocks stratigraphically underlying the Mount Musgrave Group along its western margin are of late Precambrian age, and along with associated mafic volcanic rocks constitute the Hughes Lake Complex. A pluton of similar age has been recognized at Hare Hill to the south of the Corner Brook map area by van Berkel and Currie (1988; see also Currie et al., 1992). This pluton is intrusive into Precambrian Grenvillian gneiss. Results of mapping in this southern region are summarized by Currie and van Berkel (1992a, b). In addition to the standard Humber and Dunnage zone subdivisions, they recognized a sequence south of Little Grand Lake and east of the Cabot Fault which they termed the Central Gneiss subzone, consisting of metasediments containing pods of mafic and ultramafic rock, all cut by gneissic to massive granitoids. The Central Gneiss subzone (termed the Dashwoods subzone by Williams, in press) is separated from the Notre Dame subzone by the Little Grand Lake fault (Whalen et al., 1993), which shows an older ductile history of south-over-north movement and a younger, brittle history of south-over-north movement. Williams (in press) included the Dashwoods subzone within the Dunnage Zone, but the lithological grouping within the region indicates it is better viewed as a mixture of both Humber Zone and Dunnage Zone sequences which were subsequently structurally intercalated (Fox and van Berkel, 1988; Currie and van Berkel, 1992a, b).

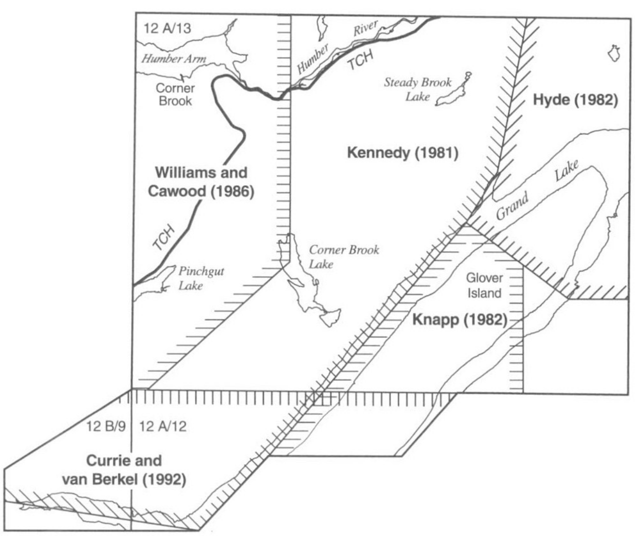


Figure 3. *Outline of Corner Brook-Glover Island region showing areas mapped by previous workers. TCH – Trans Canada Highway.*

Riley (1957) produced the first regional map which incorporated the Glover Island region. He noted the presence of psammite and pelite on the west side of the island and volcanic rocks in the east, separated by ultramafic and mafic plutonic rocks, which he interpreted as intrusions. Williams and St-Julien (1978, 1982) showed that the plutonic rocks represent an ophiolitic fragment and the structural contact with the metasedimentary rocks to the west marked the trace of the Baie Verte Line in this region. Knapp (1982) carried out the first detailed mapping of Glover Island, concentrating on its southern and central portions. He divided the metasedimentary sequence in the west into a basement assemblage of quartzofeldspathic gneiss and a structurally overlying sequence of psammite, pelite, amphibolite, and minor carbonate. He interpreted the volcanic rocks on the east side of the island to be nonconformable on the ophiolitic sequence. Riley (1957) and Knapp (1982) also noted the presence of younger silicic plutonic rocks intrusive into the Dunnage Zone sequence on the eastern side of Glover Island and along the eastern shore of the lake.

Hyde (1978, 1979a, b; 1982; Hyde, et al., 1988) carried out a detailed study of the stratigraphy and sedimentology of the Carboniferous Deer Lake Basin, the southern end of which outcrops across the northeast end of the Corner Brook-Glover Island region. This work erected the stratigraphic nomenclature for the basin and determined the relationship and setting of the Carboniferous rock units to the surrounding basement blocks.

Physiography and access

The region has moderate topography with elevation ranging from sea level at Humber Arm to over 650 m on the highlands of the Long Range Mountains between Corner Brook Lake and Grand Lake. Physiography is lithologically controlled with the thickly wooded hills of the western third of the map area corresponding with the Humber Arm Allochthon and carbonate platform sequence. Barren highlands occupying the central third of the area extend from Corner Brook Lake and Mount Musgrave in the west to Grand Lake and the valley of South Brook in the east, and are underlain by gneiss and metaclastic lithologies. The rounded but forested hills of the northeast corner of the map area correspond with the Carboniferous Deer Lake Basin. The open, boggy highlands of Glover Island and east of Grand Lake correspond with the Dunnage Zone and Silurian intrusive rocks. The present topography of the Corner Brook-Glover Island region represents a glacially modified dissected peneplain (Twenhofel and MacClintock, 1940; Brookes, 1970, 1973). Glaciation is considered to be Late Wisconsinan with the main ice flow direction swinging from an early southwest direction along the major fiords of Grand Lake and Humber River gorge to an overall westerly direction during the later stages of ice movement (Batterson and Taylor, 1990; Batterson and Vatcher, 1992; Batterson and McGrath, 1993). Glacial debris ranges from isolated erratics on the highlands to a relatively widespread veneer of till and hummock structures found along stream valleys (e.g. Corner Brook, South Brook, and Steady Brook), the area east of Pinchgut Lake, and along the eastern side of Glover Island and east of Grand Lake.

Access to large parts of the map area is provided by a network of major roads and forestry tracks. The Trans Canada Highway traverses the northern and western part of the area, running through Corner Brook, the second largest town in Newfoundland with a population of approximately 30 000. A system of logging roads, developed to supply pulp wood to the paper mill in the town, provides access to the central and southern parts of the map area, as well as Glover Island and some areas east of Grand Lake. The tracks vary considerably; in the north they are largely overgrown and bridges are washed out; in the central and southern parts around Corner Brook Lake, logging was being actively carried out while mapping was underway and roads were open to vehicular traffic; new roads were constantly added; on Glover Island and farther east logging had ceased in the early 1980s and roads were deteriorating rapidly. The Northern Harbour Road from Pasadena provides access to Grand Lake. Transmission lines trending east provide additional easy access to the area east of Corner Brook.

Acknowledgments

This project was undertaken through the Canada-Newfoundland Mineral Development Agreement 1990-1995 and was supported by the Department of Supply and Services (for the Geological Survey of Canada), an Energy, Mines and Resources Research Agreement, and a NSERC Operating

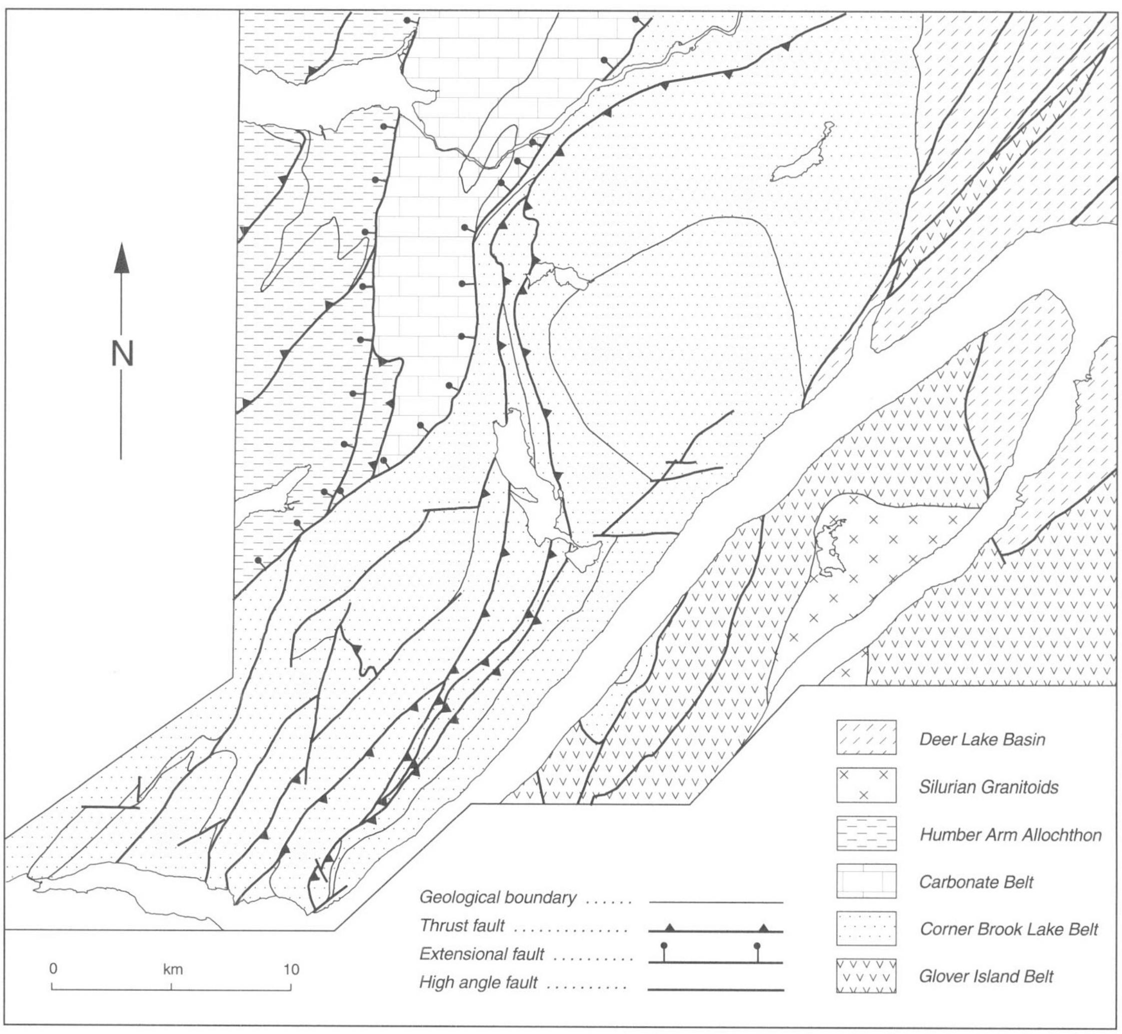


Figure 4. *Distribution of major lithotectonic units in the Corner Brook-Glover Island region.*

Grant. Ken L. Currie acted as scientific authority for the project and we thank him for his enthusiasm and support. Jim Ryan, Roger March, and Robert Taylor provided field support, and we particularly thank Jim Ryan for his assistance. Doug Boyce, Steve Colman-Sadd, Ken Currie, Greg Dunning, Rebecca Jamieson, Ian Knight, Roger Mason, Mark Piasecki, Toby Rivers, and Geoff Thurlow visited the field area and provided stimulating discussion. Aphrodite Indares provided insightful and helpful comments on the metamorphic section of this report. Mineral separates for radiometric dating were expertly prepared by Kathy Manser, Jim Ryan, and Dave West Jr. We are indebted to Greg Dunning for U-Pb ages and Dan Lux for Ar-Ar ages. The city of Corner Brook and Corner Brook Pulp and Paper provided access to private roads, and Alvin Harris and Stuart Cochrane of the Newfoundland Department of Mines and Energy allowed use of facilities at the Pasadena core storage library. This project was undertaken while the authors were at the Departmnet of Earth Sciences, Memorial University of Newfoundland. We thank the department and its staff for its support and access to facilities. This is ARC Tectonics Special Research Centre Publication No. 9.

GENERAL GEOLOGY

The distribution and interrelationship of rock units within the Corner Brook-Glover Island region are shown in map and cross-sections (Map 1893A, in pocket). The rocks constitute five lithotectonic units which from west to east are (Fig. 4): Humber Arm Allochthon, incorporating both the traditional rocks of the allochthon as well as a newly recognized sequence of structural slices at the base of the allochthon; Carbonate Belt, consisting of the carbonate dominated cover sequence within the external domain of the Humber Zone; Corner Brook Lake Belt, composed of metamorphosed siliciclastic and carbonate lithologies within the internal domain of the Humber Zone; Glover Island Belt of mafic to silicic igneous and volcaniclastic rocks, which is part of the Dunnage

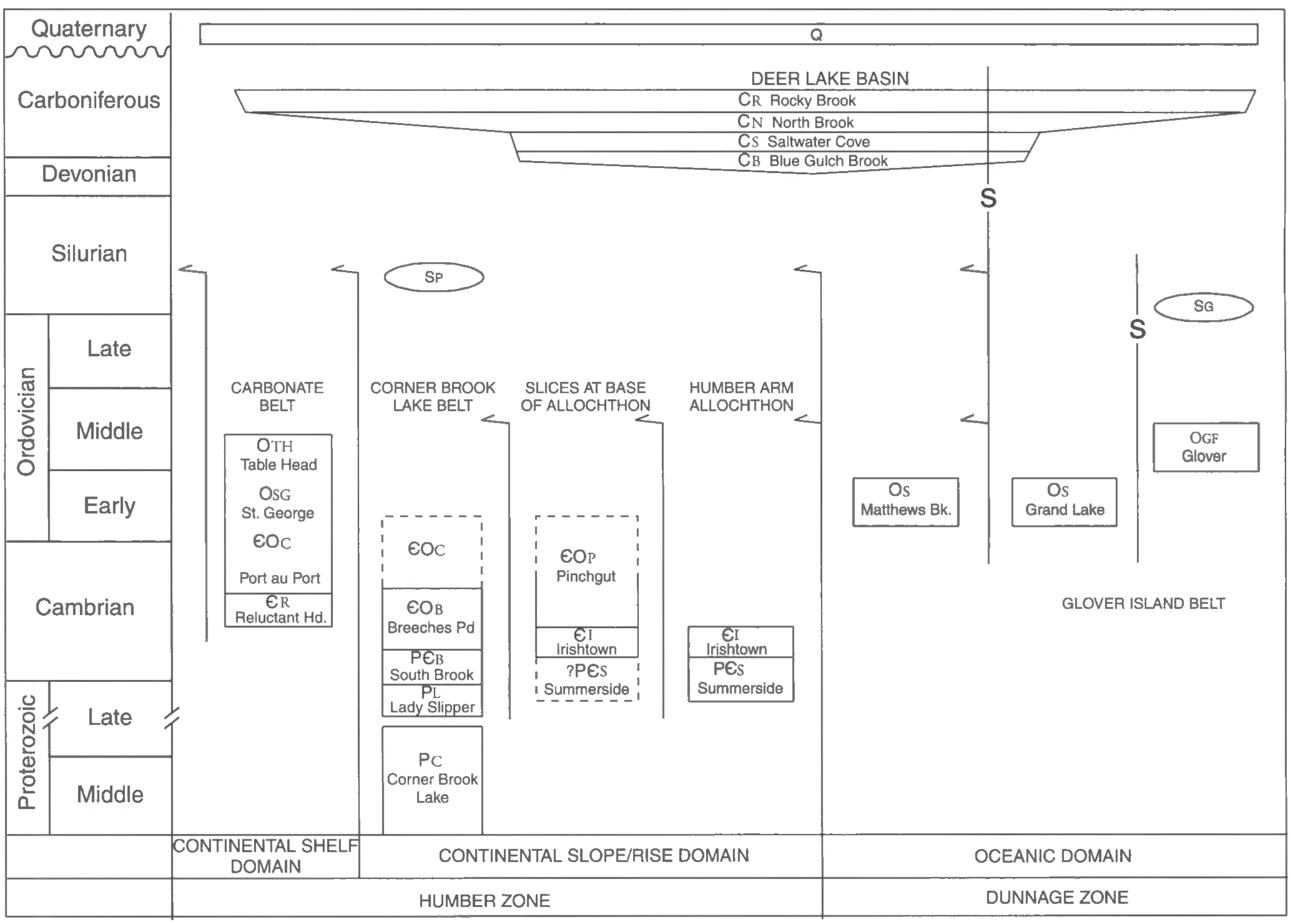


Figure 5. *Stratigraphic correlation chart and palinspastic restoration of rock units within the Corner Brook Lake region. Columns indicate age span or inferred age span of rock units. Arrows indicate overall direction of structural transport toward and onto the ancient continental margin that led to the present stacking order. Upper limit of arrows indicates approximate time of transport. Vertical lines marked with an "S" indicate strike-slip fault.*

Zone; and the Deer Lake Basin, a Carboniferous strike-slip basin containing clastic sediments. Correlation of rock units within and between the major lithotectonic divisions across the area is shown in Figure 5. The Deer Lake Basin lies unconformably on the older lithotectonic units. All boundaries between these older units are structural.

GEOLOGY OF THE HUMBER ZONE

Precambrian basement

Precambrian gneissic rocks of the Grenville Orogen form the basement to the Appalachian orogenic sequence in the Humber Zone. Basement is well exposed in the central and southern part of the metaclastic belt and is referred to as the Corner Brook Lake Complex.

Corner Brook Lake Complex

Definition and distribution

The Corner Brook Lake Complex consists of quartzofeldspathic granitoid gneiss and interbanded amphibolite, with minor quartzite and quartz-feldspar-mica paragneiss. The name is taken from Corner Brook Lake, around which the gneiss is well exposed. The region encompassing the lower reaches of the unnamed stream draining into the lake from Valley Lakes and the surrounding hills, around grid reference (g.r.) 388108 is designated as the type section. The complex incorporates strata previously mapped by Kennedy (1981) as part of the Mount Musgrave and Caribou Lake formations. It outcrops in a series of north- to northeast-trending bands (see Map 1893A, in pocket). The western band is restricted to

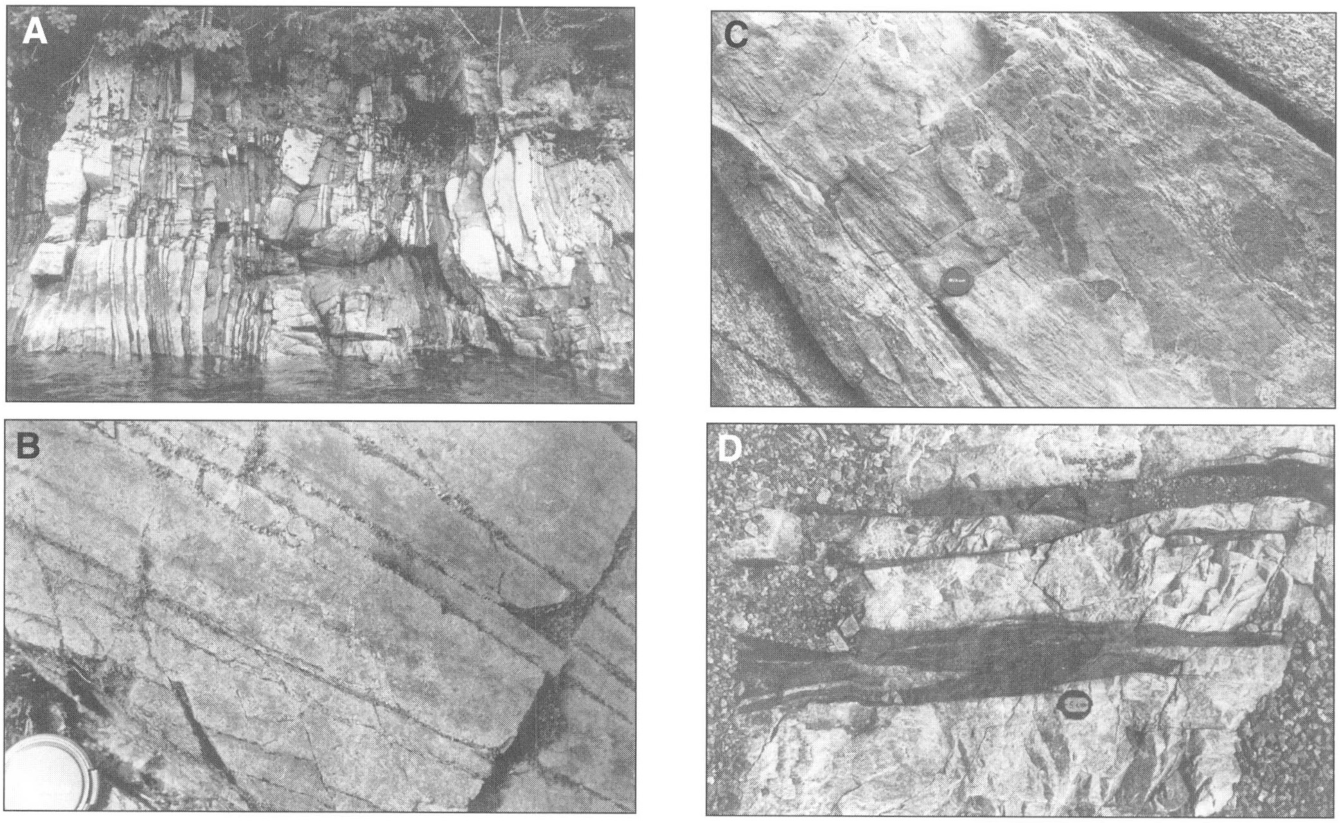


Figure 6. *Gneissic rocks of the Corner Brook Lake Complex (map unit PC).* ***A****) Interbanded felsic and mafic gneiss from western arm of Grand Lake (g.r. 225921). The gneisses are folded in west vergent F_1 folds, which are shown in the right of the photo. A diabase dyke in the centre may form part of the late Precambrian Long Range dyke swarm. View looking north. GSC 1995-149G* ***B****) Foliated homogeneous granitoid gneiss outcropping in stream draining from Valley Lakes into Corner Brook Lake (g.r. 390105). Late normal faults dipping steeply to the east cause small-scale (cm) offset of the shallow east-dipping gneiss fabric. GSC 1995-149B* ***C****) Boudinaged amphibolite dyke in granitoid gneiss cut by pegmatitic veins of Corner Brook Lake Complex (map unit PC; g.r. 343975). Banding of felsic and mafic lithologies and boudinaging of dyke are interpreted as Grenvillian features that predate penetrative Paleozoic foliation (S_1) which runs from upper left to lower right. GSC 1995-149E* ***D****) Interbanded, isoclinally folded homogeneous tonalitic gneiss and amphibolite outcropping at the northern tip of Corner Brook Lake (g.r. 384113). Folding and axial planar fabric are continuous into overlying cover sequence and are interpreted to be Paleozoic (S_1). Pervasive overprinting of the outcrop by the Paleozoic fabric prevents determination of the original relationship between felsic and mafic lithologies and whether the amphibolite is an integral part of the basement complex or a later intrusion. GSC 1995-149F*

the southern end of the map area, and consists of a single northeastward-tapering wedge extending some 6 km from the southwestern arm of Grand Lake. Rocks within this area are best exposed along forestry tracks and on hills north of the arm of Grand Lake. The eastern band can be traced approximately 25 km from the southwestern arm of Grand Lake to north of Corner Brook Lake. The eastern basement band is disrupted by thrust faulting into a number of slices. It is well exposed along streams and forestry tracks around the northeastern side of Corner Brook Lake, along forestry tracks around the southwest margin of the lake, and along streams and forestry tracks west of Little Paddle Point (g.r. 369959). In addition to the two main eastern and western bands of basement, a small block of the Corner Brook Lake Complex, consisting largely of amphibolite, occurs in Matthews Brook in the north. This area lies along strike from the eastern band and represents a structurally isolated sliver of basement. A further small block of the complex occurs on the western side of Glover Island, to the east of the main strand of the Cabot Fault system. This latter locality was informally termed the Cobble Cove gneiss by Knapp (1982), and is well exposed along the shore of Glover Island to the north of Cobble Cove[1] (g.r. 430004) and along the lower reaches of Keystone Brook (g.r. 430004 to g.r. 434004).

Lithology

Granitoid gneiss, the principal lithology of the complex, is cream to pink and consists of varying proportions of quartz, plagioclase, K-feldspar, biotite, and amphibole, with minor epidote and muscovite giving a compositional range from diorite to granodiorite to locally quartz monzonite. Small, centimetre-scale, quartz and K-feldspar pegmatitic sweats predate the S_1 foliation. The granitoid gneiss includes both banded and homogeneous varieties (Fig. 6A-C) but no systematic regional variation in the proportion of these felsic gneiss units was observed, preventing an internal lithological subdivision of the Corner Brook Lake Complex.

Amphibolite layers parallel to the compositional layering in the granitic gneiss range from several centimetres to, at least, 75 m thick (Fig. 6A, D). They consist of amphibole-plagioclase and plagioclase-amphibole-biotite schist with minor garnet and magnetite. Late, large, brassy biotite grains, often associated with euhedral titanite, are common. Biotite and hornblende are locally partially to completely altered to chlorite. Margins of amphibolite bands are either abrupt or gradational against the enclosing granitic gneiss. In the latter case, increasing feldspar content of the rock towards its margins and sometimes the presence of quartz indicate a compositional sequence from amphibolite to diorite and locally quartz diorite.

[1] Informal geographic terms introduced by previous workers are used to help locate data when no established geographic term is available. The terms 'Cobble Cove' and 'Keystone Brook' were introduced by Knapp (1982). 'Grand Lake Brook' refers to the major stream running east to west at the intersection of the Corner Brook, Harry's River and Little Grand Lake map sheets and was introduced by Kennedy (1981).

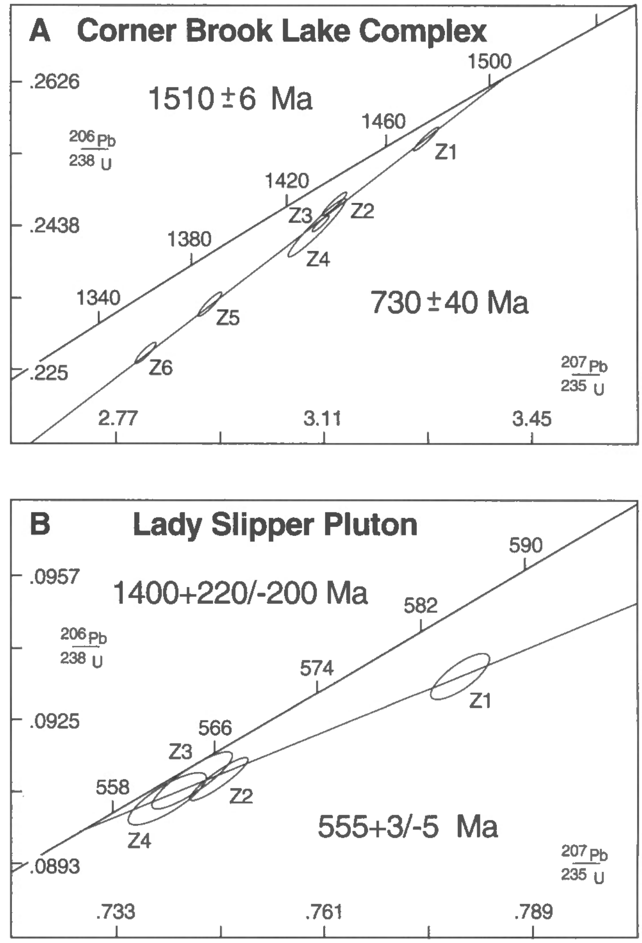


Figure 7. *U-Pb concordia plot for rocks in the Corner Brook Lake Belt.* ***A)*** *Banded granitic gneiss of the Corner Brook Lake Complex.* ***B)*** *Tonalitic gneiss of the Lady Slipper Pluton.*

The relative proportions of amphibolite and granitoid gneiss are highly variable but amphibolite tends to dominate in outcrops north of Corner Brook Lake whereas granitic gneiss forms the bulk of the outcrops to the south. Quartzite and quartz-feldspar-biotite-amphibole paragneiss are rare, localized lithologies within the complex (g.r. 343973). On Glover Island the dioritic gneiss is intruded by little-deformed, pink aplitic to pegmatitic granite dykes.

Age and relations

A sample of banded granitic gneiss from the eastern basement near the eastern shore of Corner Brook Lake (g.r. 387108), collected for a U-Pb age determination, yielded a large amount of coarse grained zircon with a short prismatic habit and rounded crystal edges. Six abraded fractions of the clearest grains were analyzed (Table 1). Z1 and Z6 were the most rounded grains, Z2 and Z4 were the most euhedral. Five fractions (excluding Z2) define a simple discordia line (10% probability of fit) with an upper intercept age of 1510 ± 6 Ma (Fig. 7A). The lower intercept of 730 ± 40 Ma likely reflects the integrated effects of Pb-loss during two events, perhaps

Table 1. U-Pb data.

		Concentration		Measured		*Corrected Atomic Ratios							Age [Ma]		
Fraction Description	Weight [mg]	U [ppm]	Pb rad [ppm]	total common Pb [pg]	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	±	$^{207}Pb/^{235}U$	±	$^{207}Pb/^{206}Pb$	±	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
CORNER BROOK LAKE COMPLEX															
Z1 anhed abr	0.047	240	62.7	8	22940	0.0984	0.25528	106	3.2771	140	0.09311	10	1466	1476	1490
Z2 best small prism abr	0.029	283	71.1	7	17020	0.0905	0.24666	128	3.1265	158	0.09193	18	1421	1439	1466
Z3 clr abr	0.024	321	79.8	9	12500	0.0900	0.24432	80	3.1052	108	0.09218	12	1409	1434	1471
Z4 euhed abr	0.076	289	71.7	28	12091	0.0896	0.24372	302	3.0980	378	0.09219	24	1406	1432	1471
Z5 large clr abr	0.132	286	67.5	17	32481	0.0823	0.23373	118	2.9235	150	0.09072	12	1354	1388	1441
Z6 small clr round abr	0.153	284	65.0	15	42237	0.0794	0.22719	96	2.8184	122	0.08997	12	1320	1360	1425
LADY SLIPPER PLUTON															
Z1 2nd best needles abr	0.135	134	12.6	28	3808	0.1142	0.09348	42	0.7792	32	0.06046	16	576	585	620
Z2 needles abr	0.124	156	14.4	25	4453	0.1130	0.09119	36	0.7468	32	0.05939	10	563	566	582
Z3 needles abr	0.072	126	11.5	28	1870	0.1107	0.09115	50	0.7431	46	0.05912	14	562	564	572
Z4 best needles abr	0.028	141	12.9	23	963	0.1153	0.09074	48	0.7398	42	0.05913	20	560	562	572
GRAND LAKE COMPLEX TRONDHJEMITE															
Z1 clr cracks abr	0.004	228	17.4	51	98	0.0739	0.07851	44	0.6165	206	0.05695	176	487	488	490
Z2 clr frags abr	0.009	260	19.9	16	758	0.0721	0.07892	42	0.6196	36	0.05694	20	490	490	489
GLOVER ISLAND GRANODIORITE															
T1 large lt brn euh abr	0.316	65	8.2	544	183	1.0195	0.07044	22	0.5383	40	0.05542	34	439	437	429
T2 small lt brn abr	0.353	80	9.6	659	207	0.9209	0.07055	34	0.5447	48	0.05600	36	439	442	452
Z1 euh needles abr	0.008	602	43.9	30	734	0.1431	0.07068	48	0.5438	42	0.05580	22	440	441	445
Z2 med lt brn abr	0.115	420	29.9	25	8708	0.1138	0.07089	24	0.5478	20	0.05604	6	442	444	454
PEGMATITE															
Z1 clr euh abr	0.088	113	8.0	17	2805	0.0150	0.07635	32	0.6119	28	0.05813	16	474	485	535
Z2 needles abr	0.085	127	8.2	8	6220	0.0033	0.07123	38	0.5527	30	0.05628	12	444	447	464
Z3 clr euh abr	0.139	136	8.8	28	2976	0.0027	0.07100	32	0.5491	26	0.05609	8	442	444	456
Z4 8 large abr	0.051	1212	77.3	44	6262	0.0002	0.07022	30	0.5397	24	0.05574	6	437	438	442
GARNET - KYANITE SCHIST															
R1 red needles abr	0.083	3	0.4	24	64	1.1510	0.07056	76	0.5506	280	0.05659	266	440	445	476
R2 blocky abr	0.413	4	0.3	106	84	0.0335	0.07006	38	0.5654	114	0.05853	106	436	455	550
M1 yellow fragments abr	0.057	1473	279.0	50	7346	2.1212	0.06896	26	0.5272	21	0.05544	6	430	430	430
M2 yellow fragments abr	0.039	1923	367.9	63	5152	2.1609	0.06881	36	0.5267	27	0.05525	8	429	430	432

NOTES: * Ratios corrected for fractionation, spike, 5-10 pg blank and initial common Pb, calculated from the model of Stacey and Kramers (1975) for the age of the sample and 1 pg U blank. 2 sigma uncertainties reported after the isotopic ratios. clr=clear, euh=euhedral, abr=abraded, lt=light, brn=brown, anhed=anhedral, med=medium, frags=fragments.

the Grenvillian and mid-Silurian orogenies. The age of 1510 Ma is interpreted to be that of igneous crystallization of the protolith granitic rock.

Relations between the various lithological units of the Corner Brook Lake Complex have not been determined, in large part because of the high-strain to which the units were subjected during Paleozoic deformation. Hence it is not known if the 1510 Ma age for the banded granitoid gneiss provides an upper or lower limit on the age of other lithologies within the complex, such as the homogeneous granitoid gneiss and quartzitic paragneiss.

All lithologies within the Corner Brook Lake Complex are considered to be Mesoproterozoic. However, the presence of Neoproterozoic or considerably younger late Precambrian felsic plutons elsewhere within the Humber Zone indicates that similar intrusions may constitute an as yet unrecognized component of the Corner Brook Lake Complex. In the Long Range Inlier on the Great Northern Peninsula, granitoid plutons intrusive into an approximately 1500 Ma gneiss complex have given a U-Pb zircon age of 1056 +4/-6 Ma (H. Baadsgaard, pers. comm., 1989, in Owen, 1991). Felsic plutons of latest Precambrian age intrude older basement at Hare Hill, immediately south of the Corner Brook-Glover Island region (Currie et al., 1992), and at the base of the Mount Musgrave Group, to the north of the map area (Hughes Lake Complex of Williams et al., 1985). The sliver of granitoid gneiss west of Sandy Point (g.r. 322921) at the junction of Grand Lake with its southwestern arm and herein mapped as part of the eastern gneiss of the Corner Brook Lake Complex was mapped as part of the Hare Hill Complex by Currie and van Berkel (1992a). This rock is strongly deformed with a well developed Paleozoic gneissic fabric and is lithologically similar to the homogeneous granitoid gneiss mapped elsewhere in the complex (e.g. Fig. 6B). In addition, the basement slice on Glover Island consists of a pegmatitic granitoid intruding banded diorite. The granitoid is relatively massive and little deformed and although restricted to the basement sequence, could be considerably younger than the diorite component of the complex.

Amphibolite within the Late Precambrian Lady Slipper Pluton and the Late Precambrian to possibly early Cambrian Mount Musgrave Group, may indicate that at least some of these bands and plutons within the Corner Brook Lake Complex are also younger intrusive phases.

The Corner Brook Lake Complex is the oldest rock unit within the Corner Brook-Glover Island region and is basement to rock units which accumulated along the eastern margin of Laurentia during the Appalachian cycle. The base of the complex is either unexposed or is faulted. The complex is overlain by the Mount Musgrave Group.

Cover sequence

The Corner Brook Lake Complex is unconformably overlain by a cover sequence consisting of a lower metaclastic succession (the Mount Musgrave Group) and an overlying carbonate dominated succession (the Breeches Pond and the Reluctant Head formations, and the undifferentiated carbonates of the Port au Port, St. George, and Table Head groups) (Fig. 5; Map 1893A, in pocket). Silicic and minor mafic igneous rocks of the Lady Slipper Pluton occur locally at the inferred base of the Mount Musgrave Group.

Lady Slipper Pluton

Definition and distribution

The name Lady Slipper Pluton is introduced for a unit of granodioritic gneiss and amphibolite that lies at the inferred base of the cover sequence. It takes its name from the Lady Slipper forestry road system along which it is best exposed (g.r. 398058 to g.r. 400058). The pluton has only been recognized at the southwest end of Corner Brook Lake where it extends for less than a kilometre along strike to the south of the lake, and has an exposed across-strike width of a few hundred metres. The pluton is differentiated from the Corner Brook Lake Complex by its leucocratic character and K-feldspar-poor composition, but it is possible that some of the homogeneous granitoid gneiss and amphibolite included within the Corner Brook Lake Complex may constitute part of the same igneous pulse (e.g., g.r 384112).

Lithology

The pluton consists predominantly of intensely lineated tonalitic to granodioritic gneiss (Fig. 8A) with varying proportions of interbanded amphibolite. In western outcrops of the body along Lady Slipper road, amphibolite is either absent or only present in thin bands (Fig. 8B), but these bands become thicker (Fig. 8C) and more prominent to the east, with the eastern margin of the unit consisting almost entirely of amphibolite (Fig. 8D). The white to cream tonalitic gneiss consists of quartz and plagioclase with minor biotite and locally magnetite. Titanite with lesser zircon and monazite are accessory phases. Thin, less than 10 cm thick, quartz-plagioclase pegmatitic sweats are locally present in the western half of the unit. The amphibolite consists of green amphibole with albite porphyroblasts (Fig. 8D), large, late biotite, minor garnet, and accessory titanite.

Age and relations

A sample of cream, homogeneous tonalitic gneiss from the central portion of the body was collected for U-Pb age determination. This rock yielded abundant needle-like zircons, and these were preferentially analyzed to avoid or minimize any inherited older component. Despite using this approach, no fraction is concordant and the four analyses of strongly abraded zircon needles (Fig. 7B, Table 1) define a mixing line (43% probability of fit) with a lower intercept age of igneous crystallization of 555 +3/-5 Ma. The upper intercept of 1400 +220/-200 Ma reflects the average Mesoproterozoic age of the inherited zircon present as cores in some grains. This upper intercept age overlaps with the age determined for Grenville basement of the Corner Brook Lake Complex and suggests that the tonalitic gneiss includes a significant component of Grenvillian basement.

The Lady Slipper Pluton is overlain to the east by psammites of the South Brook Formation of the Mount Musgrave Group. The contact is planar, parallel to the regional foliation, and although it shows minor shearing, relations are consistent with an original nonconformable contact, with the pluton acting as local basement for deposition of the siliciclastic sediments of the Mount Musgrave Group. The western contact of the pluton is not exposed. Calc-schists of the Breeches Pond Formation outcrop just to the west of the pluton, and the inferred early Paleozoic age of this unit, combined with evidence for the nonconformable eastern contact of the pluton, suggest the western contact is faulted.

Mount Musgrave Group

The Mount Musgrave Group (McKillop, 1963; Lilly, 1963; Williams et al., 1985) is divisible into Little North Pond and South Brook formations. The group takes it name from Mount Musgrave (g.r. 388199) which lies in the north of the map area just east of the Marble Mountain ski resort. The Little North Pond Formation consists of thick-bedded arkose and was considered by Williams et al. (1982, 1983, 1985) to lie at the base of the group and be stratigraphically overlain by interstratified psammitic and pelitic schists of the South Brook Formation. Although localized and discontinuous variations in lithology occur within the group in the Corner Brook-Glover Island region, no consistent stratigraphic progression in the character of the unit was noted and all rocks within the group, including those at its base, are assigned to the South Brook Formation.

South Brook Formation

Distribution and thickness

The South Brook Formation outcrops in a series of thrust slices throughout the central part of the map area and is particularly well exposed on the high barren hills extending from east of Corner Brook Lake north to the Trans Canada Highway. Metaclastic lithologies east of the Corner Brook Lake Complex rocks on Glover Island, previously mapped as the Keystone schist by Knapp (1982), are lithologically identical to the South Brook Formation and are here included within the formation. The slopes of the Marble Mountain ski resort, on the western flank of Mount Musgrave, are defined as the type section (g.r. 393217 to g.r. 388201).

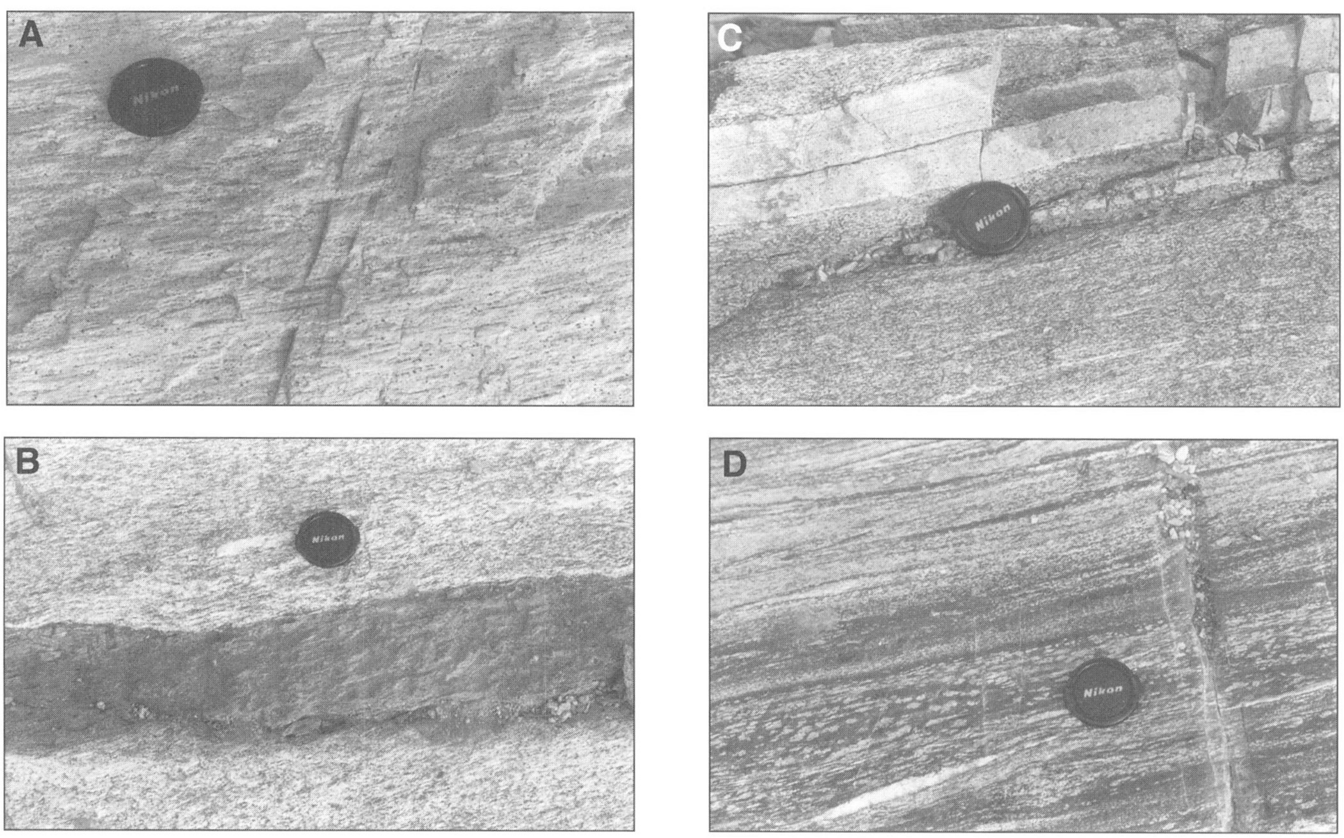


Figure 8. *Gneissic rocks of the Lady Slipper Pluton (map unit PL).* ***A)*** *Lineated, homogeneous tonalitic gneiss (g.r. 398058). GSC 1995-149C* ***B)*** *Amphibolite band in tonalitic gneiss (g.r. 399058). Felsic-mafic contact is oblique to, and overprinted by gneissic foliation. GSC 1995-149II* ***C)*** *Dioritic gneiss with interbanded tonalitic gneiss (g.r. 399058). GSC 1995-149JJ* ***D)*** *Banded amphibolite from eastern, upper contact of Lady Slipper Pluton (g.r. 400048). Banding defined by variation in proportion of amphibole to porphyroblastic albite. GSC 1995-149FF*

The thickness of the formation is highly variable between thrust slices and ranges from around 5 m or less to possibly several thousand metres. Thin slices occur within the thrust stack east of Corner Brook Lake, and consist of approximately 5 to 200 m thick sections of South Brook Formation stratigraphically between rocks of the Corner Brook Lake Complex and Breeches Pond Formation. The thrust stack east of Corner Brook Lake, termed the Valley Lakes thrust stack, is overthrust by a large coherent sheet of South Brook Formation. The absence of internal marker horizons within the overthrust block and the presence of multiple penetrative foliations and folding prevent any accurate estimate of stratigraphic thickness. The across-strike width of the thrust sheet is at least 10 km without an exposed stratigraphic top or base, suggesting a possible stratigraphic thickness of up to several thousand metres. On Glover Island the formation has a top faulted across strike width of 800 m.

Lithology

The formation is characterized by a quartz-rich metasedimentary sequence composed of interstratified quartz-muscovite schist, quartz-muscovite-albite schist, quartzite, and mica schist, along with minor quartz-pebble metaconglomerate, garnet-staurolite-kyanite-mica schist, graphitic schist, and marble (Fig. 9). Amphibolite lenses occur locally.

Siliciclastic units contain varying amounts of quartz, feldspar, and muscovite with coarser grained lithologies generally rich in quartz, and finer grained sequences rich in mica (Table 2). Albite, muscovite, garnet, staurolite, kyanite, biotite, epidote, and magnetite, and less commonly chloritoid, calcite, and hornblende as porphyroblasts, along with accessory titanite, ilmenite, apatite, rutile, monazite, and zircon are variably distributed throughout the siliciclastic sequence depending on original composition and subsequent metamorphic grade. Apart from bedding, defined by abrupt lithological changes or heavy mineral concentrations within massive arkose, primary sedimentary features are absent.

Metaconglomerate is a distinctive siliciclastic lithology which occurs at, or near, the base of the South Brook Formation (Fig. 9C). This unit is best exposed at the basement-cover interface along the southwestern arm of Grand Lake (Piasecki, 1991), but also along strike of this contact (g.r. 258940), as well as in separate thrust slices overlying basement at the southern end of Corner Brook Lake (g.r. 391057) and on Glover Island (g.r. 436015, 434003). In addition, granule to fine pebble conglomerate occurs locally throughout the metaclastic sequence. The conglomerate consists of quartz clasts with minor granitic to pegmatitic gneiss, feldspar and psammite clasts in an arkosic matrix of quartz, feldspar, muscovite, and opaque minerals. Albite porphyroblasts are present in some outcrops in both clasts and matrix. Clasts are generally of pebble size but locally reach up to

Table 2. Modal analyses of South Brook formation (data from Kennedy, 1981).

	conglomerate		psammite and pelite					albite schist			
Sample No.[1]	246	157	161	259	205	169	305	158	134	282	288
Quartz	52	57	45	43	7	20	42	29	23	22	6
K-feldspar	21	-	-	-	-	-	-	-	7	8[2]	-
Plagioclase	6	26	6	9	16	28	35	35	41	40	57
Muscovite	18	9	30	27	40	30	8	28	14	18	7
Biotite	-	3	3	7	-	6	11	-	11	-	20
Epidote	-	-	-	-	10	tr	-	1	1	2	2
Chlorite	-	3	3	3	14	2	2	1	1	7	7
Oxides	3	2	3	1	4	3	1	6	2	3	1
Tourmaline	-	-	-	-	9	-	-	-	-	-	-
Apatite	-	-	-	-	-	-	-	-	-	-	tr
# Points	1084	989	1027	1004	1038	910	1095	1059	1096	1039	957

Samples:
246 - green-grey and pink, coarse-grained, quartz-feldspar metaconglomerate (g.r. 363015).
157 - buff, fine-grained, thinly layered, quartzofeldspathic schist (g.r. 403158).
161 - grey, fine-grained, thinly layered, garnetiferous quartz-mica schist (g.r. 388199).
259 - grey, fine-grained, thinly layered, garnetiferous quartz-mica schist (g.r. 430180).
205 - green, fine-grained, muscovite-albite-tourmaline schist (g.r. 389214).
169 - grey to buff, fine-grained, garnetifeous, quartz-albite-mica schist (g.r. 502204).
305 - buff to pink, fine-grained, quartzofeldspathic schist (g.r. 421071).
158 - orange-pink, medium-grained, albite-quartz-muscovite schist (g.r. 417153).
134 - orange-pink, coarse-grained, albite-quartz-muscovite schist (g.r. 388004).
282 - pink, medium-grained, albite-quartz-muscovite schist (g.r. 406003).
288 - green, medium-grained, albite-muscovite schist (g.r. 470081).

[1] Sample numbers and descriptions taken from Kennedy (1981, Tables 5 and 7). Grid references for sample locations are approximate and are adapted from Figure 30 in Kennedy (1981). Abbreviation: tr - trace amount.
[2] mainly adularia vein

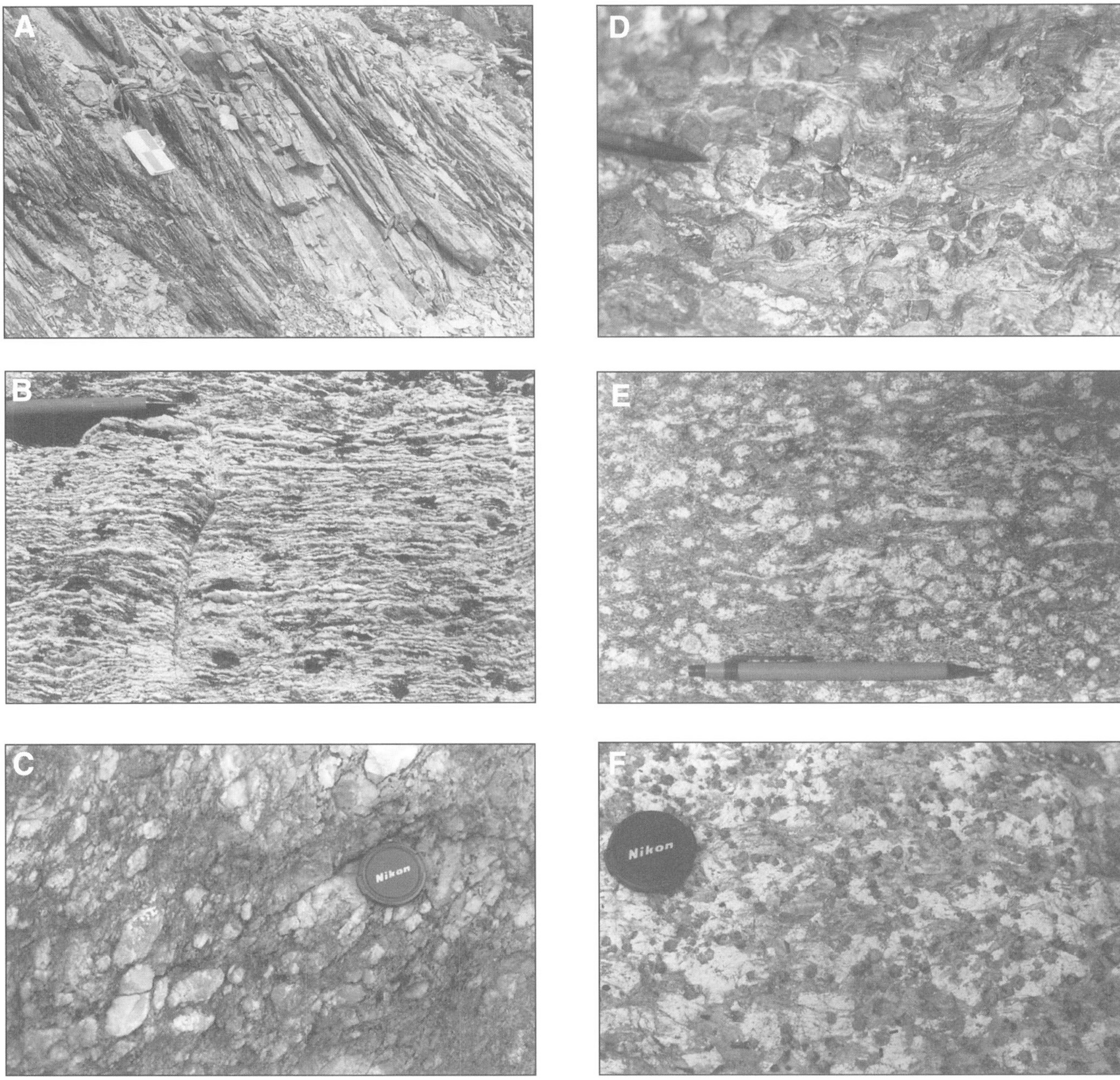


__Figure 9.__ Rocks of the South Brook Formation (map unit P€SB). __A)__ Interbedded quartzite and pelitic schist (g.r. 339981). Notebook is 18 cm long. Pelitic schist is the dark unit under the book and the quartzite is the lighter, more massive band to the right. The main foliation (S_1) is parallel to (S_0). GSC 1995-149DD __B)__ S_1 differentiated layering in quartz-muscovite schist on the powerline trail east of Breeches Pond (g.r. 390188). GSC 1995-149CC __C)__ Quartz pebble conglomerate overlying basement of the Corner Brook Lake Complex (g.r. 391057). GSC 1995-149BB __D)__ Mica schist with garnet porphyroblasts (g.r. 392106). The garnets have an internal fabric (S_1), which continues into the matrix. Kyanite, staurolite, and biotite are also present within the schist at this locality. GSC 1995-149AA __E)__ Round albite porphyroblasts in quartz-mica schist (g.r. 476108). The elongate white lenses are quartz. GSC 1995-149Z __F)__ Large albite porphyroblasts overgrowing garnet, kyanite, staurolite, and tourmaline in mica schist of South Brook Formation (map unit P€SB) on the shore of Corner Brook Lake (g.r. 403062). GSC 1995-149Y

15 cm. The thickness of the metaconglomerate ranges from 0-15 m. Immediately south of the map area on the southern shore of the west arm of Grand Lake the unit attains an across strike width of approximately 10 m (K.L. Currie, pers. comm., 1993). Thin granule to pebble conglomerate beds (<1 m thick) interstratified with psammite also occur directly overlying basement on Glover Island. Modal data for two metaconglomerate samples analyzed by Kennedy (1981) are presented in Table 2; sample 246 is from a conglomerate within a fault sliver of South Brook Formation whereas sample 157 is from a unit of granule to pebble conglomerate interstratified within the main psammitic sequence.

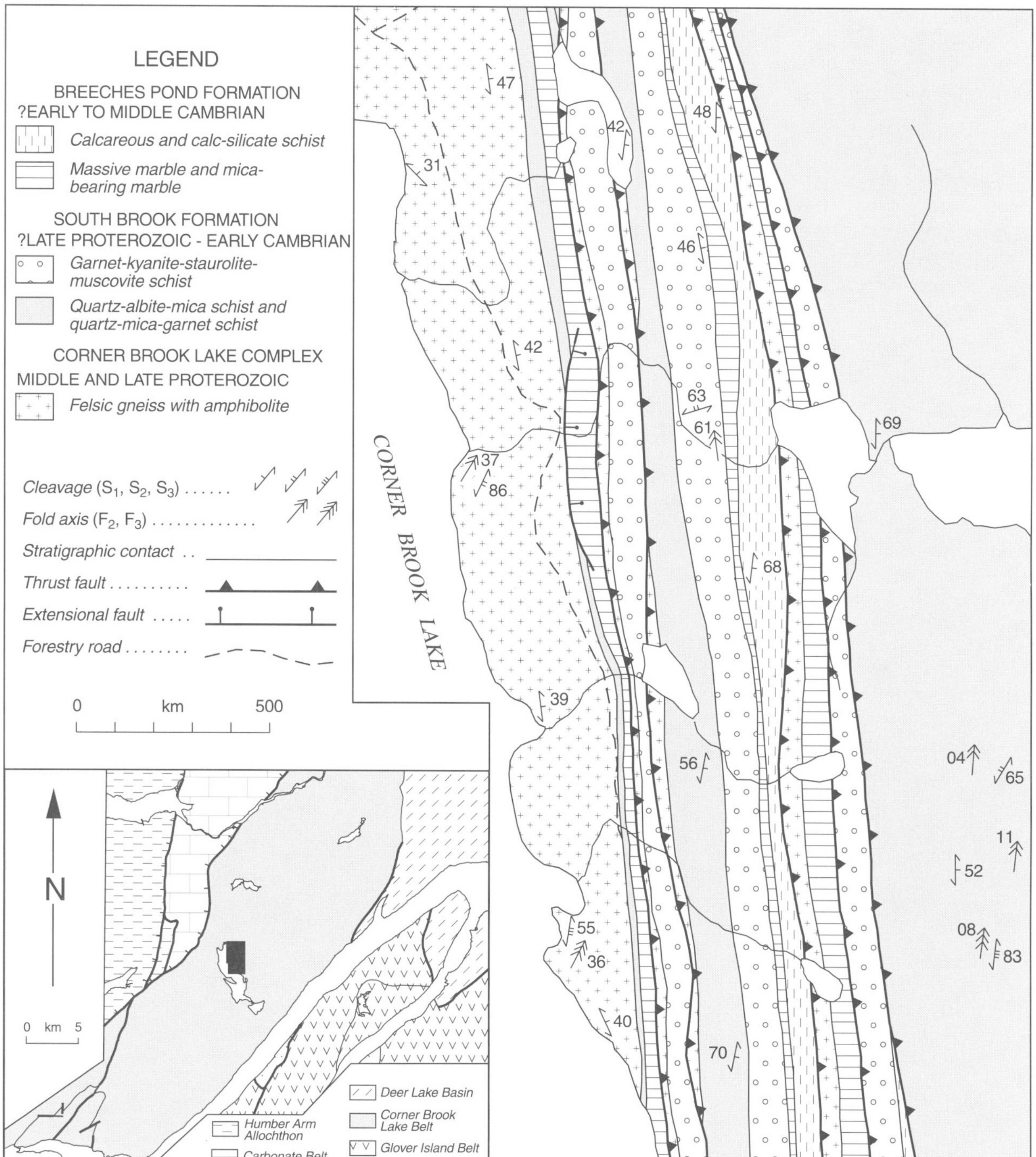


Figure 10. Geological map of the Valley Lakes thrust stack northeast of Corner Brook Lake showing distribution of basement and cover rock units within the imbricate thrust stack, based on mapping by the authors (1991) and Ryan (1992). The inset shows the location of the map within the Corner Brook-Glover Island region.

A pelitic schist with abundant garnet, staurolite, and kyanite porphyroblasts is well developed within the South Brook Formation east of Corner Brook Lake (Fig. 9D) and was also reported by Piasecki (1991) along the shores of the southwestern arm of Grand Lake. The matrix consists of muscovite with plagioclase±quartz±biotite. Epidote, tourmaline, chlorite, and iron oxides are accessory phases. Graphitic beds up to 5 cm thick occur locally. The unit is up to 40 m thick, and in the Valley Lakes thrust stack east of Corner Brook Lake it forms a distinctive marker horizon that can be traced for up to 5 km along strike (Fig. 10).

Amphibolite occurs in bands, parallel to bedding, generally up to a few metres thick, but one band in Keystone Brook on Glover Island is about 70 m thick (Knapp, 1982). It consists of plagioclase and hornblende, plus epidote, chlorite, titanite, and porphyroblastic biotite.

Marble and calc-schist occur in beds up to 2 m thick interstratified within the siliciclastic lithologies. The marble, which is often micaceous and rich in pyrite, is best observed in the Glover Island sections of the formation (Cawood and van Gool, 1992; Knapp, 1982) and along the southwestern arm of Grand Lake (Piasecki, 1991). Thin-bedded, grey, calcareous schist and limestone occur in the large overthrust block of South Brook Formation along the ridge immediately south of the Trans Canada Highway (g.r. 456253) and just west of the Northern Harbour road (g.r. 550219) where the calc-schist is fuchsite-bearing.

The amphibolite is restricted to lower parts of the formation. On Glover Island the marble, along with the biotite and graphitic schist, occurs in the upper segments of the unit, and the spatial variability between these lithologies and the distribution of amphibolite formed the basis for Knapp's (1982) informal division of the formation into lower and upper members. Along the southwestern arm of Grand Lake marble occurs within a few metres of the basement-cover contact (Piasecki, 1991). Amphibolite is absent from these sections of the South Brook Formation.

Albite porphyroblastic psammitic schist is characteristic of the eastern outcrops of the South Brook Formation (Fig. 9E; see map, unit P€SBa) but is also locally developed within the main part of the unit (unit P€SB). The distribution of this schist corresponds in large part with the Caribou Lake formation of Kennedy (1981). Porphyroblasts are generally less than 1 cm in diameter but range up to 5 cm across. South of Corner Brook Lake this unit is restricted to a 1.5 to 2 km wide belt, but north of the lake, folding in the Yellow Marsh antiform increases the width of the unit to approximately 10 km. Albite porphyroblasts are pervasive throughout this region and are present in all lithologies from quartzite to pelitic schist (Fig. 9F), although overall the formation consists largely of massive psammitic schist with only minor pelitic schist. Pervasive porphyroblast growth makes recognition of protolith lithologies difficult and the lithological sequences within the core of the Yellow Marsh antiform may include orthogneiss (?basement) as well as paragneiss.

Relations

The South Brook Formation unconformably overlies both basement of the Corner Brook Lake Complex and tonalite of the Lady Slipper Pluton, and is conformably overlain by carbonate and calc-schist of the Breeches Pond Formation. To the north in the Pasadena map area (12H/4) the formation is considered to conformably overlie the Little North Pond Formation which in turn overlies mafic and silicic igneous rocks of the Hughes Lake Complex which contains the Round Pond granite, dated by U-Pb zircon at 602 ± 10 Ma (Williams et al., 1985). In the map area basal stratigraphic contacts are exposed around Corner Brook Lake (g.r. 400058 and 392098), along the west coast of Glover Island (g.r. 436015), along Keystone Brook (g.r. 434004), and along the southern shore of the southwestern arm of Grand Lake (Fig. 11; g.r. 264911; Piasecki, 1991). At all localities, psammite and quartz granule to quartz pebble conglomerate stratigraphically overlie granitoid gneiss. These lithologies along with the metaconglomerate recognized elsewhere at, or near, the basal contact are similar to sequences described for the Little North Pond Formation in the Pasadena map area. However, in the Corner Brook-Glover Island region, these lithologies are restricted aerially and represent facies variations within the South Brook Formation, rather than a distinct mappable lithostratigraphic unit.

The upper stratigraphic contact of the South Brook Formation with the Breeches Pond Formation is marked by massive and bedded carbonate, limestone conglomerate, and calc-schist. The contact is exposed in the thrust slices along the eastern side of Corner Brook Lake (Fig. 10). Where the

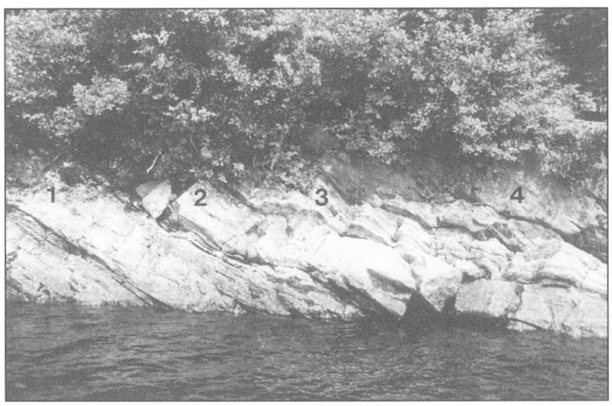


Figure 11. *Unconformity between Corner Brook Lake Complex and South Brook Formation on the southwestern arm of Grand Lake (g.r. 264911). Basement (**1**) consists of a highly foliated homogenous quartzofeldspathic gneiss and is restricted to a 2 m band shown in the left side of the photograph. The cover rocks consist, from left to right, of about 1 m of massive quartz-magnetite schist (**2**) containing albite porphyroclasts, 1 m of banded, coarse grained pyritiferous marble (**3**), and a thick sequence of garnetiferous garbenschiefer (**4**) of the Breeches Pond Formation. GSC 1995-149X*

upper contact is faulted it is generally against other lithostratigraphic units of the Humber Zone. On Glover Island, however, the South Brook Formation is overthrust by ultramafic and mafic plutonic rocks of the Grand Lake Complex of the Dunnage Zone. The juxtaposition of Matthews Brook Serpentinite against South Brook Formation (g.r. 483264) and Breeches Pond Formation (g.r. 308990) may be a similar but more disrupted example than that preserved on Glover Island, of Dunnage Zone overthrusting Humber Zone.

Age

No age data are available for the South Brook Formation, but ages on enclosing units and regional relations suggest it is of latest Precambrian to early Cambrian age. The 555 Ma age of the Lady Slipper Pluton and the inferred nonconformable relationship between this unit and the South Brook Formation imply that at least parts of the formation are latest Precambrian or younger. The overlying Breeches Pond Formation is correlated with the Reluctant Head Formation (Fig. 5) which contains trilobites as old as late Middle Cambrian (Knight and Boyce, 1991).

Breeches Pond Formation

Definition, distribution, relations, and thickness

The term Breeches Pond Formation was introduced by Cawood and van Gool (1992) for all carbonate, pelite and psammite unconformable overlying basement of the Corner Brook Lake Complex. It included a lower siliciclastic member and an upper carbonate dominated member. Subsequent mapping has shown that the lower siliciclastic sequence constitutes part of the South Brook Formation. The Breeches Pond Formation is therefore redefined to include only the carbonate sequence, stratigraphically overlying the South Brook Formation.

The formation outcrops in a series of fault blocks and thrust sheets defining an overall northeast-tapering wedge extending from the southwestern arm of Grand Lake to just southeast of Humber Gorge.

The unnamed stream draining into Corner Brook Lake from Valley Lakes (g.r. 389105 to g.r. 391105) is designated as the type section, and the unit is well exposed along strike for a kilometre to both the north and south of this section. The major east-trending stream at the northern end of the Little Grand Lake map sheet (Grand Lake Brook; between g.r. 315989 to g.r. 328989) provides an additional representative section and exposes a pelitic schist and calc-schist dominated segment of the formation, in contrast to the marble and limestone conglomerate dominated character in the type section. The lower contact of the formation at the type section is marked by a late brittle normal fault which cuts out the thin intervening unit of South Brook Formation and directly juxtaposes the Breeches Pond Formation against the Corner Brook Lake Complex. Stratigraphic continuity with the South Brook Formation is observed along strike to the north. The contact with the South Brook Formation is mapped at the incoming of calc-silicate schist and marble. Where the Breeches Pond Formation appears to directly overlie basement of the Corner Brook Lake Complex, a thin sliver of South Brook Formation, too thin to show on the map and sometimes no thicker than 5 m, is always present (see map and Fig. 10). The upper contact of the Breeches Pond Formation is generally faulted either by early thrust faults, resulting in repetition of the basement-cover sequence such as in the Valley Lake thrust stack (Fig. 10), or by later high-angle faults, such as in the southern end of the map area. In the southwest corner of the area, immediately north of the southwestern arm of Grand Lake, the formation passes up conformably into massive carbonate of the main cover sequence (?Port au Port Group).

The formation has a maximum thickness in the type section of approximately 100 m. In the thrust slices and fault blocks west and south of Corner Brook Lake, the unit has an across-strike width of up to several kilometres. However, complex folding and faulting and the lack of internal marker horizons prevent any accurate estimation of thickness within this region.

Lithology

The Breeches Pond Formation consists of marble, limestone conglomerate, calc-schist, and pelitic schist. Marble and pelitic schist represent end members of a compositional spectrum which includes impure marble and calc-schist, depending on variation in the carbonate to silicate ratio. The marble is buff to cream, bedded to massive limestone (Fig. 12A) commonly with a sugary recrystallized texture. Detrital quartz grains and metamorphic muscovite occur locally, and where muscovite is widespread this results in a well developed penetrative fabric. Pelitic schists are grey and locally graphitic. The buff to grey calcareous schist and calc-schist consist of varying proportions of calcite-dolomite-muscovite-plagioclase-quartz± biotite±hornblende±garnet±tremolite±zoisite±epidote± chlorite±talc±diopside. Amphibole within garbenschiefer occurs in spectacular bow-tie structures (Fig. 12B).

Limestone (marble) conglomerate is a distinctive lithology of the Breeches Pond Formation and occurs in beds from 30 cm to at least 3 m thick. Clast size ranges from pebble to boulders up to 30 cm diameter, with the majority of clasts in the 4 to 10 cm range (Fig. 12C). Clasts within the conglomerate include equidimensional coarsely recrystallized marble and finer grained more platy limestone, the latter probably of intraformational origin. The matrix is calcareous schist.

Coarse-grained siliciclastic units are generally absent from the Breeches Pond Formation, although at the southeast end of Corner Brook Lake (g.r. 363017) near the structural top of a wide block of the formation, a 5 m thick band of quartzite and quartz-bearing psammite is interstratified with carbonate and pelitic schist.

Age

No fossils have been found in the formation. Pervasive deformation and extensive recrystallization of the limestone and shale protoliths of the marble and pelite probably destroyed any fossiliferous material within the rocks. The unit is lithologically similar to, and probably correlative with, the Reluctant Head Formation of late Middle Cambrian to Late Cambrian age (Fig. 5; Knight and Boyce, 1991).

Reluctant Head Formation

Definition, distribution, and thickness

The Reluctant Head Formation is the basal unit of the Carbonate Belt exposed within the Corner Brook-Glover Island region (Fig. 4, 5). It consists of bedded carbonate, slate, and limestone conglomerate. The term was originally introduced by Lilly (1963) for rock units exposed in the Pasadena map sheet around Old Mans Pond. The formation can be traced from the Old Mans Pond region south into the Corner Brook map area (Williams et al., 1983), where it is best exposed to the north of Humber Gorge around Wild Cove Lake and Rubber Lake (see map). South of the gorge, the formation wedges out against the Humber River fault, which separates the Carbonate Belt from the Corner Brook Lake Belt.

Multiple pervasive deformational events prevent any meaningful estimation of thickness of the formation. In the Old Mans Pond region where the formation is less disrupted, Gillespie (1983) estimated a thickness of 250 m.

Lithology

The Reluctant Head Formation consists of thin-bedded, ribbon limestone with interstratified slate and limestone conglomerate. Thick-bedded limestone and dolostone forming regionally mappable ridges (unit ЄRc) occur around Wild Cove Lake and Rubber Lake.

The ribbon limestone consists of grey lime mudstone intercalated with yellow-weathering dolomitic mudstone or grey slate (Fig. 13A, B). Carbonate beds are 1-5 cm thick with slate interbeds ranging from 1 mm to 1 cm thick. Limestone conglomerates up to several metres thick consist of pebble to cobble and locally boulder size clasts in a grey lime mudstone or yellow weathering dolomitic mudstone matrix. The majority of the limestone clasts are platy and probably of intraformational origin, but equidimensional grainstone clasts up to cobble size are a minor component of some thicker beds.

Age and relations

Knight and Boyce (1991) reported a trilobite fauna of late Middle Cambrian to early Late Cambrian age from the upper part of the Reluctant Head Formation exposed on the Hughes Brook-Goose Arm logging road system south of Old Mans Pond. The formation is base faulted within the Corner Brook map area but is stratigraphically overlain by thick-bedded carbonates of the Berry Head Formation of the upper Port au Port Group (Knight and Boyce, 1991). Conformable upper

Figure 12. *Rocks of the Breeches Pond Formation (map unit ЄOB). **A**) Mylonitic banded marble exposed on the shore of Corner Brook Lake (g.r. 393086). Shear bands offsetting the banding in the top left-hand part of the photograph indicate top up to the northwest movement on the shear zone. View looking north. GSC 1995-149NN **B**) Garbenschiefer with well developed bow-tie aggregates of amphibole and small garnet porphyroblasts (g.r. 351000). GSC 1995-149EE **C**) Limestone conglomerate (g.r. 379078). GSC 1995-149W*

contacts of the formation were observed on the Trans Canada Highway in Humber Gorge (g.r. 383220) and at the head of Wild Cove Lake (g.r. 383249).

Port au Port, St. George, and Table Head groups

Carbonates of the undifferentiated Port au Port, St. George, and Table Head groups form a prominent belt within the western half of the Corner Brook-Glover Island region. They extend some 15 km from the head of Humber Arm south to Meadows Pond where they wedge out against the structural boundary with the internal domain of the Humber Zone, and then reappear south of Pinchgut Lake and extend southeast to the head of the southwestern arm of Grand Lake at the southwest corner of the map area.

The carbonate sequence consists of massive to thick-bedded and nodular limestone and dolostone (Fig. 13C). Disruption of the carbonate sequence, particularly in the east, and recrystallization of the carbonates to marble under greenschist facies conditions hindered differentiation of the sequence into its constituent groups and formations apart from a few localized well exposed sections. Continuous

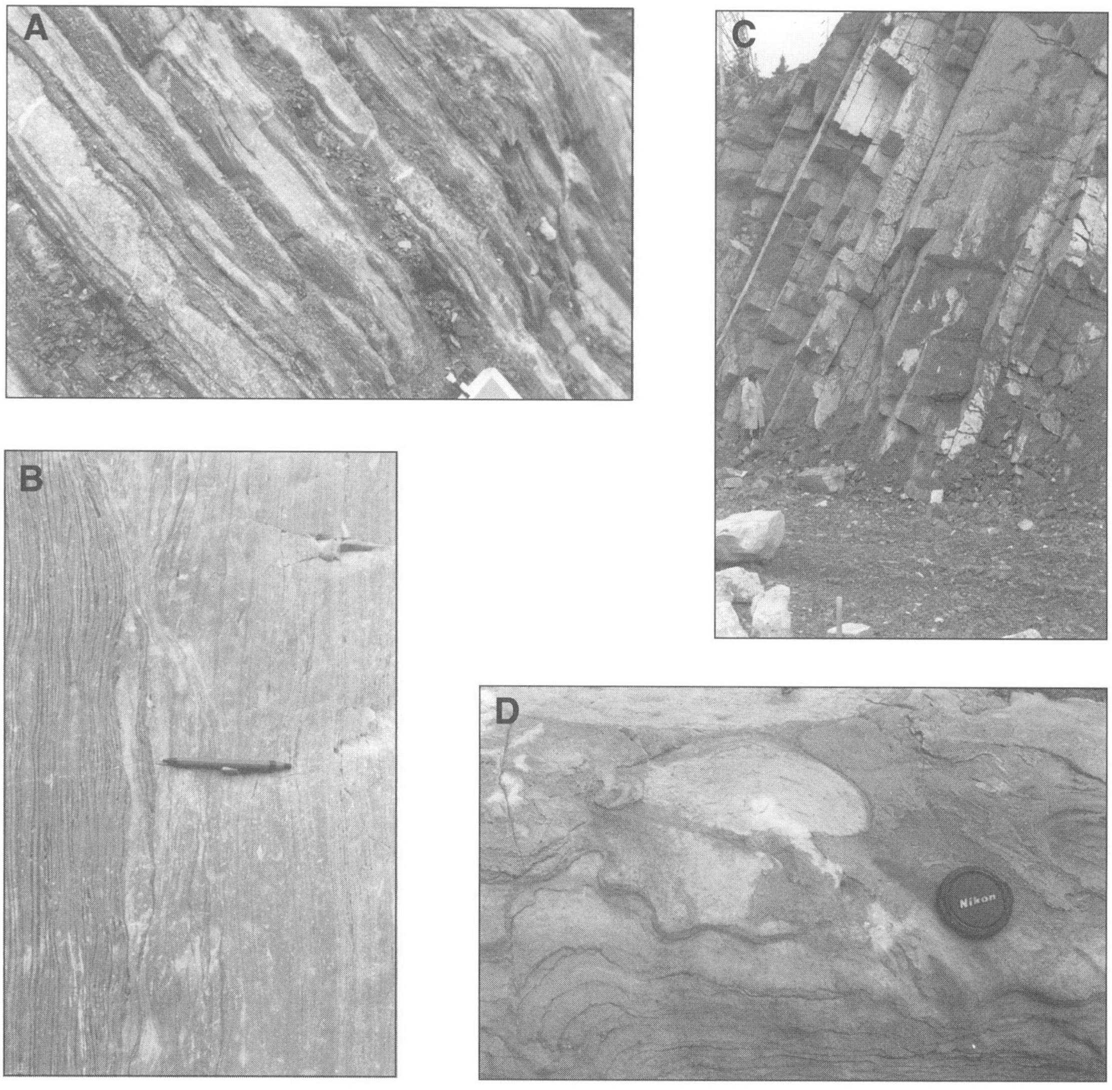


***Figure 13**. Rocks of the Carbonate Belt. **A**) Thin-bedded limestone and slate of the Reluctant Head Formation (map unit ЄR; g.r. 383220). The foliation (S_1) is parallel to bedding. GSC 1995-149V **B**) Thin-bedded limestone of the Reluctant Head Formation (map unit Єr; g.r. 383220). In the left one-third of the photograph, the bedding is accentuated by thin slate partings. S_1 is parallel to S_0. GSC 1995-149T **C**) Thick-bedded limestone of the (?)Table Point Formation at the stratigraphic top of the undifferentiated sequence of Port au Port, St. George, and Table Head groups (map unit ЄOC), adjacent to the contact with the Humber Arm Allochthon (g.r. 331207). GSC 1995-149O **D**) Stromatolite mound within dolostone of the Aguathuna Formation at the top of the St. George Group, adjacent to the contact with the Meadows Pond formation of the Table Head Group (g.r. 333149). GSC 1995-149N*

exposure through the carbonate sequence, at the head of Humber Arm on the north shore of Wild Cove, at new road cuts along the Trans Canada Highway in Humber Gorge, in the vicinity of Meadows Pond, and along the Lady Slipper road system east of Lady Slipper Pond, enabled recognition of the Catoche and Aguathuna formations of the St. George Group, and the Table Point Formation of the Table Head Group (cf. Knight and James, 1987; Stenzel et al., 1990). Lithological and facies types in the least deformed and metamorphosed western outcrops of the carbonate sequence are generally similar to those described from the well known, low-grade exposures through the carbonate cover sequence on the Great Northern Peninsula and the Port au Port Peninsula (e.g. Cumming, 1983; Knight and James, 1987). Stromatolite and thrombolite mounds are locally preserved in the upper St. George Group (g.r. 353242 and g.r. 333149; Fig. 13D).

Near Meadows Pond, dolostones of Aguathuna Formation of the upper St. George Group are stratigraphically overlain by thin-bedded, fine-grained grey limestone with slaty lime mudstone. This unit is herein informally referred to as the Meadows Pond formation of the Table Head Group, and is distinct from the thick-bedded limestones of the Table Point Formation, which elsewhere occupy this stratigraphic level within the carbonate sequence. I. Knight (pers comm., 1993) has recognized a similar lithology within the Table Head Group along strike to the north in the Pasadena map area (12H/4). Immediately north of Meadows Pond (g.r. 332147) the unit contains a fauna of gastropods, brachiopods, crinoids, trilobites, sponges, cephalopods, ostracod, and bivalves. Fossils determined by W.D. Boyce (pers. comm., 1994) include brachiopods of the genera *Orthambonites* (sp. cf. *O. marshalli*, Wilson, 1926) and *?Pleurortis* and gastropods of the genera *Helicotoma* (sp. undet) and *Maclurites* (*Maclurites oceanus*, Billings, 1865) which indicate a range from the Catoche Formation to the Table Point Formation. The age of this fauna is of the *Orthidiella* zone (W.D. Boyce, pers. comm. 1994). The thin, continuous bedded nature of the unit suggests it is a deep-water, offshore facies of the Table Point Formation.

Humber Arm Allochthon

Rock units temporally equivalent to the cover sequence in the Carbonate and Corner Brook Lake belts are also preserved in thrust slices of the Humber Arm Allochthon (Fig. 5). These transported rocks were assigned to the Humber Arm Supergroup by Stevens (1970).

Summerside Formation

Definition and distribution

The Summerside Formation (Weitz, 1953; Stevens, 1965; Brückner, 1966) is the lowest stratigraphic unit within the lower structural slices of the Humber Arm Allochthon. The formation is well exposed in roadside outcrops on the north side of Humber Arm around the village of Summerside and on the south side of the arm around Crow Hill (see map, in pocket; Williams and Cawood, 1986, 1989). In addition, a thin sliver of probable Summerside Formation outcrops along Watsons Brook on the eastern side of Corner Brook. This sequence was originally mapped by Williams and Cawood (1986) as the Watsons Brook member of the overlying Irishtown Formation, but the present mapping program has shown that it occupies the core of an antiform and is stratigraphically overlain by Irishtown Formation, which combined with its lithological character, suggests that it should be included within the Summerside Formation.

Lithology

The formation consists of greywacke, quartzite, quartz pebble conglomerate, and grey, green, and purple slate (Williams and Cawood, 1989). Finer grained lithologies become more dominant towards the top of the unit (Stevens, 1965). The coarser grained, stratigraphically lower units of the formation are best exposed north of Humber Arm, near the community of Summerside, and at Crow Hill. They comprise cream to pale yellow-green weathering, medium-bedded arkose and quartz pebble conglomerate with interbedded red and green sandstone and siltstone. In the upper, western sections of the formation, the coarser grained lithologies die out and the sandstone and siltstone are interbedded with grey slate. Although variations in colour of the finer grained siliciclastic lithologies are principally bedding plane controlled, local along strike colour variations indicate a diagenetic origin. The unit of Summerside Formation outcropping along Watsons Brook consists of red, green, and grey slate, similar to the upper parts of the formation west of the Crow Hill thrust, but is differentiated from these western outcrops by the localized presence of minor amounts of thin-bedded grey carbonate and calc-slate.

Thickness, relations, and age

Deformation in the Summerside Formation, combined with the lack of marker horizons, has prevented an accurate estimate of its thickness. Stevens (1965) suggested a thickness of at least 300 feet (approx. 100 m) whereas Brückner (1966) estimated a thickness of 800 feet (approx. 250 m), but this may have included part of the structurally underlying Irishtown Formation (Botsford, 1988). A thickness of approximately 150 m was proposed by Botsford (1988).

The base of the formation is invariably faulted and thrust eastward along the Crow Hill Thrust over the Irishtown Formation (see map and cross-section B). Correlation of the formation with the lower Labrador Group and the Mount Musgrave Group of the Fleur de Lys Supergroup (Fig. 5) suggests it probably originally accumulated on Grenvillian basement. Along its western margin, the formation is conformably overlain by the Irishtown Formation with the contact defined on the last outcrop of red slate (Williams and Cawood, 1986).

The Summerside Formation is unfossiliferous but is assumed to be Early Cambrian and older because of its stratigraphic contact with the overlying Lower Cambrian Irishtown Formation, and on the basis of correlations with the lower Labrador Group of the cover sequence (Cawood et al., 1988, Fig. 10) and the Lower Cambrian Blow-Me-Down

Brook Formation which outcrops in structural slices of the Humber Arm Allochthon (Lindholm and Casey, 1988, 1989; Williams and Cawood, 1989).

Irishtown Formation

Definition and distribution

The Irishtown Formation, originally mapped by Stevens (1965) as Meadows formation was first termed Irishtown Formation by Brückner (1966). The formation appears in two major thrust sheets, separated by the Crow Hill thrust, within the northwestern corner of the Corner Brook map area. The eastern belt is fault bounded between the carbonate sequence to the east and the Summerside Formation to the west. The western belt lies west of the Crow Hill thrust. In addition, a small block of the formation lying at the base of an inferred thrust sheet occurs east of Pinchgut Lake on the Trans Canada Highway.

Lithology and thickness

The formation consists of dark grey slate (Fig. 14A) with prominent white- to buff-weathering quartzite units (Fig. 14B), and local pebble to boulder conglomerate. The slate is thin bedded to laminated with local small-scale cross lamination. Most is grey but a grey-green and a bluish-weathering grey unit occurs at the base of the formation. Silvery grey and green slate units occur locally. Quartzite is

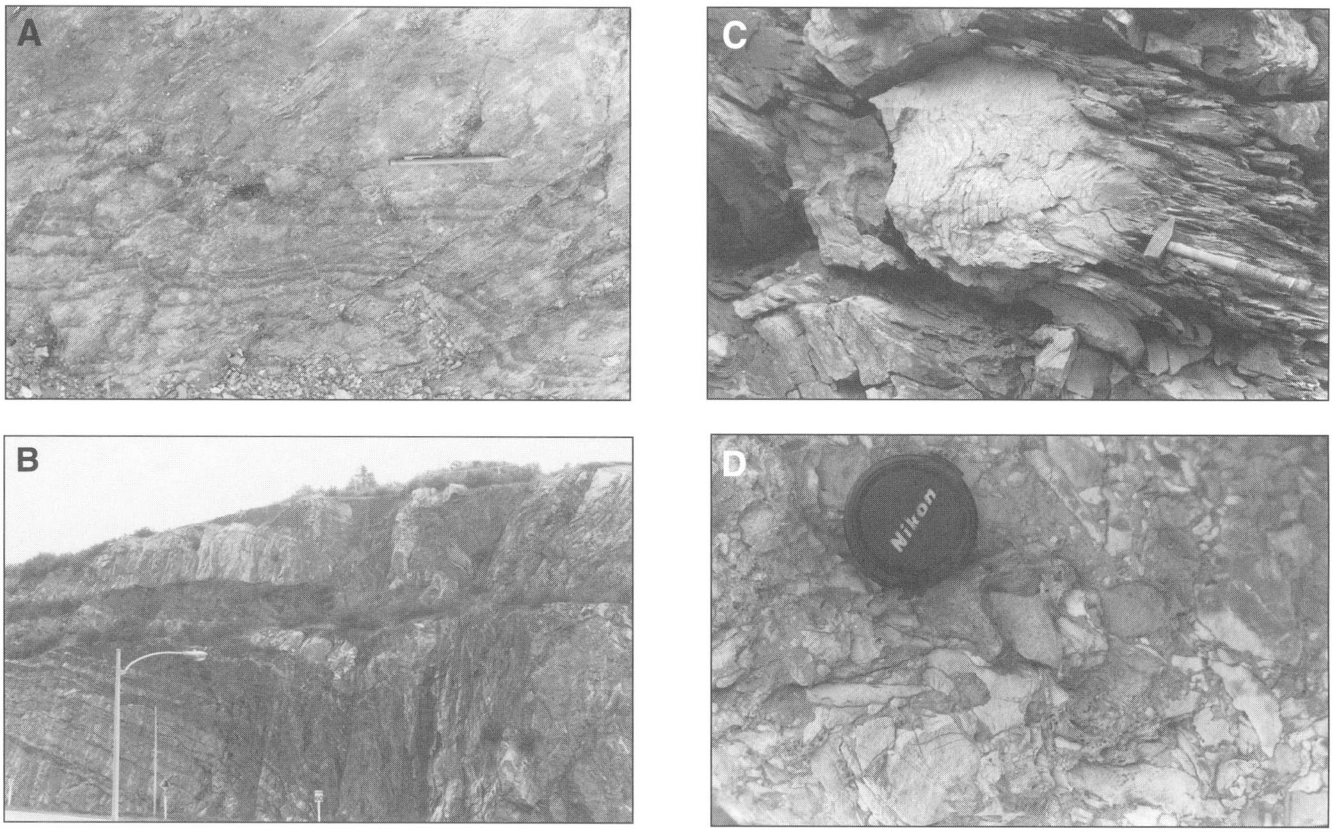


Figure 14. *Rocks of the Humber Arm Allochthon.* ***A)*** *Thin-bedded grey slate of the Irishtown Formation (map unit ꞒI) exposed by a communication tower (g.r. 288219). Prominent S_2 slaty cleavage traverses photograph from lower left to upper right, at a high angle to the bedding. GSC 1995-149M* ***B)*** *Thick-bedded, buff weathering quartzite beds with interbedded grey slate of the Irishtown Formation (map unit ꞒI), exposed in the core of the Seal Head syncline in a road cut along the Lewin Parkway, Corner Brook (g.r. 309227). View looking southeast. Note the well developed axial planar cleavage (S_2) in the hinge. The eastern normal limb is shallower than the western forelimb, in accordance with the east vergence of this F_2 structure. GSC 1995-149L* ***C)*** *Thin-bedded dolostone and slate of the Pinchgut Formation (map unit ꞒOP) on O'Connell Drive (g.r. 291212). The steeply dipping sequence is deformed by F_2 folds with shallow dipping axial planar cleavage (S_2) which is well developed in the slate, but absent in the dolostone. These folds are in the overturned limb of an F_2 structure in the footwall of the Crow Hill thrust. View looking south. GSC 1995-149P* ***D)*** *Limestone conglomerate of the Pinchgut Formation (map unit ꞒOP) exposed on the forestry track to Big Feeder Pond (g.r. 274156). The conglomerate contains platy clasts of lime mudstone of probable intraformational origin as well as more equidimensional clasts of grainstone of probable extra-basinal origin. GSC 1995-149K*

well sorted and thick bedded to massive. Conglomerate units are well exposed in a large cutting at the old railway yard in Corner Brook (g.r. 312234). These are coarse and polymictic, with pebbles to large boulders of intraformational slate and quartzite, limestone, and medium- to coarse-grained massive to sheared, and pyrophylitized granite. Quartz grains, up to 5 cm in diameter, occur in the matrix. Granite clasts were probably derived from Grenville basement, whereas sedimentary clasts represent facies equivalents of the Labrador Group. Minor, thin, grey calcareous slate and limestone beds are interstratified with quartzite and grey slate in the vicinity of the K-Mart plaza at Corner Brook.

Botsford (1988) informally divided the formation into three intervals with a total thickness of around 550 m: a 30 m thick lower slate dominated unit with thin interbeds of quartzose siltstone to fine sandstone; a 450 m thick intermediate interval dominated by thick-bedded quartzite with slate interbands; and an approximately 75 m thick uppermost interval of thin-bedded, orange-weathering quartzose siltstone with scattered thin beds of quartzite and slate, passing up into slate at the top of the interval. The conglomerate horizons occur in the middle unit.

Age and relations

Carbonate clasts within the conglomerate contain Early Cambrian trilobites, salterellids, and archaeocyathans (Walthier, 1949; McKillop, 1963; Stevens, 1965; James and Stevens, 1982). These are inferred to approximate the age of deposition of the unit as the overlying Cooks Brook Formation contains in situ late Early to Middle Cambrian fossils (Botsford, 1988). The Irishtown Formation is correlated with the Forteau and Hawke Bay formations of the Labrador Group (e.g. Cawood et al., 1988, Fig. 10).

The Irishtown Formation is conformably overlain by a thin-bedded carbonate dominated sequence, which, east of the Crow Hill thrust, is termed the Pinchgut Formation (Fig. 5) and to the west of the thrust, beyond the map area, is the Cooks Brook Formation (Brückner, 1966). The contact is marked by thin-bedded light grey limestone above a sequence of dark grey slate (Williams and Cawood, 1986).

Pinchgut Formation

Definition, distribution, and thickness

The term Pinchgut group was introduced by Williams and Cawood (1986) for a distinctive sequence of thick-bedded to massive limestone conglomerate, thin-bedded platy limestone, dolostone, and slate, outcropping south of Corner Brook. Williams and Cawood (1986) noted that the group was spatially associated with, but faulted against, a sequence of green-grey sandstone and slate which they termed the Whale Back formation. Exposures at new outcrops in the town of Corner Brook and along forestry tracks running west from the Trans Canada Highway to Big Feeder Pond, reveal that contacts between the Pinchgut and Whale Back lithologies are stratigraphic, and that the two are interstratified. In addition, the overall uniform lithological character of the sequence, with the absence of any mappable internal divisions, favours formation rather than group status for the unit. We combine and redefine these units as the Pinchgut Formation.

The Pinchgut Formation extends from Corner Brook south-southeast to the edge of the map area near Pinchgut Lake. It is best exposed in road cuts along the Trans Canada Highway (between g.r. 312129 and g.r. 308118), which is designated as the type section, and along the forestry track (between g.r. 269157 to g.r. 288158 and g.r. 307162 to g.r. 323161). The unit is differentiated from the Cooks Brook Formation, which occupies a similar stratigraphic position above the Irishtown to the west of the map area, by the higher proportion of grey slate and the presence of interstratified green sandstone and siltstone.

The formation has an across strike width of at least several kilometres, but the lack of internal marker horizons, widespread mesoscopic folding, and penetrative deformation prevent any estimation of its stratigraphic thickness.

Lithology

The Pinchgut Formation consists of grey slate, thin-bedded grey lime mudstone and buff-weathering dolostone, limestone conglomerate and bedded, fine green sandstone and siltstone. Grey to green-grey, and locally purple slate is the principal lithology, but its friable and recessive nature mean that it is under-represented in all but the freshest outcrops. The limestone and dolostone beds (Fig. 14C) are generally less than 10 cm thick and are locally contorted by soft sediment slumping. The carbonate beds range from limestone to impure calcareous slate.

Limestone conglomerate (Fig. 14D) occurs in beds less than a metre to more than 10 m thick. They contain platy or rounded limestone clasts ranging from pebble to cobble and 2 m boulder size. Most appear to be intraformational with oversize boulder clasts consisting of contorted and partially disaggregating thin-bedded limestone and slate. Equidimensional pelletoidal and oolitic grainstone indicate a minor extra-basinal clast component. The matrix is dolomitic lime mudstone or slate. A mappable sequence of bedded grey grainstone, limestone conglomerate, and thin-bedded limestone (map unit ЄOPc) occurs within the formation along the ridge west of South Bells Brook on the southern edge of Corner Brook.

Limy to non-calcareous green and grey-green lithic sandstones constitute a minor but distinctive lithological component exposed in a road side quarry (g.r. 286090), by Southwest Pond and the Serpentine Lake road (g.r. 270114 to g.r. 287130), and around Corner Brook (g.r. 290212 and g.r. 284197). The sandstones are quartz-rich but also contain detrital feldspar, quartzose siltstone fragments, and minor epidote and opaque minerals.

Age and relations

No macrofossils were observed in the Pinchgut Formation and no microfossils have been recovered from processed samples of the carbonate. The formation is lithologically similar to the Reluctant Head and Breeches Pond formations of the cover sequence and the Cooks Brook Formation of the allochthon. The late Cambrian age of these other units, extending into the early Ordovician for the Cooks Brook Formation, suggests a similar age for the Pinchgut Formation. Williams and Cawood (1986; see also Dean and Meyer, 1985, and Williams and Cawood, 1989) suggested that the sandstone was similar to the Goose Tickle Group (Cooper, 1937; Stenzel et al., 1990). Such a correlation implies the formation may be as young as Middle Ordovician (Fig. 5). The Pinchgut Formation has a conformable lower contact with the Irishtown Formation, but the upper contact is invariably faulted.

GEOLOGY OF THE DUNNAGE ZONE

Introduction

Rock units of the Notre Dame subzone of the Dunnage Zone occur along, and east of, the Cabot Fault system, principally on Glover Island and east of Grand Lake, but also in fault bound blocks along the west shore of Grand Lake, north of Northern Harbour, and along Matthews Brook (see map). Stratigraphic units of the subzone within the map area are the Grand Lake Complex, Glover Formation, and Matthews Brook Serpentinite. All units have been subjected to regional greenschist facies metamorphism. In addition, gabbros of the Grand Lake Complex show evidence for an early, possibly sea-floor, high temperature amphibolite facies metamorphism (Knapp, 1982).

Grand Lake Complex

Definition, relations, distribution, and thickness

The Grand Lake Complex of mafic and ultramafic plutonic rocks outcrops along the western and central parts of Glover Island (see map). The term was informally introduced by Knapp (1982), who recognized that the western (lower) contact of the complex with the Humber Zone lithologies of the Mount Musgrave Group was faulted. Knapp (1982) considered the eastern (upper) contact of the complex to be nonconformable with his Glover group, but our mapping has revealed a shear zone along this boundary, which we refer to as the Kettle Pond shear zone.

The Grand Lake Complex (Knapp, 1982) is well exposed on the west shore of the island, on high hills south of Bluff Head, and along Keystone Brook. No single area provides a complete section through the complex but Keystone Brook (from g.r. 442005 to g.r. 456010) has good exposures through the lower parts of the complex and is designated as the type section. The western shore of Glover Island (from g.r. 471056 to g.r. 484068) exposes the upper parts of the complex and its upper contact, and is nominated as a reference section. The complex has an across strike width of up to 2.5 km, which combined with the steep dip of igneous layering within the unit (see map and cross-sections D and F) suggests a thickness of at least 2 km.

Lithology

The Grand Lake Complex consists predominantly of gabbro, but also includes ultramafic rocks, greenschist, trondhjemite, and mafic dykes. The gabbro is divisible into a lower, layered cumulate sequence defined by alternating melanocratic and leucocratic layers (Fig. 15A), and an upper massive section of mainly leucocratic gabbro. The cumulate sequence is best exposed along the shore of Glover Island south of Bluff Head and along Keystone Brook, whereas the upper leucocratic gabbro is well exposed on the barren tops south of Bluff Head (see map). Grading within the cumulate sequence is consistently east-facing. This thickness of cumulate layers ranges from a few centimetres to a metre, with most in the decimetre range. Small patches of pegmatitic hornblende gabbro occur in the upper parts of the complex. The gabbro contains a primary mineral assemblage of hornblende-clinopyroxene-plagioclase-opaques. The plagioclase is pseudomorphically altered to albite and clinozoisite. Knapp (1982), on the basis of microprobe analysis, showed that the clinopyroxene is of salitic composition and the hornblende is a ferroan-pargasite.

Ultramafic and related rocks of the Grand Lake Complex include talc-schist, variably serpentinized peridotite, and wehrlite. They are best developed along the western base of the complex but also occur as isolated bodies within the complex (see map). The base of the complex is characterized by variably serpentinized and metasomatized ultramafic rock abruptly and structurally overlying the metaclastic schist of the South Brook Formation. In Keystone Brook and south, the serpentinite is separated from the schist by up to 30-40 m of green to purple banded and massive greenschist, consisting of chlorite, epidote, plagioclase, carbonate and opaque oxides. This greenschist probably represents mafic igneous rock, metamorphosed and metasomatized during tectonic emplacement. Orange-brown weathering talc-carbonate rocks overlie the greenschist in Keystone Brook, and elsewhere mark the base of the Grand Lake Complex. They consist of talc, iron carbonate, serpentine and oxide minerals, and rare fuchsite. White- to orange-weathering massive serpentinized peridotite, up to several hundred metres thick in Keystone Brook, directly overlies the talc-carbonate rocks. The ultramafic rocks contain serpentine minerals, tremolite, magnetite, and relict igneous clinopyroxene, and were probably originally wehrlite. Knapp (1982) determined a diopsidic composition for the clinopyroxene, and noted the presence of probable orthopyroxene exsolution lamellae. Tremolite occurs as overgrowths on clinopyroxene and the serpentine (antigorite) presumably replaces olivine and/or orthopyroxene.

Serpentinized ultramafic rocks also form discontinuous lenses up to a kilometre long within lower segments of the gabbro sequence. In contrast to the serpentinized ultramafic rocks along the base of the complex, which are often penetratively deformed, these bodies are little deformed, and also generally less serpentinized. They range from wehrlite to clinopyroxenite.

Mafic dykes, from 30 cm to 2 m thick, with sharp, linear, well developed chilled margins, intrude the ultramafic and gabbroic segments of the Grand Lake Complex. They consist of clinopyroxene, partially rimmed by actinolite, albitized plagioclase, chlorite, and intergranular titanite, the last at least in part replacing an opaque mineral. The dykes are generally aphyric with local plagioclase-phyric segments. Dyke orientations are consistently northwest striking and steeply dipping at an angle to the regional foliation. Near Kettle Pond, Knapp (1982) noted pargasite-bearing and titanaugite-bearing diabase dykes intruding gabbro. Although these dykes are restricted to the Grand Lake Complex, their sharp contacts with the gabbro indicate a time gap between gabbro emplacement and dyke injection, and it is not clear if they are genetically related to the other rocks of the complex.

Trondhjemite, occurring locally at the top of the complex, consists of blue quartz, albitized plagioclase, and primary ferromagnesian minerals replaced by chlorite. It outcrops both as isolated bodies possibly representing small plutons, and as small intrusions (<5 m diameter) into gabbro. At least some trondhjemite outcrops have been incorporated into the Kettle Pond shear zone. Knapp (1982) mapped a greenschist unit at the top of the Grand Lake Complex which he considered constituted a host for the trondhjemite. Our mapping shows that this unit is part of the matrix of the shear zone, and is therefore removed from the complex.

Age

The intrusive relationship of the trondhjemite into the gabbro indicates that this provides an upper age limit for the complex. The close spatial association of trondhjemite and gabbro suggests the two are petrologically related and probably of similar age.

Trondhjemite from near the upper, eastern margin of the Grand Lake Complex (g.r. 459993) was collected for U-Pb age determination. Two fractions of fine-grained zircon were analyzed. The poor ratio of radiogenic Pb to common Pb in analysis Z1 is reflected in the large error ellipse (Fig. 16A; Table 1). The fraction Z2 is perfectly concordant and reliable, and the two analyses combined indicate an age of 490 ± 4 Ma for crystallization of this rock. No inherited component was detected.

Glover Formation

Distribution, definition, internal divisions, and thickness

The Glover Formation consists of a sequence of mafic to silicic volcanic and high level intrusive igneous rocks and minor volcaniclastic rocks that underlie the northern and eastern

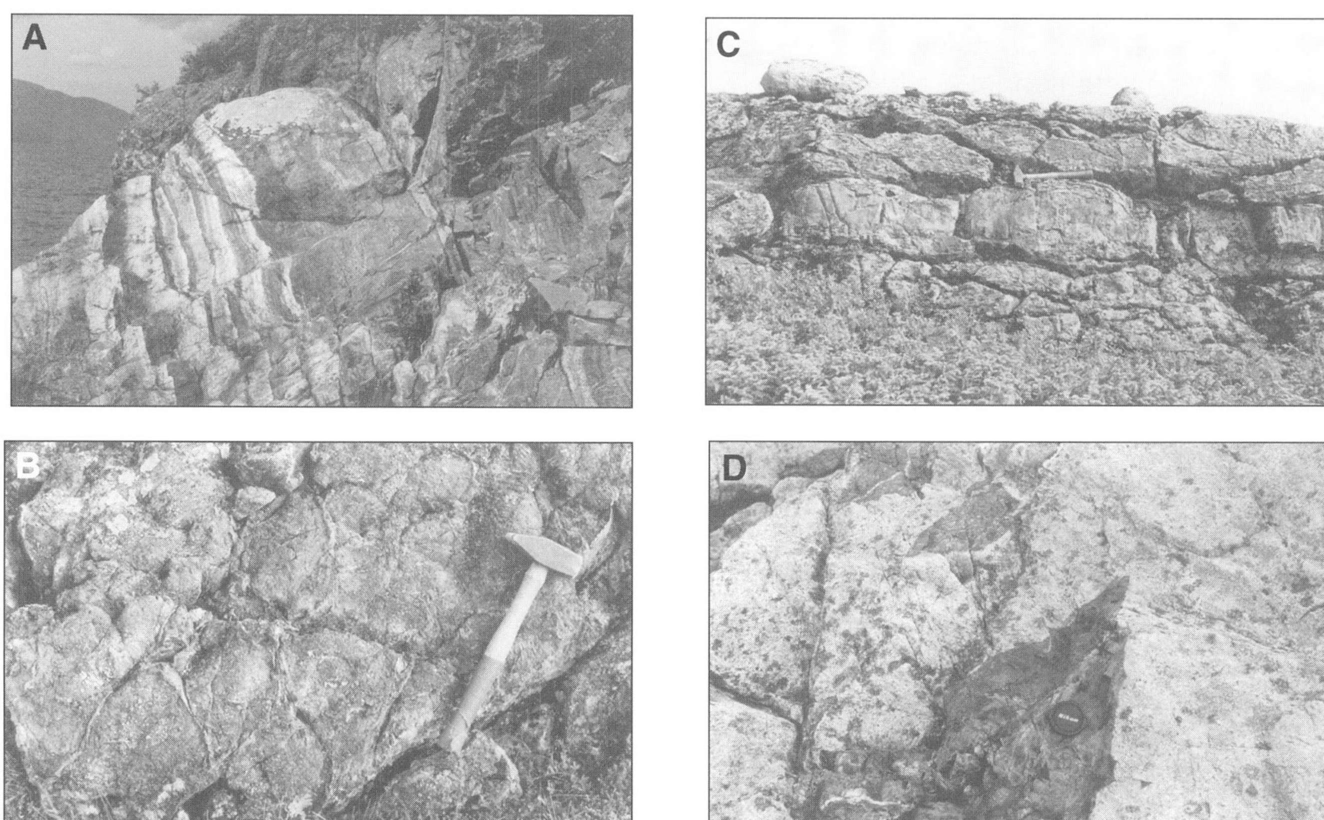


Figure 15. *Rocks of the Glover Island Belt.* ***A****) Layered gabbro of the Grand Lake Complex (map unit OGg) exposed south of Bluff Head on the west shore of Glover Island (g.r. 466050). Outcrop width is approximately 3 m. GSC 1995-149A* ***B****) Mafic pillow basalt of the Glover Formation (map unit OGF) exposed east of Grand Lake (g.r. 608054). GSC 1995-149Q* ***C****) Flattened pillow basalt with an intercalated 30-40 cm thick diabase sill (below hammer) of the Glover Formation (map unit OGF) exposed east of Grand Lake (g.r. 616053). GSC 1995-149R* ***D****) Mafic xenoliths of Glover Formation in Glover Island Granodiorite on the east shore of Glover Island (g.r. 533033). GSC 1995-149D*

segments of Glover Island as well as the area east of Grand Lake. Exposure along the shores of the lake, particularly the northwest coast of Glover Island is excellent, but outcrop in the interior of the island and east of the lake is poor.

The term Glover Formation was originally introduced by Riley (1957) for all the volcanic rocks east of the granite gneiss and metaclastic schist (Corner Brook Lake Complex and South Brook Formation). Knapp (1982) raised the unit to group status and redefined it to include the Kettle Pond and Tuckamore formations, and excluded rocks he informally assigned to his Grand Lake complex, Otter Neck group, Red Point formation, and Corner Pond formation (the last three units lie outside the present map area). The Kettle Pond formation was restricted to a sequence of predominantly quartz-sericite schists stratigraphically overlying the Grand Lake complex, whereas the Tuckamore formation corresponds largely with the original character of the unit as established by Riley (1957). Our mapping has shown that the rocks mapped by Knapp (1982) as the Kettle Pond formation, as well as immediately adjoining segments in the Tuckamore formation and Grand Lake Complex, lie within a large shear zone, and that the contact with the Grand Lake Complex is structural

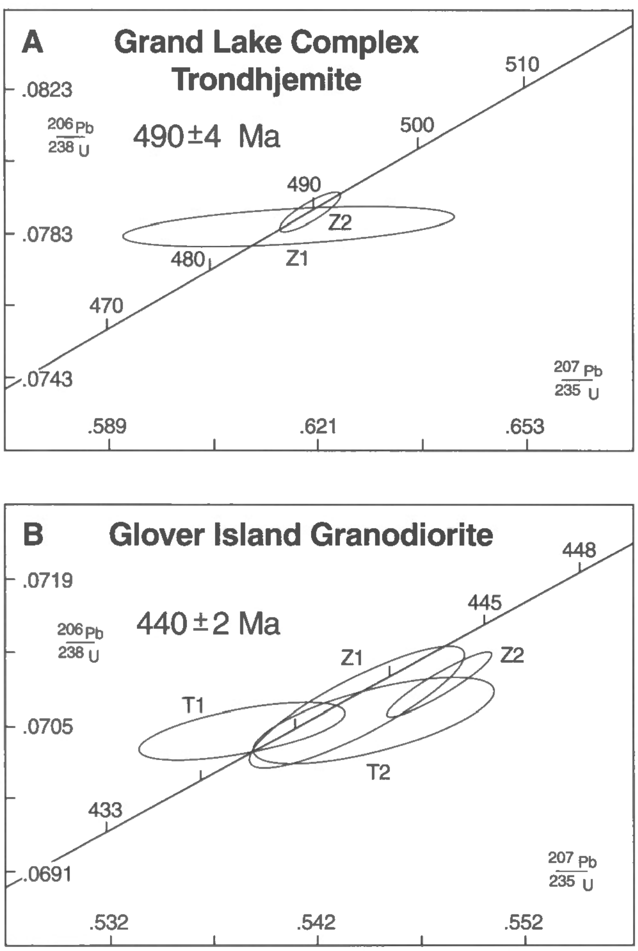


Figure 16. *U-Pb concordia plots for the Glover Island Belt.* ***A)*** *Trondhjemite at the top of Grand Lake Complex.* ***B)*** *Glover Island Granodiorite.*

rather than stratigraphic. We therefore discontinue use of the Kettle Pond and Tuckamore formations as stratigraphic units and revert to the term Glover Formation for igneous rocks east of the Grand Lake Complex. The shear zone separating these two units is herein informally referred to as the Kettle Pond shear zone.

The Glover Formation is locally informally divisible into a unit of mafic igneous rock, mainly diabase and pillow basalt (map unit OGFp), and a unit of interbanded mafic and felsic volcanic rock and tuff (map unit OGFf). In the interior of Glover Island where outcrop is poor, no attempt was made to subdivide the formation. The mafic igneous unit is well exposed on the eastern shore of Grand Lake and on the barren hills to the east, but is also exposed on the northwestern shore of Glover Island, opposite Northern Harbour. The interbanded volcanic and tuffaceous sequence is best observed on the western shore of Glover Island, north of Bluff Head (see map). Rare younging directions determined from pillow basalts and least deformed tuffaceous rocks in both units, are consistently to the east which, combined with the steep orientation of bedding and the extensive across-strike width of the formation, imply a thickness of at least several thousand metres. However, outcrop is insufficient, particularly in the inland sections, to determine if there is internal repetition within the formation, or if the relationship between the mafic sequence and the interbanded mafic and felsic sequence is stratigraphic or structural. Minor shear zones within the well exposed coastal sections on Grand Lake suggest at least some structural disruption of the formation. The overall inferred east-facing of the unit suggests that the pillow basalts and diabase east of Northern Harbour and east of Grand Lake form the stratigraphically lowest and highest segments of the formation, and that the interbanded mafic and felsic volcanics lie within the lower to intermediate parts of the formation.

Lithology

Mafic igneous lithologies within the Glover Formation range from pillow lavas (Fig. 15B), tuffs and massive flows to high level intrusive diabase and microgabbro. They occur throughout the formation but extrusive phases dominate along the eastern half of the island and east of Grand Lake. Well developed diabase sills and dykes (Fig. 15C), and small microgabbro intrusions occur within the pillow basalt sequence opposite Northern Harbour and east of Grand Lake. Pillows range in diameter from 20 cm to over 1 m, with most averaging between 30 cm to 60 cm. Deformation is less east of Grand Lake than on the west side of Glover Island.

The mafic igneous rocks consist of albitized plagioclase phenocrysts in a groundmass of actinolite, epidote, chlorite, titanite, opaque minerals, and rare biotite. Calcite amygdules were observed at a few locations. Rare igneous amphibole, determined to be titaniferous pargasite by Knapp (1982), occurs in some samples, but primary clinopyroxene is rare or absent.

Silicic igneous rocks, ranging from dacitic to rhyolitic volcanic rocks and tuffs, occur interstratified with the mafic rocks in western sections of the formation, and are well exposed along the northwest shore of the island. They contain quartz and albitized plagioclase phenocrysts in a mesostasis of quartz and albite with minor chlorite, probably derived from devitrified glass. Quartz grains show well developed volcanic embayments. The tuffs locally contain oversized flattened pumice fragments up to 10 cm in diameter. Thin bands of extensive pyrite alteration, locally up to several metres wide, occur within both the mafic volcanic rocks and silicic tuffs.

Age and relations

No direct age data are available for the Glover Formation. However, outside the map area, Knapp (1982) considered the formation to be unconformably overlain by the Corner Pond formation, which contains Arenig graptolites and conodonts (L.V. Rickards, pers. comm., in Dean, 1978; G.S. Nowlan, pers. comm., in Knapp, 1982; D. Skevington, pers. comm., in Knapp, 1982; Williams, 1989). This suggests an Early Ordovician or older age for the formation.

Matthews Brook Serpentinite

Thin fault bounded slivers of serpentinized ultramafic rock occur at the northern end of the Corner Brook map area, just east of the Trans Canada Highway, in Matthews Brook (g.r. 483264), and near the northwest corner of the Little Grand Lake map area (g.r. 308989 and g.r. 308998). All localities are herein termed the Matthews Brook Serpentinite. The serpentinite at the Matthews Brook locality is restricted to a narrow sliver along, and immediately along strike from, the lower reaches of the brook. It is generally altered to an assemblage which includes talc, carbonate, magnesite, actinolite, chlorite, siderite, magnetite, fuchsite, and quartz. Asbestos fibres are locally present. The localities in the Little Grand Lake map area consist of two slivers, each less than 100 m long and approximately 1 km apart. The southern block was originally mapped by Kennedy (1981) and both are less altered than the Matthews Brook locality. They consist largely of green serpentine (antigorite), plus minor chromite and opaque minerals. Magnesite and talc are rare alteration products. Although Kennedy (1981), on the basis of the pronounced aeromagnetic anomaly associated with the serpentinite, proposed that other anomalies in the area might also represent ultramafic material, our mapping failed to locate any.

The Matthews Brook Serpentinite at the northern locality is juxtaposed against quartz-mica schist of the South Brook Formation, but is also near amphibolite and minor granitoid gneiss of the Corner Brook Lake Complex (see map). The southern serpentinite localities are juxtaposed against the Breeches Pond and South Brook formations.

LATE ORDOVICIAN AND YOUNGER ROCK UNITS

Igneous and sedimentary rock units postdating the mid-Ordovician and older Humber Zone and Dunnage Zone sequences occur within the map area. Principal amongst these sequences are Late Ordovician to Silurian silicic plutonic bodies, the largest of which is the Glover Island Granodiorite, and Carboniferous sedimentary rocks within the Deer Lake Basin.

Late Ordovician to Silurian igneous activity

Glover Island Granodiorite

Definition, distribution, and relations

The term Glover Island Granodiorite was introduced by Cawood and van Gool (1993) for an unnamed pluton exposed over some 35 to 40 km^2 on the eastern side of Glover Island (see map). The intrusive contact of the pluton into the surrounding Glover Formation is exposed at the southern margin of the body on the east shore of Glover Island (g.r. 490990). This region is characterized by numerous aplitic dykes cutting Glover Formation greenschist, as well as large rafts of greenschists, at least tens of metres across, within the pluton. Mafic xenoliths, presumably of the enclosing Glover Formation, occur throughout the pluton but are best observed in shoreline outcrops (Fig. 15D). Xenoliths range from centimetres to tens of metres with the majority less than 10 cm in diameter. Some contain a foliation which is at an angle to, and truncated by, the foliation in the granite, requiring deformation prior to inclusion within the granite. Granitic and aplitic dykes are common in the adjoining Glover Formation.

Lithology and character

The granodiorite is a medium-grained, cream to grey, equigranular, relatively uniform rock with no internal divisions. It is composed of quartz, plagioclase, biotite, and potassium feldspar, along with minor muscovite, titanite, epidote and an opaque phase, probably ilmenite or magnetite. Plagioclase is commonly sericitized, particularly in its cores, suggesting a normal zoning pattern to the crystals from sodic cores to more calcic rims. Potassium feldspar (microcline) is unaltered, polysynthetically twinned, and occurs locally in graphic intergrowth with quartz. Some biotite is strongly chloritized. Large euhedral titanite crystals occur in all samples, commonly in association with biotite, and smaller anhedral to subhedral titanite grains also occur as an alteration product with chloritized biotite. Muscovite is a common minor phase. Variation in microcline content of the samples locally gives the pluton a quartz monzonite, rather than a granodiorite, composition. Pegmatitic and aplitic dykes up to 1 m across, and mineralogically similar to the main granodiorite, locally cut the pluton and are assumed to be consanguineous with it.

Age

A sample of coarse-grained Glover Island Granodiorite (g.r. 521085) collected for U-Pb dating, yielded both titanite and brown zircon, the latter with a high uranium content. One zircon fraction (Z2, Fig. 16B; Table 1) is displaced to the right of concordia and is interpreted to contain a minor component of inherited zircon. The remaining fractions are concordant within uncertainties. The titanites yield ^{206}Pb-^{238}U ages of 439 ± 2 Ma and 439 ± 3 Ma while that of the zircon is 440 ± 3.5 Ma. These in combination are taken to indicate an age of 440 ± 2 Ma for igneous crystallization for this granodiorite.

Red Indian Brook Granodiorite

Distribution, character, and correlation

The Red Indian Brook Granodiorite outcrops along the east shore of Grand Lake from near the mouth of Red Indian Brook to a point some 4.5 km north, and occupies some 8 km^2. The pluton is compositionally similar to, and may represent an offshoot of, the Glover Island Granodiorite. Any contact between the two lies under the east arm of Grand Lake. At the mouth of Red Indian Brook (just south of the area on the east shore of Grand Lake) and south of the brook the pluton is intrusive into diabase and gabbro. Numerous aplitic and pegmatitic dykes up to several metres wide and intrusive into Glover Formation were observed in, and around, Conners Brook. One small outcrop of felsic igneous rock in fault contact with adjoining gabbro of the Glover Formation, and representing either an aplite dyke related to either the Glover Island/Red Indian Brook granodiorites (cf. Hyde, 1979b) or a felsic phase of the Glover Formation, occurs on the shore of Grand Lake some 3 km south of the mouth of Conners Brook (g.r. 577055).

Plutons along the Cabot Fault system: Little Paddle Point and Island Pond plutons

Definition and distribution

Two fault bounded blocks containing deformed and altered granitoid plutons intrusive into mafic igneous rocks occur along the Cabot Fault. The southern block extends along the western shore of Grand Lake in the vicinity of Little Paddle Point for some 16 km, but its exposed width is never more than 100 m. It is faulted against the South Brook Formation in the west, and its eastern margin is not exposed but is inferred to be faulted by the splay of the Cabot Fault lying along the southwestern arm of Grand Lake. The northern block extends from Northern Harbour to north of Island Pond, a distance of 18 km (Hyde, 1979b, 1982). The maximum width of the block is approximately 1 km. Plutonic rocks within these two blocks are referred to as the Little Paddle Point pluton for the southern block and the Island Pond pluton for the northern block. Mafic igneous rocks are assigned to the Glover Formation.

Lithology

The Little Paddle Point pluton consists of strongly brecciated and altered medium- to coarse-grained granite and associated pegmatite. The rock is quartz monzonitic with approximately equal proportions of plagioclase and potassium feldspar, and quartz contents of greater than 20%. It is locally cut by pseudotachylyte. Hematite staining is pervasive and no primary ferromagnesium phase is preserved.

The Island Pond pluton consists of medium- to coarse-grained granite and pegmatite consisting of quartz, plagioclase, minor potassium feldspar, and biotite. The biotite is pervasively chloritized and the feldspar is hematite stained. The pluton is strongly foliated with aligned and stretched quartz and feldspar crystals, is locally mylonitic, and is disrupted by late brittle fracturing with widespread chlorite development on fracture planes.

Age and relations

The quartz monzonitic to granodioritic composition of the Little Paddle Point and Island Pond plutons and their emplacement into mafic igneous rocks of the Glover Formation suggests they may be temporally and petrologically related to the Glover Island Granodiorite.

Unnamed granite/pegmatite

Distribution, relations, and character

Coarse pegmatite and associated minor granite are intrusive into the Corner Brook Lake Complex and the Mount Musgrave Group of the Humber Zone around the southwest margin of Corner Brook Lake. Most are isolated outcrops, less than 100 m in diameter, and too small to show

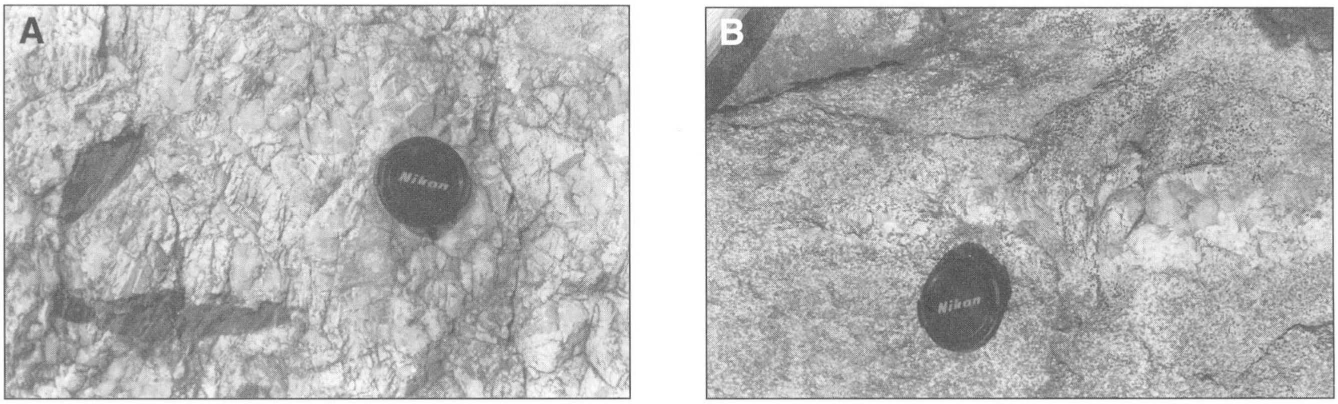


Figure 17. *Rocks of the pegmatite suite (map unit SP) near the southern end of Corner Brook Lake.* ***A****) Coarse grained massive pegmatite containing feldspar (white), quartz (grey), and amphibole (dark), exposed on the shore of Corner Brook Lake (g.r. 392074). GSC 1995-149HH* ***B****) Pegmatite grading into foliated granite, exposed on the shore of Corner Brook Lake (g.r. 392074). GSC 1995-149LL*

on the map. The concentration of intrusions within one geographic area and the similar mineralogy of the outcrops suggest all are consanguineous. The best exposures of the granite/pegmatite sequence are seen on the small peninsula in Corner Brook Lake (g.r. 392074) and along the stream draining into the lake (g.r. 394055). Finer grained phases are granitic and consist of quartz, calcic and potassic feldspar, biotite, muscovite, and magnetite, with rare amphibole. Epidote, chlorite, and hematite are alteration products. Pegmatitic phases are dominated by potassium feldspar with some quartz, minor muscovite, and rare green amphibole (Fig. 17A). Potassium feldspar crystal are up to 15 across, muscovite books are up to 10 cm across, and tabular amphibole crystals are up to 30 cm long.

Deformation of the granite/pegmatite is highly variable with finer grained granitic phases showing a weak to moderately developed principal foliation (Fig. 17B) defined by mineral alignment cut but a later fracture cleavage. Pegmatitic phases are relatively massive and undeformed. Xenoliths of country rock within the least deformed phases, as well as country rock adjacent to pegmatitic dykes, are also little deformed, suggesting that the coarser grained parts of the intrusion formed a component lithology relatively resistant to deformation.

Age

A sample of pegmatite (g.r. 394055) was collected for U-Pb dating. This sample yielded zircons of variable morphology, including very large grains and needles. An inherited core component is visible in some grains under the microscope. A fraction of needle-like grains (Table 1; Z2) and a fraction of only 8 grains (Z4) were analyzed in an attempt to avoid inheritance. While inheritance is minor in three analyses it was not eliminated, and the calculated mixing line (60% probability of fit) yields a lower intercept age of crystallization of 434 +2/-3 Ma (Fig. 18). The upper intercept is 1021 ± 68 Ma and might indicate a source of predominantly this age.

Carboniferous Deer Lake Basin

Carboniferous rocks of the Anguille and Deer Lake groups of the Deer Lake Basin are restricted to the northern end of the map area (see map). Outcrops are generally poor, except along shoreline and stream sections. The geology of the basin was studied in detail by Hyde (1978, 1979a, b, 1982, 1984) and only an overview of the rock units is given here.

Exposures of the *Anguille Group* (Hayes and Johnson, 1938) occur along the northwest shore of Grand Lake and east of the Northern Harbour road. Plant and spore data from the group indicate a Tournasian age (Heyl, 1937; Baird, 1959; Popper, 1970; Barss, 1974). The group is divisible within the map area into the basal Blue Gulch Brook Formation and the overlying Saltwater Cove Formation (see map; Hyde, 1982).

The *Blue Gulch Brook Formation* (map unit CB) consists of grey pebble to cobble conglomerate, along with grey micaceous sandstone, dolomitic sandstone, and rare impure limestone. Conglomerate horizons contain carbonate and quartz clasts, derived from the Humber Zone basement. The unit runs along the western flank of the Fisher Hills, and is base faulted along its western margin by the Deer Lake Fault. It is conformably overlain by the Saltwater Cove Formation. Hyde (1979a) estimated a thickness for the unit of 400 m.

The *Saltwater Cove Formation* (map unit CS) consists of grey sandstone and siltstone, black carbonaceous mudstone and minor interbedded conglomerate, limestone and dolostone. Clast types within the conglomerate include quartz, felsic volcanic and plutonic rock, quartz-mica schist, and limestone. The unit is well exposed along the northwest shore of Grand Lake, north of Northern Harbour, but extends inland as far west as the Fisher Hills. Outcrop is disrupted in the vicinity of Island Pond into two belts by a fault-bounded basement block consisting of Dunnage Zone mafic igneous rocks intruded by silicic plutonic rocks. Hyde (1979a) estimated a thickness for the formation of 2700 m. Paleocurrent data from outcrops along the west side of Grand Lake indicate paleoflow to the northwest and northeast (Hyde, 1979a).

The *Deer Lake Group* (Werner, 1956) is composed of poorly indurated siliciclastic rocks along with oil shale and limestone. Two formations outcrop within the map area: the basal North Brook Formation and the overlying Rocky Brook Formation (Werner, 1956; Hyde, 1979a, b, 1982). The latter unit gave a late Viséan to early Namerian age (Belt, 1969; Hyde, 1979b). The Deer Lake Group is inferred to unconformably overlie the Anguille Group (Hyde, 1979a), and basement rock units. The group outcrops in two structural belts separated by the uplifted block of Anguille Group lithologies. The northern belt lies northwest of the Deer Lake Fault and its western margin is unconformable on Humber Zone lithologies of the Mount Musgrave Group. The southeastern block lies to the east of the Island Pond Fault and its eastern margin is marked by the Grand Lake Fault.

The *North Brook Formation* (map unit CN) consists of red-brown siltstone and sandstone with interstratified pebble conglomerate and minor limestone. Within the eastern block of the formation, clast types within the conglomerates include granitoid, silicic volcanics, and quartz, whereas in the northern block clast types are mainly quartz pebbles with minor

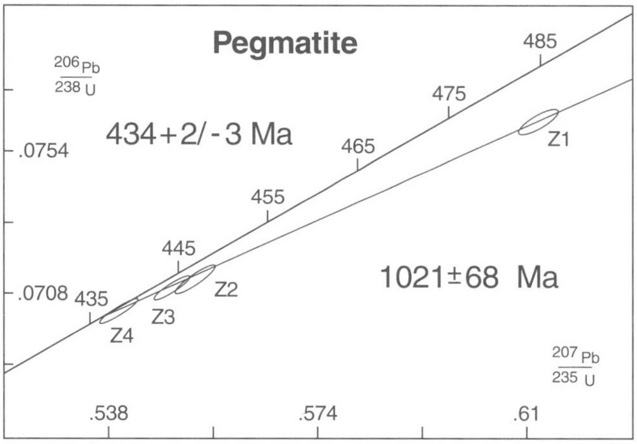


***Figure 18.** U-Pb concordia plot of unnamed pegmatite.*

chloritized mafic volcanic rocks. The formation is well exposed around the northern end of Glover Island where it attains its maximum thickness of 1200 m (Hyde, 1979a).

The *Rocky Brook Formation* (map unit CR) contains red-brown siltstone and mudstone with sandstone and oil shale. It attains a thickness of around 1500 m. The unit, within the map area, is restricted to the eastern block of the Deer Lake Group, exposed on the east shore of Grand Lake by Grand Pond Point.

STRUCTURAL DEVELOPMENT

Assembly of the rock units into their present configuration caused a structural imprint which can be recognized as several major phases of deformation in each of the lithotectonic units, some of which can be correlated across unit boundaries. The description of these structures is based on relations displayed in the geological map and cross-sections (in pocket), plots of orientation data, and descriptions, sketches and photographs of outcrops.

The cross-sections were constructed using data both from along the lines of section and from areas to the north and south which were projected onto the plane of section, supplemented with interpretations at depth (e.g. the level of the basement underneath the Deer Lake Basin in sections A and B). The field sketches in the sections are of representative outcrops taken predominantly from along the line of section, but locally also from outcrops at a distance, resulting in them being projected either into the subsurface or above the topography. They are oriented in the same way as the sections, which means that some appear as a mirror image of their outcrop appearance. Note that sketch D2 is in plan view.

Structures are described separately for five of the six lithotectonic units shown in Figure 4 (not including the Silurian granitoids). These units are in most cases separated by late structures. The orientation data are subdivided and presented for several domains in each of the lithotectonic belts. Where sufficient data were available, the data points were contoured using a Gaussian counting function. Eigenvectors were calculated to construct π-girdle and fold axis orientations.

Map pattern and cross-sections

The five lithotectonic units (Fig. 4) define the large-scale structural patterns of the Corner Brook-Glover Island region. The Humber Arm Allochthon was originally emplaced as a thrust sheet in a west-directed movement and contains cryptic west-vergent structures. It is now dominated by east-vergent inclined to overturned folds and the east-directed Crow Hill thrust. The Carbonate Belt forms a south-tapering wedge which lenses out between Pinchgut and Corner Brook lakes. It consists in the north of kilometre-scale anticline-syncline pairs (cross-section B) one of which dies out to the south, where the belt contains stacked thrusts. The carbonate sequence at the southwest corner of the Corner Brook-Glover Island region, although containing the same stratigraphic units as the Carbonate Belt, occupies a different structural position and is part of the Corner Brook Lake Belt. The Corner Brook Lake Belt is a fold-and-thrust belt in which Proterozoic basement and its late Proterozoic to early Paleozoic cover were thrust, folded and faulted in a series of deformational events. The belt consists in the southwest of kilometre scale upright to inclined folds, which are bounded to the east by a stack of thin-skinned thrust slices, the Valley Lakes thrust stack, which in turn are overlain by a many kilometres thick fold nappe consisting of metaclastic rocks of the South Brook Formation, named the Yellow Marsh fold nappe. The Corner Brook Lake Belt is truncated to the east by the Cabot Fault and is in the north unconformably overlain by Carboniferous sedimentary rocks.

In the Glover Island Belt the Glover Formation and Grand Lake Complex show a less complex ductile deformation history than rock units of the Corner Brook Lake Belt. The belt is dominated by a steep foliation related to the development of

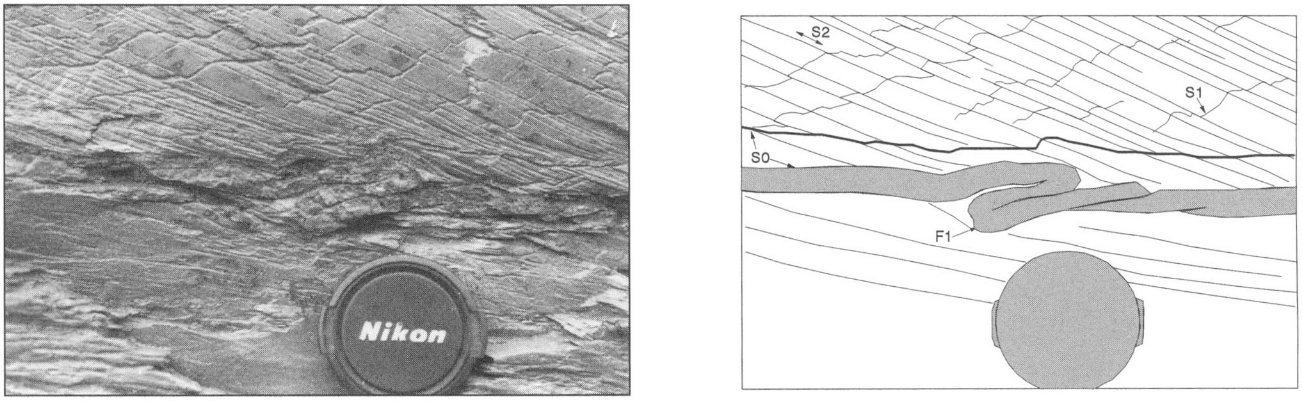


Figure 19. *Photo and line drawing of isoclinal F_1 fold in quartzite layer in slate of the Summerside Formation west of Crow Hill (g.r. 285227), overprinted by a penetrative S_2 fabric which dips to the west. The F_1 fold is west-vergent, whereas the S_0-S_2 angular relationship is east-vergent. Weathering has emphasized the S_1 foliation, which is axial planar, but, as a result of later deformation, at an angle to the F_1 fold. Looking south. GSC 1995-149KK*

the Kettle Pond shear zone that separates the two lithological units. In the west-central part of Glover Island, vestiges of the Humber Zone are preserved, and are in fault contact with rocks of the Dunnage Zone. All lithological units on Glover Island and their separating shear zones are folded in a south-plunging antiform, termed the Kettle Pond antiform, which is cut by faults of the Cabot Fault system.

The Carboniferous rocks in the northeast of the area form part of a pull-apart basin, created by strike-slip movement on the Cabot Fault system. Most contacts in map view are high-angle faults which are predominantly transcurrent, but some have large extensional components. Locally compressional configurations of the fault system caused intense folding and penetrative deformation of the Carboniferous strata.

Humber Arm Allochthon

The rocks in the Humber Arm Allochthon display evidence of at least three phases of deformation, and are at lower greenschist metamorphic grade, with either brittle or ductile structures. The map pattern of the allochthon is dominated by an east-vergent fold train and a repetition of the stratigraphy in a stack of east-vergent thrust sheets, resulting in an interfingering of rocks of the Summerside, Irishtown, and Pinchgut formations. These relationships are best displayed in, and just south of, the city of Corner Brook (see map, cross-sections B and C). The east-vergent thrusts and thrust-related folds and foliation are the dominant structures on both a macroscopic and a mesoscopic scale. The main foliation in the allochthon is a second generation structure (S_2).

D_1 west-vergent folding

The oldest structural event (D_1) in the Humber Arm Allochthon is a phase of west-vergent, thrust-related folding. D_1 structures are generally cryptic and only locally exposed. In a road outcrop west of Crow Hill (g.r. 285227) centimetre-scale, west-vergent, isoclinal folds in quartzite layers in thin-bedded slate (Fig. 19, 20, sketch B1) represent F_1 folds, which are overprinted by the younger S_2 foliation. On the Curling road, at the Nova recycling site (g.r. 264225, about 1 km west of the Curling turnoff, just outside the map area), an F_1 fold of several metres wavelength is overprinted by an intense F_2 foliation and small-scale F_2 folds. Figure 21A is a sketch of the F_1 fold, which is not obvious in outcrop, but can be reconstructed by tracing bedding and bedding-cleavage relationships through the outcrop. The change in asymmetry of the small-scale F_2 folds and S_0-S_2 angular relationships do not agree with the shape of the fold and must postdate the larger scale fold. Larger F_1 folds, overprinted by D_2 structures, were recognized on a larger scale across the road from this outcrop (Fig. 21B) and in rocks of the Summerside Formation, west of the community of Summerside. In other locations where asymmetries of small F_2 folds appeared inconsistent with large fold structures, F_1 folds are inferred, but not demonstrated.

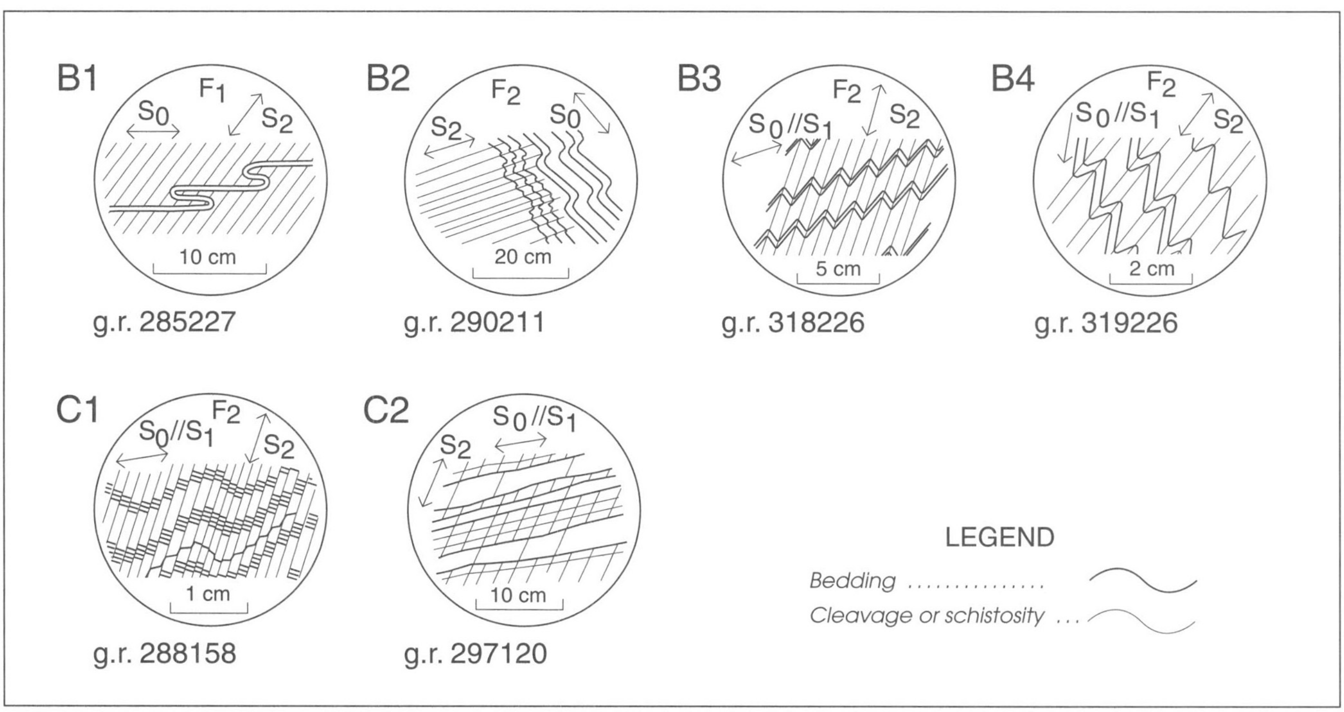


Figure 20. Sketches outlining structural relations within the Humber Arm Allochthon. Letters and numbers refer to the cross-sections (Map 1893A, in pocket) and their relative positions in the sections from which they were compiled. All sketches are oriented with the west to the left.

On the top of Crow Hill (g.r. 293225) sandstone of the Summerside Formation is overturned with fining-down graded beds. Since the angular relation between S_0 and S_2 suggests that this is a normal limb with respect to F_2, this must be an overturned limb of a recumbent F_1 fold. Where F_1 folds were recognized or interpreted, they are consistently west-verging.

S_1 is only rarely visible in thin section, such as in F_2 fold hinges, where S_1 and S_2 are perpendicular, or in rocks of low D_2 strain. It is a slaty cleavage, approximately parallel to bedding, consisting of elongated, quartz-rich microlithons, surrounded by films of sericite, chlorite, and opaque minerals.

No obvious large-scale D_1 structures were observed in the Humber Arm Allochthon, but they may be locally present, causing complications in the map pattern. West of the map area, approximately from the map boundary to the community of Cooks Brook, bedding is fairly steep, and is assumed to be the forelimb of a multi-kilometre-scale F_1 fold. This places the rocks of the Cooks Brook Formation around Cooks Brook (e.g. Williams and Cawood, 1989) in a large F_1 syncline. The area around Corner Brook forms a normal F_1 limb with an F_1 antiform at the western margin of the map.

D_2 east-vergent folding and east-directed thrusting

The second phase of deformation is the dominant structural element in the rocks of the allochthon and defines the map pattern. It was a period of east-directed thrusting and folding, and S_2 formed as the dominant cleavage in the allochthon.

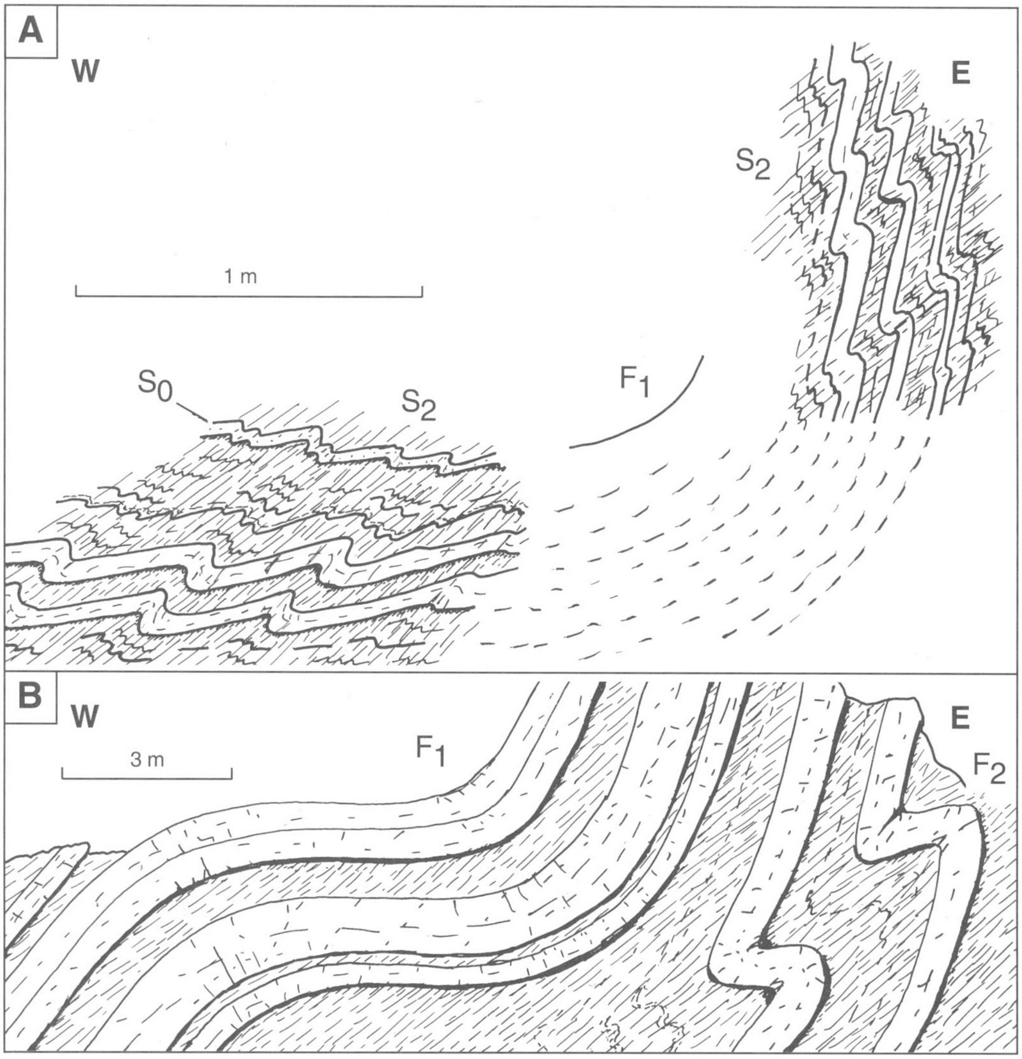


Figure 21. *Sketches of F_1-F_2 overprinting relationships in the Humber Arm Allochthon, west of Corner Brook.* ***A)*** *Fold asymmetries and bedding-cleavage relationships in a metre-scale F_1 fold, overprinted by smaller F_2 folds. Road outcrop of Irishtown slate (g.r. 264225).* ***B)*** *Large-scale west-vergent F_1 fold, overprinted by S_2 foliation which is axial planar to east-vergent F_2 folds. Road outcrop of interbedded sandstone and slate of the Irishtown Formation (g.r. 264225).*

F_2 folds are cleavage folds which range in size from milli-metres (Fig. 20) to kilometres (see map and cross-sections B and C). They have a well developed axial planar cleavage, S_2, in the slates. Fold styles vary with lithology from predominantly parallel in sandstone and bedded carbonate, to similar in slate. The folds are consistently asymmetric and predominantly east-vergent (Fig. 22A). Axial planes are steeply to moderately west dipping (Fig. 23). F_2 fold axes in outcrop show a range of orientations in a moderately west-dipping plane, which has approximately the same orientation as the majority of the S_2 foliation (Fig. 23), suggesting that F_2 folds are curvilinear, as was locally observed in the field.

S_2 is predominantly a slaty cleavage (Fig. 14A, 19), but in massive sandstone beds or bedded limestone, it is either a fracture cleavage or it is not developed (Fig. 14C). Solution along the cleavage planes cause an apparent contractional or extensional offsets of bedding or S_1 planes along S_2 cleavage planes (Fig. 19). Petrographic studies showed that the slaty cleavage is commonly a crenulation cleavage on a microscopic scale (Fig. 22B). Microlithons of quartz, feldspar, and chlorite stacks are separated by bands of fine-grained muscovite, chlorite, and opaque minerals. High concentrations of opaque oxides along the cleavage planes may also be a result of solution along S_2. The quartz-rich microlithons locally preserve hinges formed by a relict S_1 or cross-micas. The fracture cleavage in the limestone and quartzite is assumed to be formed by pressure solution.

The Seal Head syncline is a map-scale F_2 structure exposed in several locations in downtown Corner Brook (Fig. 14B, 22C). The best exposure is along the abandoned railway yard at Seal Head (g.r. 314236). The western limb of the structure is steeper than the eastern limb, in accord with the east vergence of the D_2 structures. Although, this syncline cannot be traced to the south of Corner Brook Stream, it lies in the extension of a syncline with an associated out-of-the syncline thrust just south of the Corner Brook ring road (cross-section B; g.r. 305193). It is the only well-defined large-scale F_2 fold in the allochthon. Other large-scale F_2 folds have been reconstructed by tracing small-scale F_2 fold asymmetries, bedding-cleavage angular relationships, and stratigraphic boundaries.

Lineations are common in the slates and are formed by the intersection of S_0, S_1, and S_2 planes. Bedding intersections on the cleavage planes are common and have orientations that are approximately parallel to the meso-scale fold axes. S_1-S_2 intersection lineations ($L^2{}_1$) and S_0-S_1 intersections ($L^1{}_0$) are rare.

Although meso-scale F_2 fold axes are shallowly north plunging, the overall plunge of the structures in the allochthon is towards the south, as expressed in the exposure of progressively higher stratigraphic units towards the south. This regional southern plunge may be caused by the earlier D_1 deformation.

The F_2 folding is related to a phase of east-directed thrusting. Several small and large scale thrusts faults were observed, and others were inferred from stratigraphic repetition. The thrusts dip shallowly to moderately to the west, and the few measured stretching lineations on the fault planes suggest that the thrusting was slightly oblique to the east-northeast, but the low number of measurements (3) and the spread of their orientations makes the actual direction of movement uncertain. Asymmetries of east-vergent F_2 folds related to the thrusting are the only kinematic indicators noted. The Crow Hill thrust is exposed at James Cook Lookout (Fig. 22D), on the road at the foot of Crow Hill, just east of Crow Gulch (Fig. 22E, F) and north of the Humber Arm, east of the community of Summerside. It placed rocks of the Summerside Formation on top of the younger Irishtown Formation. In most exposures the thrust surface is parallel to S_2 (Fig. 22E, F) with intense F_2 folding close to the thrust plane. Near the top of Crow Hill, the thrust is a sharp planar feature that truncates S_2 in the footwall (Fig. 22D), which may indicate a later reactivation of the thrust fault. The thrust plane dips between 25° and 30° to the west.

Small intraformational reverse faults parallel to S_2 were observed in a few locations (e.g. on the ring road west of Crow Hill, g.r. 287227, and in outcrops along the southern shore of the Humber Arm). These are only visible in ramps, where they cut through bedding, and they may be more common than their low observation count would suggest.

D_3 fracture cleavage

In the eastern part of the Humber Arm Allochthon, near the contact with the Carbonate Belt, a weak, spaced cleavage (S_3) is developed in the slate. This is best observed in outcrops along the new part of the Trans Canada Highway east of the Maple Valley Industrial Park, and near Margaret Bowater Park and Corner Brook stream to the east. At the last location, S_2 is locally crenulated by S_3 but overall S_3 resembles more a fracture cleavage. The dip is generally steeply to the east-southeast.

Extensional faults

The youngest generation of structures in the allochthon is a set of extensional faults, which cut both the S_2 cleavage and the east-directed thrusts. An age relation with the S_3 cleavage could not be established. The offset on the faults where this could be determined was in the range of 1-10 m. Figure 24 shows an example of a normal fault at the Curling turnoff (g.r. 272223), and a similar extensional fault offsets the Crow Hill thrust at the foot of Crow Hill (Fig. 22E, F). Slickensides are common on these faults and indicate a downdip extensional displacement.

Orientation data

S_0 and S_1 planes in both domains in Figure 23 have approximately similar orientations. S_0 in domain 1 is concentrated in a cluster for the dominant west-dipping limb, but there is a weak great circle distribution, which defines a fold axis that is about parallel to the maximum orientations of F_1, F_2, and F_3 fold axes. In domain 2 the distribution of S_0 and S_1 planes shows a weak cluster, because the bedding is folded over F_1 and F_3 fold axes that are at right angles to the F_2 fold axes. S_2 forms a strong west-northwest-dipping cluster in both domains.

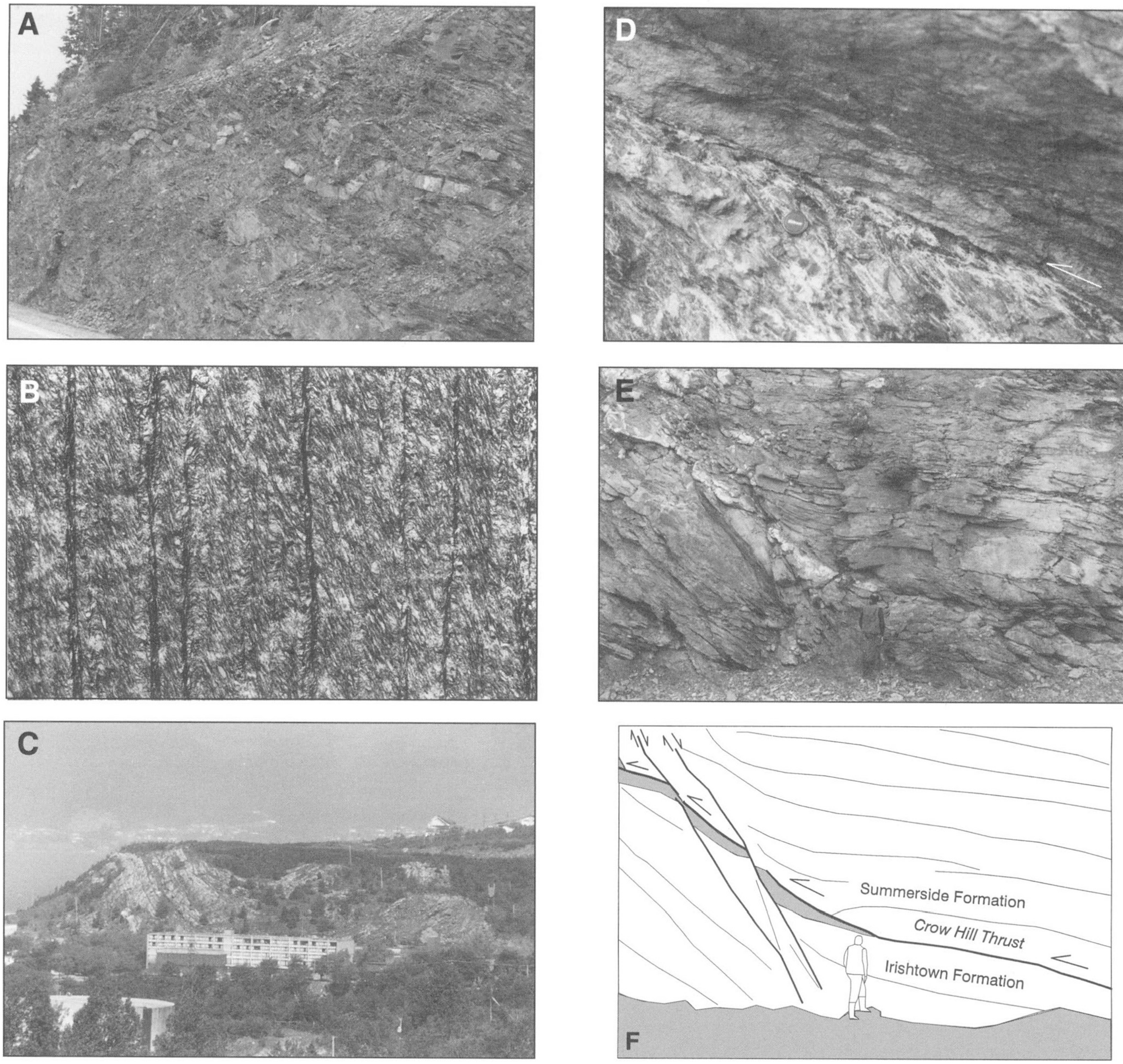


Figure 22. *D_2 structures in rocks of the Humber Arm Allochthon.* ***A****) Asymmetric F_2 folds in sandstone bed in slate of the Irishtown Formation, at the Curling Turnoff (g.r. 272222). The folds are east vergent and have a well developed axial planar cleavage, dipping to the right. View looking south. GSC 1995-149U* ***B****) Photomicrograph of an S_2 crenulation cleavage which mesoscopically forms a slaty cleavage in a slate of the Summerside Formation, east of Crow Hill (g.r. 280274). Fine-grained muscovite and chlorite define S_1, which is subparallel to S_0 and folded in F_2 crenulations. Opaque oxides and micas are concentrated along the cleavage planes. Plane-polarized light. Width of the photo represents 1.35 mm. GSC 1995-149MM* ***C****) Seal Head syncline (F_2) behind the Holiday Inn in downtown Corner Brook, viewed from the south. The rocks are thick-bedded sandstone and minor slate of the Irishtown Formation. GSC 1995-149J* ***D****) Exposure of the Crow Hill thrust at James Cook Lookout in Corner Brook. Sandstone of the Summerside Formation lies on top of slate of the Irishtown Formation. The thrust plane (diagonally through the photo, at the arrow) is quite sharp and truncates the steep S_2 foliation in the footwall. View looking southeast. GSC 1995-149S* ***E and F****) Photo (GSC 1995-149VVV) and line drawing of the exposure of the Crow Hill thrust in a road outcrop at the foot of Crow Hill in Corner Brook. Slate and fine sandstone of the Summerside Formation lie on top of slate of the Irishtown Formation. The thrust is offset by normal faults (viewed from the north).*

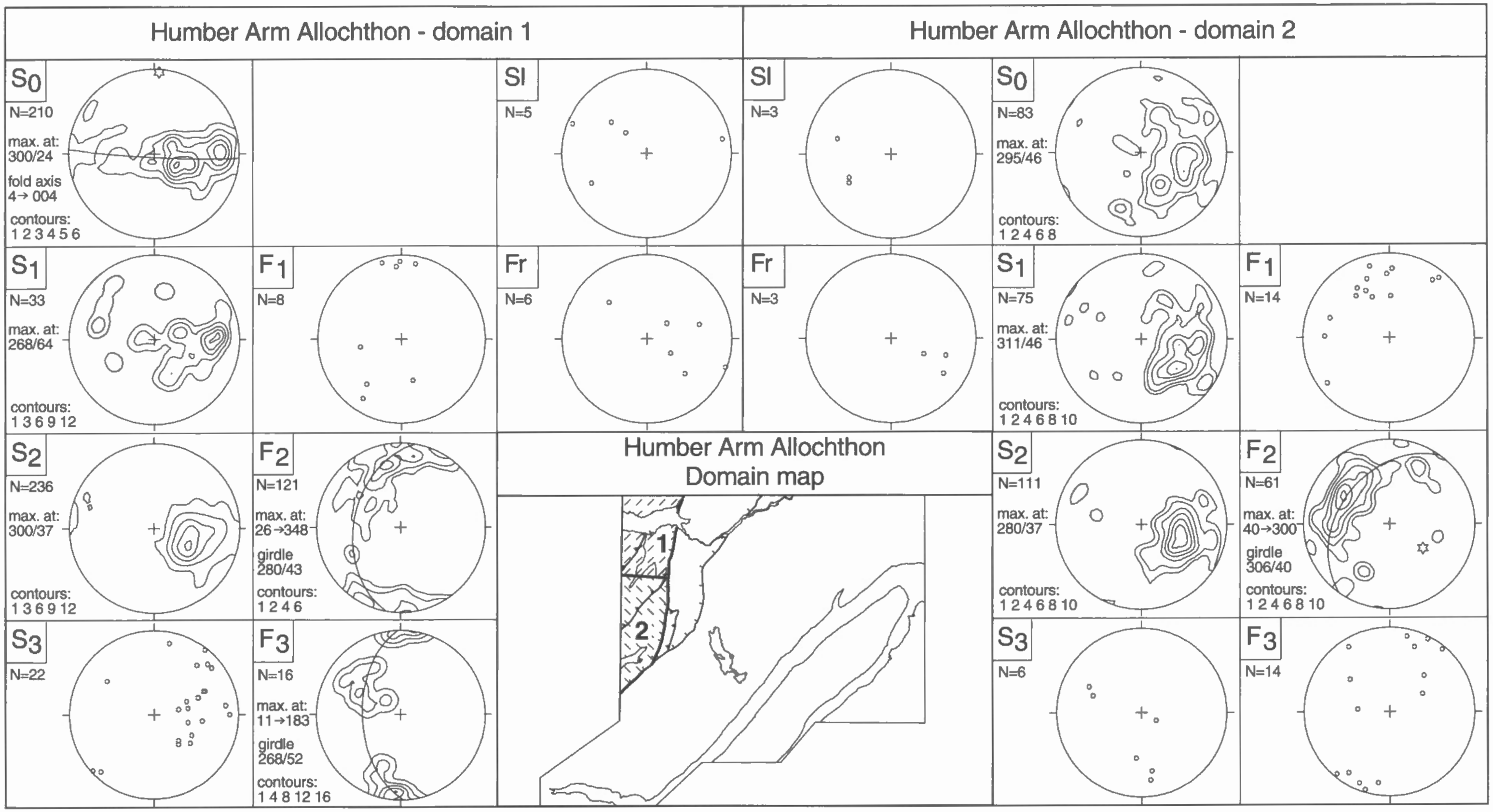


Figure 23. *Structural orientation data for the Humber Arm Allochthon. Linear data and poles to planar data, plotted on a lower hemisphere, equal area projection. Contour levels are in multiples of uniform distribution. Orientations of planes are quoted as dip direction and dip. Fr = fractures; Sl = slickensides.*

The distribution of the F_2 fold axes in a west-northwest-dipping great circle about parallel to S_2 is presumably a result of this overprinting of F_2 folds on previously folded strata. In the rocks of the Pinchgut Formation (domain 2) the F_2 fold axes are predominantly oriented down-dip in the S_2 plane, which may reflect a different orientation of the bedding in this domain, possibly because the Pinchgut Formation lacks thick-bedded, competent sandstone. F_3 in domain 1 has a similar orientation to F_2, but in domain 2 the F_3 is again mainly shallowly north-northeast or south-southwest-dipping, perpendicular to F_2 .

Carbonate Belt

Structures in the carbonate rocks of the Reluctant Head Formation and the Port au Port, St. George, and Table Head groups suggest that two penetrative phases of deformation have affected the Carbonate Belt, and locally three generations of penetrative structures can be recognized. These are followed by a phase of brittle faulting. Outcrop-scale structures are best developed in the thin-bedded limestone and slate of the Reluctant Head Formation, whereas the massive carbonates show them only in the southern part of the belt where deformation has a more ductile character than in the north. In the northern extension of the Carbonate Belt, around Old Man's Pond, Knight (1994) reported a first phase of deformation that involved west-directed thrusting, followed by a second phase of west-vergent folding, and a third phase of east-vergent fold-nappe generation. Observed D_1 structures in the Corner Brook area probably embrace the D_1 and D_2 structures in the Old Mans Pond area, and D_2 in the Corner Brook map area correlates with Knight's (1994) D_3 event.

D_1 structures

D_1 structures are locally well developed in the slate and thin-bedded limestone of the Reluctant Head Formation along the Trans Canada Highway in the Humber Gorge at Duncan's Rock (g.r. 362217), at Marble Mountain (g.r. 382220), and around Wild Cove Lake (g.r. 383245). Here, S_1 forms the main foliation, defined by alignment of micas in the slate or by thin horizons of finer grained and flattened carbonate grains in the limestone beds (Fig. 13A, B). Locally S_1 can be virtually obliterated by S_2 overprinting, and S_1 is generally cryptic in the massive carbonates. F_1 folds are tight to isoclinal with an axial planar slaty cleavage (Fig. 25A; Fig. 26, sketch B6). The best examples of F_1 folds are in the predominantly slate outcrop along the Trans Canada Highway at Duncan's Rock. The metre-size F_1 folds are west vergent, have a well developed axial planar, slaty cleavage, are open to tight, and have rounded hinges. In most outcrops S_1 is subparallel to bedding (Fig. 13A). Near Meadows Pond (g.r. 334149) and along Lady Slipper road in the southern part of the Carbonate Belt, an S_1 foliation is visible in some thick-bedded units, parallel to bedding, and consists of a foliation defined by flattened carbonate grains in the thin-bedded rocks (Fig. 25B) and a fine-spaced fracture cleavage or a spaced pressure solution ("stylo") cleavage in the more massive carbonates (Fig. 25C). There is insufficient information to reconstruct large-scale D_1 structures, but available data are consistent with an overall westerly vergence.

Figure 24. Normal fault in road outcrop of thick-bedded sandstones and slates of the Irishtown Formation at the Curling turnoff near Corner Brook (g.r. 272223). View looking southwest. GSC 1995-149UUU

South of Meadows Pond, slate and minor thin-bedded carbonate of the Pinchgut Formation overlie massive carbonates of the Table Head and St. George groups. The contact may be the original or reworked thrust contact between the Humber Arm Allochthon and the autochthonous carbonate sequence. The thrust boundary is not exposed, but it traces the shape of F_2 folds in the underlying carbonates and is therefore assumed to be a D_1 structure. A second inferred thrust contact in the footwall farther east in the belt, is folded by F_2 and could be either a D_1 structure or a D_2 back thrust, related to the east vergent F_2 folding.

D_2 structures

D_2 in the map area is mainly expressed as upright folding on the scale of centimetres to kilometres. The Mount Patricia syncline north of Humber River (g.r. 355230; Fig. 25D), which is visible from the Trans Canada Highway, and the associated Meadows Pond anticline are well-exposed large-scale F_2 structures. The Trans Canada Highway around Corner Brook cuts several times through this anticline and syncline, and they are well exposed in the outcrops between the Humber Gorge and Maple Valley Industrial Park. The hinge of the Meadows Pond antiform exposed along the highway west of the Humber Gorge (g.r. 344225) is faulted, as is the hinge of the Mount Patricia syncline exposed along the highway south of Massey Drive (g.r. 339194). Slickensides on bedding planes in the forelimb are perpendicular to the large-scale F_2 fold axis and indicate a reverse sense of shear, suggesting bedding-parallel slip on the forelimb during F_2 folding. On outcrop scale, the massive and thick-bedded carbonates in the north locally show a weakly developed, steep S_2 fracture cleavage which is approximately parallel to the axial plane of the large structures. No outcrop-scale parasitic folds were observed in the northern part of the belt.

In outcrops of the Reluctant Head Formation, in Humber Gorge and near Wild Cove Lake, mesoscopic F_2 folds and an S_2 axial planar cleavage are well developed. F_2 folds have

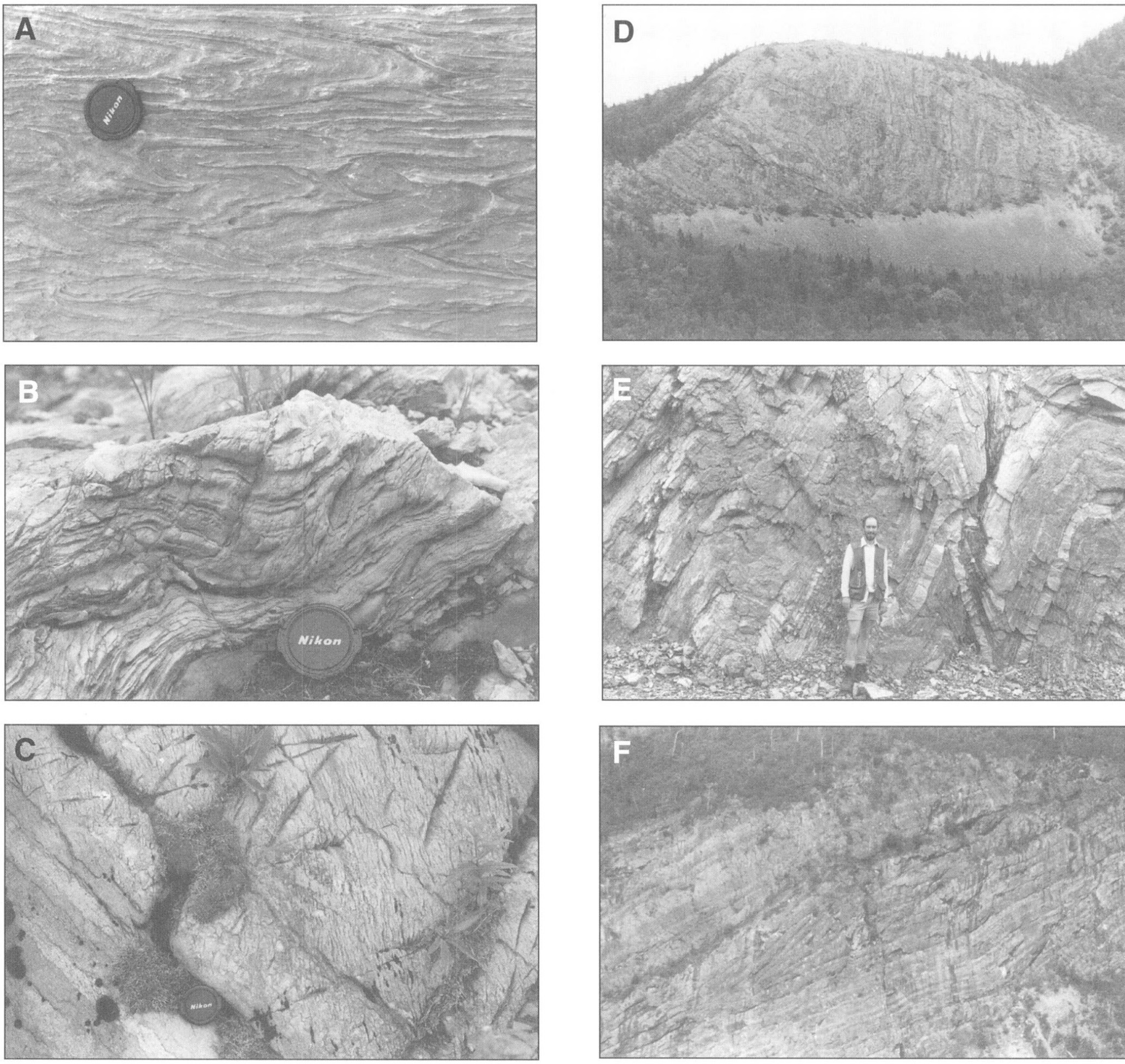


Figure 25. Structures in the Carbonate Belt. ***A****) Oblique cut through F_1 folds in thin bedded limestone of the Reluctant Head Formation along the Trans Canada Highway near Marble Mountain (g.r. 381220). The S_1 axial planar cleavage is well developed in the slaty parts of the rocks, but it is not clearly visible in the photo. View looking south. GSC 1995-149PP* ***B****) F_2 fold in thin-bedded carbonate of the St. George Group on Lady Slipper Road (g.r. 337152). An S_1 foliation is parallel to bedding. The S_2 axial planar cleavage is a spaced fracture cleavage. View looking north. GSC 1995-149SS* ***C****) Foliation in thick-bedded limestone along Lady Slipper Road (g.r. 339154). S_1 is a pressure solution (stylo) cleavage parallel to bedding. The steeply dipping, spaced fracture cleavage is S_2. View looking north. GSC 1995-149WW* ***D****) Mount Patricia syncline (F_2) in thick-bedded carbonate of the St. George Group at the western end of the Humber Gorge (g.r. 355230). View looking north. GSC 1995-149TTT* ***E****) Set of west-vergent F_2 folds in thin-bedded limestone of the Reluctant Head Formation in the Humber Gorge (g.r.368214). View looking south. GSC 1995-149WWW* ***F****) Duplex in the massive carbonate of the St. George Group, north of Wild Cove Brook (g.r. 372256). View looking north-northeast. GSC 1995-149AAAA*

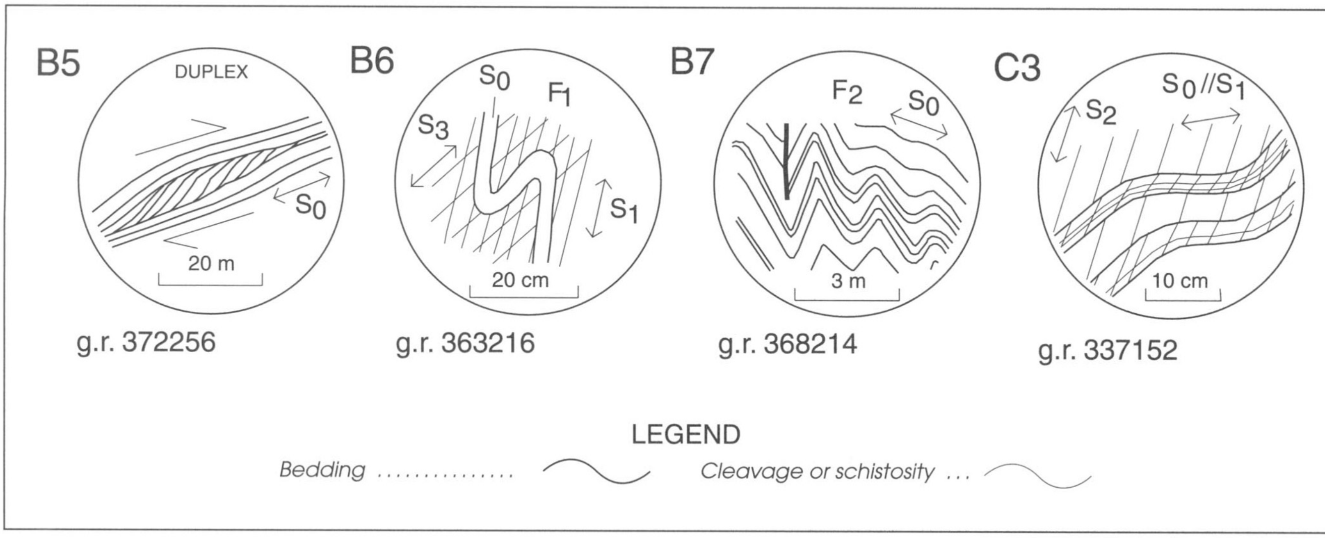


Figure 26. Sketches of structural relations in outcrops in the Carbonate Belt. Letters and numbers refer to the cross-sections (Map 1893A, in pocket) and their relative positions in the sections from which they were compiled. All sketches are oriented with the west to the left.

generally angular hinges and a similar fold style. Figure 25E shows a set of metre-scale F_2 folds in thin-bedded limestones in the core of the eastern F_2 anticline along the Trans Canada Highway in the Humber Gorge (Fig. 26, sketch B7). The axial planar cleavage is not well-developed in this outcrop and consists of a widely spaced fracture cleavage. Cleavage folds are locally abundant in more slaty parts of the Reluctant Head Formation. S_2 is a slaty cleavage or a very fine crenulation cleavage.

In the southern end of the Carbonate Belt, along the Lady Slipper Road system, D_2 structures are more ductile in the thick-bedded carbonates of the Port au Port, St. George, and Table Head groups. Many outcrops in the southern part of the belt show a penetrative S_2 fabric, which is generally developed as a fracture cleavage (Fig. 25C), but in zones of high strain it occurs as a schistosity of flattened grains or of narrow zones of intense grain-size reduction. On outcrop-scale, parasitic F_2 folds occur locally as open folds (Fig. 25B).

In the massive carbonates only one thrust-related structure was recognized. An inferred intraformational duplex (Fig. 25F; Fig. 26, sketch B5) occurs in a rock face north of Wild Cove Brook (g.r. 372256). It indicates a component of east-directed thrusting in the carbonate rocks. In combination with the east-vergent nature of the F_2 folds and the observations of east-directed thrusts and folds by Knight (1994) to the north in the Old Mans Pond region, this duplex is interpreted as a D_2 structure.

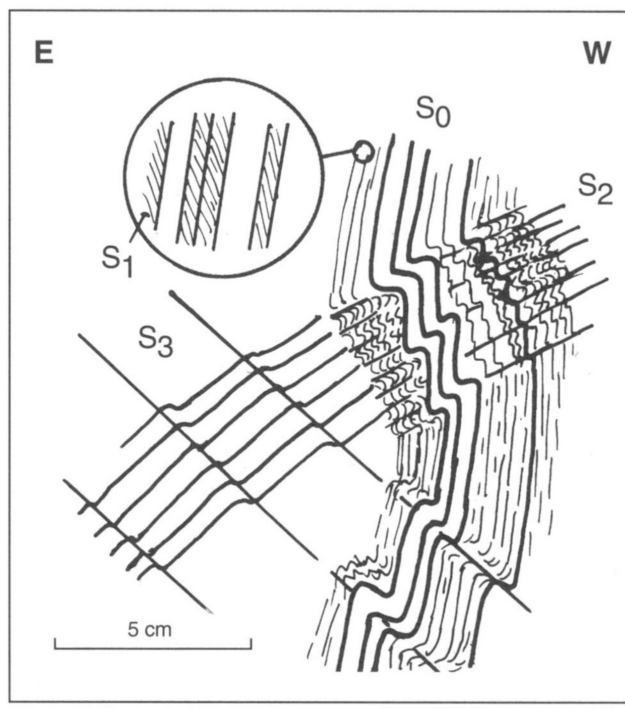


Figure 27. Schematic overprinting relationships in slate of the Reluctant Head Formation near Wild Cove Lake (g.r. 383245). Angular relationships indicate that D_1 and D_3 structures are west vergent, while D_2 structures are east vergent.

D_3 and younger structures

The youngest penetrative structures in the Carbonate Belt resemble the D_3 structures in the Humber Arm Allochthon. In slates of the Reluctant Head Formation, S_3 occurs as a spaced (2-3 cm) fracture cleavage, which locally crenulates the older cleavages (Fig. 27). Asymmetry of the crenulations is consistent and indicates a westward vergence.

In the northern part of the Carbonate Belt, late, high-angle faults are present in the massive and thick-bedded carbonate rocks exposed in Humber Gorge. They are predominantly east-west striking, and slickensides indicate a strike-slip dextral displacement. Locally the faults are associated with a bright red fault breccia, or with chimneys filled with boulders in a red matrix. These may be Carboniferous features, as hematite staining of early Paleozoic carbonates is characteristic of Carboniferous tectonic activity elsewhere within the Newfoundland Humber Zone.

Orientation data

The structural orientation data in Figure 28 show a great circle distribution of the poles to bedding with a dominant west-dipping orientation, which represents the dominant long limb of the east-vergent F_2 folds. The pole to the great circle indicates a southern plunge of the structures, which confirms the map pattern. Poles to S_1 planes have a maximum orientation close to that of S_0, but only a partial great circle distribution. S_2 is steeper than the previous two and data are concentrated

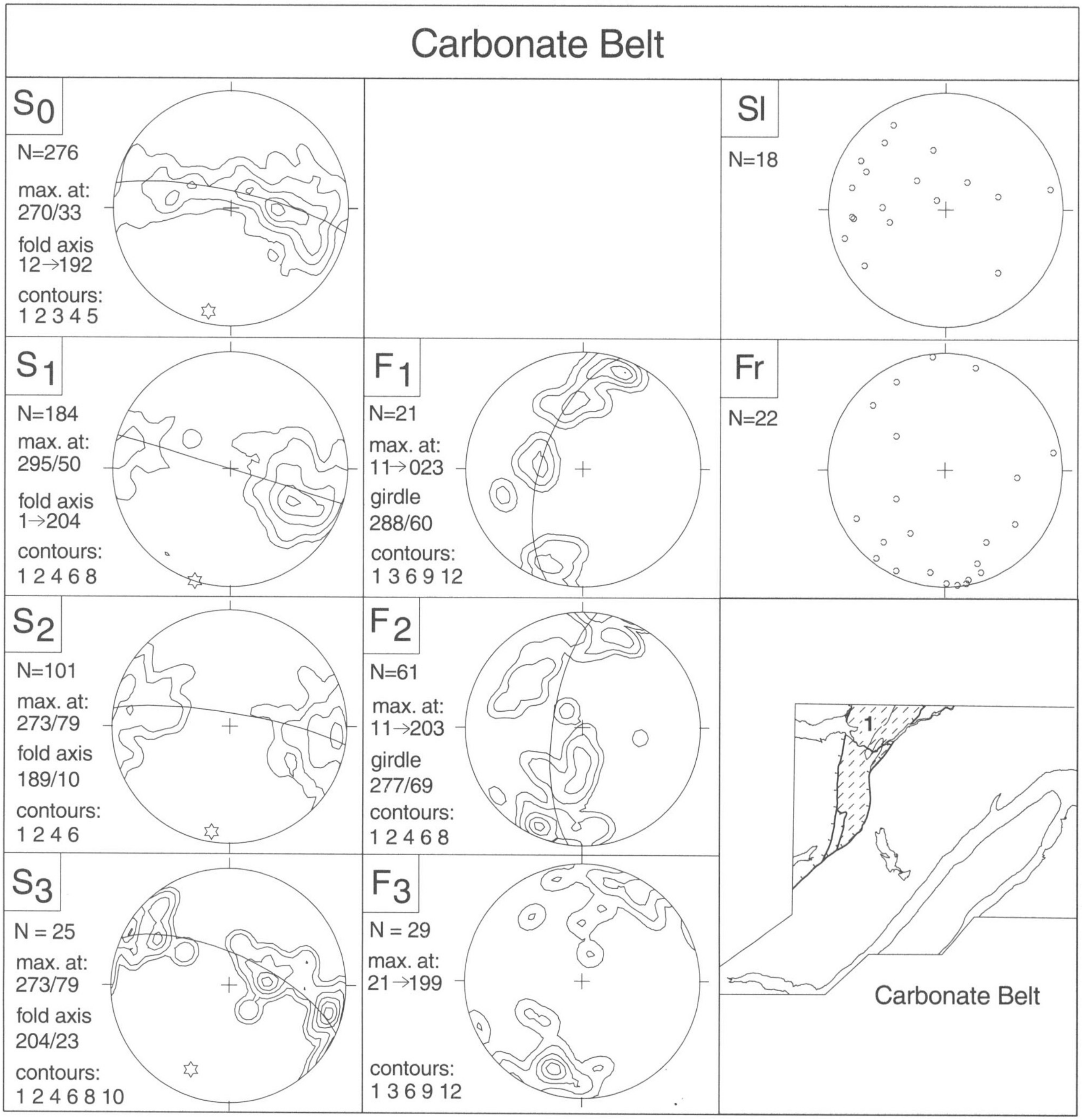


Figure 28. Orientation data for the Carbonate Belt. Linear data and poles to planar data, plotted on lower hemisphere, equal area projection. Contour levels in multiples of uniform distribution. Orientations of planes are quoted as dip direction and dip. Fr = fractures; Sl = slickensides.

in a cluster, which is dispersed somewhat as a result of fanning or later folding of the foliation planes. S_3 planes show various clusters in one great circle. All fold axes have a partial great circle distribution as a result of overprinting of folds, and lie in a moderate to steeply west-dipping plane. Maximum concentrations of fold axis orientations are plunging shallowly to the south-southwest, which is in accord with the orientations of the π-girdles of the planar structural elements. Fractures and slickensides do not show a consistent pattern, because they represent different ages of faults which have not been separated in the plots.

Boundary between the Humber Arm Allochthon and the Carbonate Belt: the Hughes Brook fault

The Humber Arm Allochthon was originally juxtaposed against the rocks in the Carbonate Belt during Taconian, west-directed thrusting (e.g. Stevens, 1970; Cawood and Williams, 1988). The Hughes Brook fault, which presently separates these two lithotectonic units, probably reactivated or excized this contact. The fault is exposed near the turnoff from the Trans Canada Highway to Massey Drive (g.r. 329210), 500 m to the south in Corner Brook stream (g.r. 328203), and another 500 m south along the water supply pipeline (g.r. 326200). The fault plane is slightly steeper than, or parallel to, bedding in the carbonates and dips around 75 to 80° west. Kinematic data adjacent to the fault do not give an unequivocal sense of movement. Slickensides on bedding planes and faults in the carbonates near the contact, indicate a reverse, top to the east movement. It is not obvious if this reverse fault movement is related to the final juxtaposition of the allochthon and the carbonates or to an earlier, or later, episode of faulting. In an outcrop west of the Trans Canada Highway on the logging road to Big Feeder Pond (g.r. 322161) shear bands occur in the carbonates near the contact between rocks of Table Head Group and Pinchgut Formation. The shear bands indicate a top to the west movement and, in their present orientation of steeply dipping to the west-northwest, indicate extensional movement on the boundary. The ductile character of these shear bands is not in accord with the brittle character of the boundary elsewhere, and could suggest that these bands are reoriented D_1 structures related to the initial west-directed thrusting and emplacement of the Humber Arm Allochthon.

Omission of stratigraphic units across the Hughes Brook Fault, combined with its rectilinear trace, western dip, and regional relations, suggests downward movement of the western block, making it an extensional structure. The absence of part, or all, of the Table Head Group and overlying Goose Tickle Group from the top of the Carbonate Belt east of the fault, as well as the absence of mélange and the lowest stratigraphic units of the allochthon immediately west of the fault, indicate that the Hughes Brook Fault is not a primary contact related to initial juxtaposition of the two lithotectonic belts, but postdates this primary contact. F_2 deformation appears not to have affected this boundary, and juxtaposition of the two lithotectonic units is likely to postdate D_2. Therefore it is tentatively linked with the late extensional faulting in the allochthon. The similarity of the D_2 structures on either side of the Hughes Brook Fault, both in geometry and orientation, suggests that the movement on the fault is not major.

Corner Brook Lake Belt

The Corner Brook Lake Belt has a complex structural history. In the northeast it contains a kilometre-scale antiform, the Yellow Marsh antiform, which is interpreted as a hanging wall antiform of a thrust nappe (Yellow Marsh thrust nappe) emplaced over an imbricate stack of thrust sheets around Corner Brook Lake and south (Valley Lakes thrust stack). The latter forms a west-vergent, thin-skinned thrust stack which involves both basement and cover rocks. In the southwest, the Corner Brook Lake Belt is complexly folded, and includes large basement blocks and rare folded thrusts. This area is cut by a set of northeast-striking, high-angle faults that can be traced for many kilometres and probably have offsets in the order of tens to hundreds of metres.

A minimum of six generations of structures can be recognized locally in the belt. These are not penetratively developed throughout the belt, and overprinting structures are not necessarily the result of different tectonic events. Careful tracing of orientations and vergences of structures, and study of overprinting relations were used to unravel the complicated history, but in areas of scarce exposure the correlation of relative ages of structures is difficult. The area around the Marble Mountain ski resort has many well-exposed and accessible outcrops which show most of the structural features. The trail beneath the powerline between Breeches Pond and Northern Harbour, which runs approximately along cross-section B, provides a good section through the Yellow Marsh antiform and contains examples of most of the generations of structures.

Early brittle thrusts (pre-D_1)

The oldest structures recognized in the belt are discrete planar features which repeat the stratigraphy. They are brittle thrusts which early in the development of the belt caused stacking of thrust sheets in a westward vergence. In most of these thrusts no relation to ductile structures was observed. They are overprinted by the S_1 foliation and later folds. Well-exposed and accessible examples of these brittle thrusts occur on: the Lady Slipper Road 10 km southwest of Corner Brook Lake (g.r. 335983), where psammite and quartzite of the South Brook Formation overlie pelite and calc-schist of the Breeches Pond Formation (Fig. 29A); in a small creek east of Corner Brook Lake (g.r. 391098) where a thin slice of basement gneiss overlies banded marble of the Breeches Pond Formation; and on an abandoned forestry track north of Grand Lake Brook (g.r. 307000) where quartz-albite schist of the South Brook Formation overlies graphitic schist of the Breeches Pond Formation (Fig. 29B; Fig. 30, Sketch E3). In all examples the foliation does not change in intensity towards the thrust contact or show any other genetic relation to the thrust. The orientation of the foliation does not change across the thrusts, other than where rheological changes cause refraction. In the example in Figure 29B the foliation is slightly steeper than the fault in hanging wall and footwall. Furthermore, the fault and the foliation are folded by an F_3 fold. These features suggest that the faults existed before the development of S_1.

A few thrust contacts show an increase in intensity of the foliation near the thrust plane, suggesting ductile deformation along the thrust surface. This is presumably a result of later ductile thrusting, which locally may have reactivated pre-existing brittle thrusts. An example of a ductile thrust occurs in an unnamed brook draining into Corner Brook Lake from Valley Lakes (g.r. 390105), where a slice of high strain basement granitoid overlies schist of the Breeches Pond Formation.

The sense of displacement of the thrusts is not evident. Stretching lineations related to the brittle thrusting were not observed, having been obliterated by younger ductile deformation and metamorphism. The relative consistency of the thrust sheets along strike suggests that the main component of displacement was most likely at a high angle to the present strike. The overall eastern dip of the fault planes suggests that thrust movement was to the west, which is consistent with the kinematics of the subsequent ductile west-directed thrust movement.

Some thrust faults in the area show crosscutting relationships which are inconsistent with thrusting in a regular (piggy back) sequence, and they are considered to be out-of-sequence thrusts. The floor thrust of the Yellow Marsh fold nappe truncates several thrusts and stratigraphic contacts in its footwall, in a fashion that suggests that it cut through a pre-existing thrust stack. Incorporation of a slice of serpentinite of the Dunnage Zone along this contact at Matthews Brook requires multiple thrusting events rather than a single phase, regular-sequence thrust event. Around Breeches Pond several thrusts appear to be cutting down-section through the stratigraphy (see cross-section B), which would also suggest an out-of-sequence origin for these structures. The out-of-sequence thrusts are not exposed, and their timing relative to the main S_1 foliation and younger features was not determined.

Pre-D_1 foliation

The oldest foliation recognized in the field (S_1) is developed throughout the belt as a penetrative schistosity or a differentiated layering. However, detailed petrography resulted in the recognition of a fabric that is older than S_1. In Figure 31A, a photomicrograph of an S_1 differentiated layering, the dominant foliation is S_1 which crenulates an older, close-spaced schistosity. The older, pre-S_1 fabric is recognized in thin sections as a schistosity of interlayered fine-grained muscovite and either chlorite or biotite. Locally it is preserved as trails of opaque oxides, which are commonly included in garnet, albite or large muscovite porphyroblasts. This oldest fabric has not been recognized in the field, and therefore it is not represented in the numbering of the generations of structures, which is based primarily on field observations.

D_1 structures

The character of the S_1 foliation varies with lithology. In the quartz-muscovite schist of the Corner Brook Lake Belt, S_1 is a very regular, close-spaced differentiated layering in which quartz-rich layers of the order of 1 mm thickness are separated by thin muscovite-rich (+ chlorite in lower metamorphic grades) films (Fig. 9B; 31A, B). The area around the ski-slopes of Marble Mountain shows many good examples of the very regular differentiated layering in quartz-rich rocks of the South Brook Formation. In more mica-rich rock types, S_1 is a schistosity of aligned mica with less extensive differentiation (Fig. 9A). In thin section these rocks also show isolated relict fold hinges, generally outlined by trails of

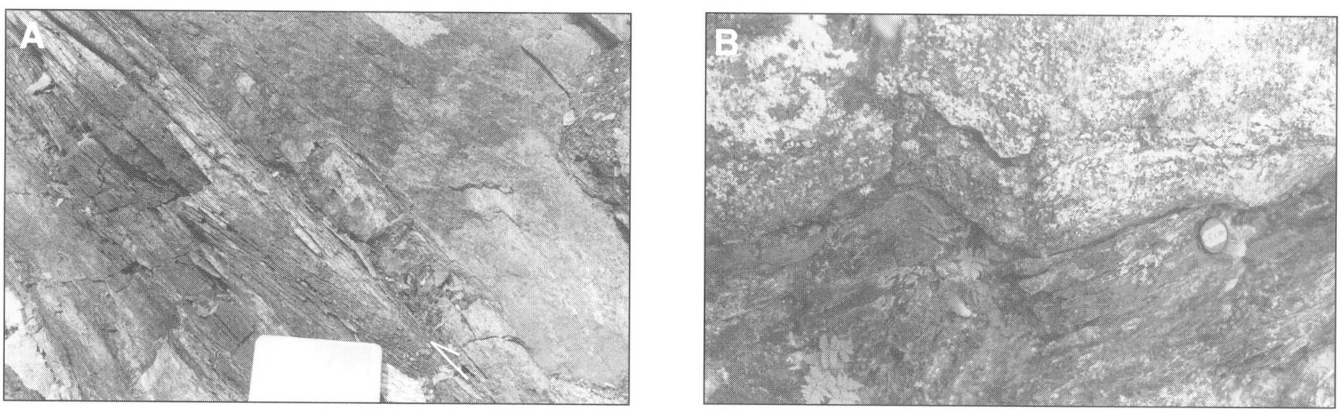


Figure 29. *Brittle thrusts overprinted by younger deformation.* ***A)*** *Thrust contact between psammite of the South Brook Formation to the right and pelite of the Breeches Pond Formation to the left. The thrust contact is indicated by the arrow to the right of the fieldbook, 10 km southwest of Corner Brook Lake on Lady Slipper Road (g.r. 335983). View looking north. GSC 1995-149SSS* ***B)*** *Thrust contact between quartz-albite schist of the South Brook Formation on top of graphitic schist of the Breeches Pond Formation. The S_1 foliation is slightly steeper than the thrust in both footwall and hanging wall. The thrust and foliation are folded by an F_3 fold. Close to abandoned forestry track north of Grand Lake Brook (g.r. 307000). View looking southeast. GSC 1995-149QQQ*

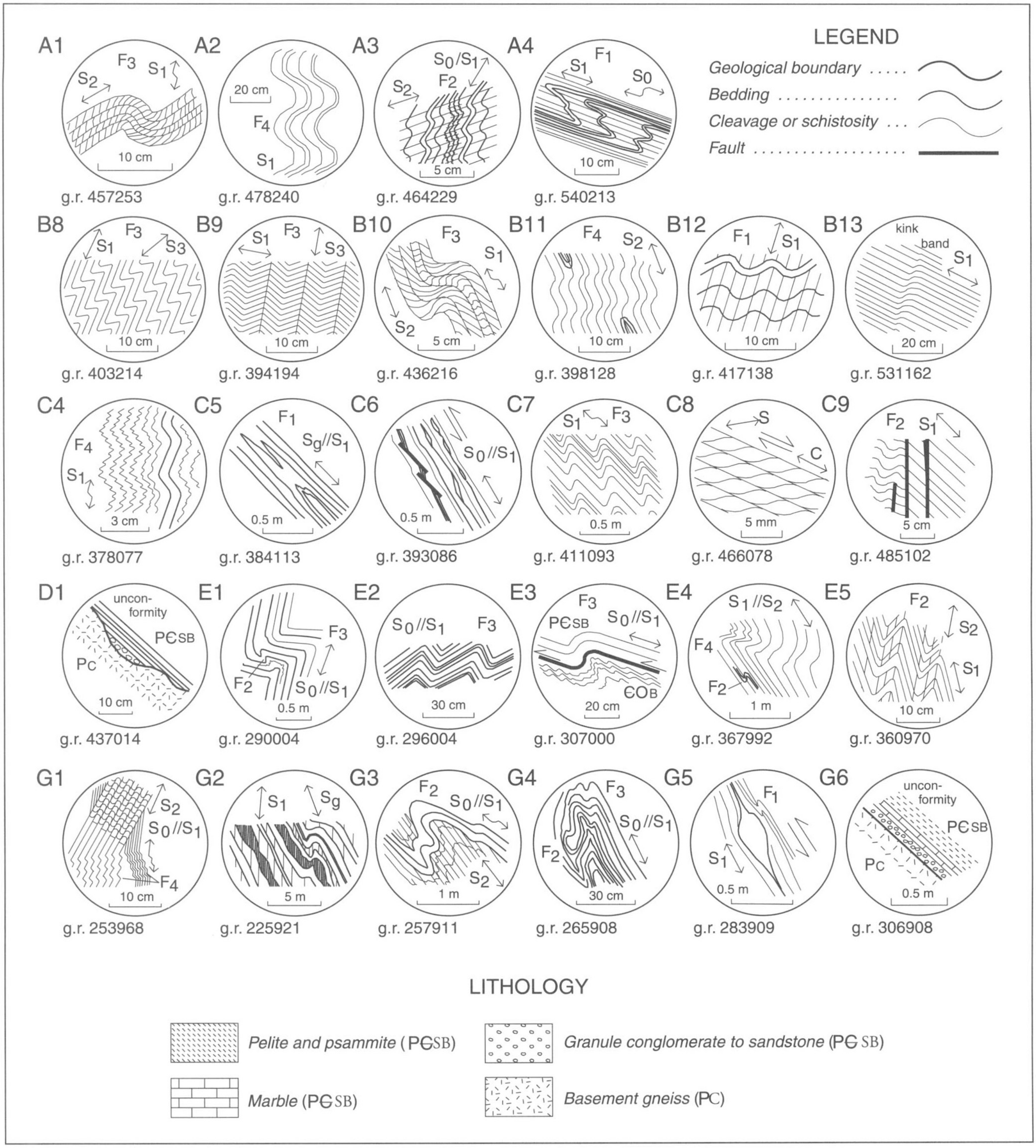


Figure 30. *Sketches of structural relations in outcrops from the Corner Brook Lake Belt. Letters and numbers refer to the cross-sections (Map 1893A, in pocket) and their relative positions in the sections from which they were compiled. All sketches are oriented with the west to the left.*

fine-grained opaque phases. In quartzofeldspathic rocks the foliation is defined by elongated quartz and/or feldspar and isolated mica with a preferred orientation. In the micaceous marble of the Breeches Pond Formation the foliation is defined by the preferred orientation of mica, and locally differentiation of carbonate minerals and mica was observed. D_1 is poorly developed in the massive marble of the Corner Brook Lake Belt and, where present, is expressed as either a weak elongation of carbonate minerals or a fracture cleavage. However, marble conglomerate in zones of high D_1 strain show a flattening of the clasts. The S_1 foliation seems consistently penetratively developed throughout the belt, but is locally overprinted by a younger foliation. Bedding and S_1 are virtually parallel on outcrop-scale in most of the map area.

S_1 is axial planar to isolated isoclinal folds, which were observed in bedding planes and thin quartz lenses (Fig. 31C; Fig. 30, sketch A4). Many of the F_1 folds are rootless, intrafolial folds, suggesting that the bedding, which is generally parallel to S_1, was transposed to the foliation plane. At very few locations, more open F_1 folds occur (Fig. 30, sketches B12, G2). No consistent vergence of the F_1 folds was determined, mainly due to a lack of sufficient observations.

D_1 mylonitic fabrics

Locally in the Corner Brook Lake Belt, S_1 has a mylonitic character, defining shear zones, particularly along thrust contacts (Fig. 30, sketches C6, C8 and G5). The main ductile

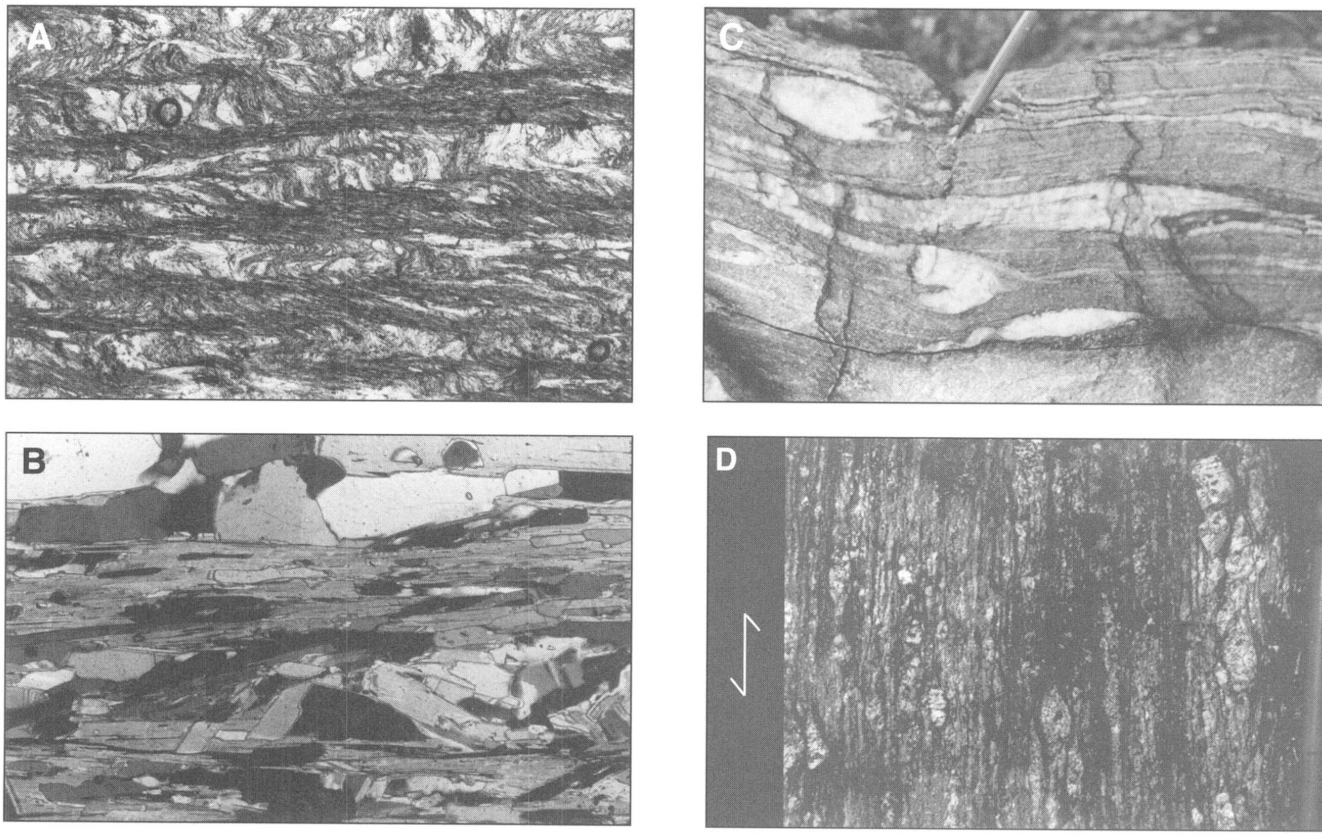


Figure 31. *D_1 structures in the Corner Brook Lake Belt.* ***A****) Photomicrograph of S_1 crenulation cleavage developed into a differentiated layering in a graphitic mica schist of the South Brook Formation (g.r. 456256). Micas in the crenulation hinges are remnants of a pre-S_1 foliation, but are totally recrystallized. The original trace of the pre-S_1 foliation is outlined by trails of graphite and extremely fine-grained micas. Plane-polarized light. Width of the photo represents 1.35 mm. GSC 1995-149JJJ* ***B****) Photomicrograph of finely spaced S_1 differentiated layering in muscovite schist of the South Brook Formation, north of Grand Lake Brook (g.r. 308992). Muscovite-rich layers alternate with quartz-rich layers. Cross-micas in the quartz-rich domains represent remnants of hinges of recrystallized F_1 crenulations. Cross-polarized light. Width of the photo is 1.35 mm. GSC 1995-149PPP* ***C****) Isoclinal F_1 folds in quartz lenses in quartz-muscovite schist of the South Brook Formation at the Marble Mountain ski slopes (g.r. 386202). The S_1 differentiated layering is axial planar to the folds. GSC 1995-149OOO* ***D****) Mylonite in a granodiorite of the Corner Brook Lake Complex, near the base of the basement thrust sheet north of Corner Brook Lake. The movement zone is vertical, due to F_4 folding. The sense of movement is east side up as shown by the domino-type rotation of porphyroclast fragments in the upper right-hand corner. View looking north (g.r. 375146). GSC 1995-149TT*

thrusts in the map area are the floor thrust of the basement sheet through Corner Brook Lake (Fig. 31D) and the floor thrust of the Yellow Marsh fold nappe. Indications of non-coaxial flow are recognized throughout the Corner Brook Lake Belt, also outside the thrusts, specifically in the Valley Lakes thrust stack (Fig. 12A).

Kinematic indicators were observed both in outcrop and in thin section. They are asymmetric, lozenge-shaped lenses of quartz, feldspar, or carbonate (Fig. 30, sketch G5), tailed porphyroclasts (Fig. 32A), C-S planes (Fig. 32B), shear bands (Fig. 12A), rotated garnets (Fig. 32C), and domino-style rotation of fractured porphyroclasts (Fig. 31D). C-S fabrics can be cryptic due to recrystallization during postkinematic metamorphism. Mylonitic features are best exposed in the Valley Lakes thrust stack , specifically in the outcrops on the eastern shore of Corner Brook Lake (Fig. 12A), in outcrops along the southwestern arm of Grand Lake (Piasecki, 1991), and along the western shore of Grand Lake. Asymmetric fabric elements at most locations indicate reverse movement (virtually dip-slip with locally a small southern pitch of L_S) of the hanging wall to the west-northwest, where the shear plane is southeast dipping. However, locally the mylonitic foliation is folded to vertical or overturned orientations, the latter resulting in apparent extensional or normal displacement.

The ductile shearing is presumably a D_1 event, because the mylonitic foliation is affected by the post-S_1 folding events (F_2 to F_4). This is consistent with relations between the foliation in the shear zones and relative timing of metamorphic mineral growth. The metamorphic peak and retrograde metamorphism, combined with local strain induced recrystallization during post-D_1 deformation, caused recrystallization of the minerals that formed the shear fabric, which made these fabrics locally difficult to recognize, and the exact extent of the shear strain is not well established. Overall, these fabrics are best developed in psammitic and gneissic rocks. The mylonitic deformation presumably affected all rocks, but the shear movement concentrated in zones, forming ductile thrusts, which locally reactivated brittle thrust planes. The ductile shearing is interpreted as a progressive development of the brittle thrusting, further building up the thrust stack.

Mineral lineations are present throughout the belt, but not everywhere well developed. In most cases they represent a true elongation lineation (L_S), but examples of mineral lineations parallel to intersections of foliation planes were also observed. Most lineations plunge either south-southeast, or north-northwest, depending on which limb of F_2 and F_3 folds they are measured on (Fig. 33, L_S in domain 1). Stretching lineations

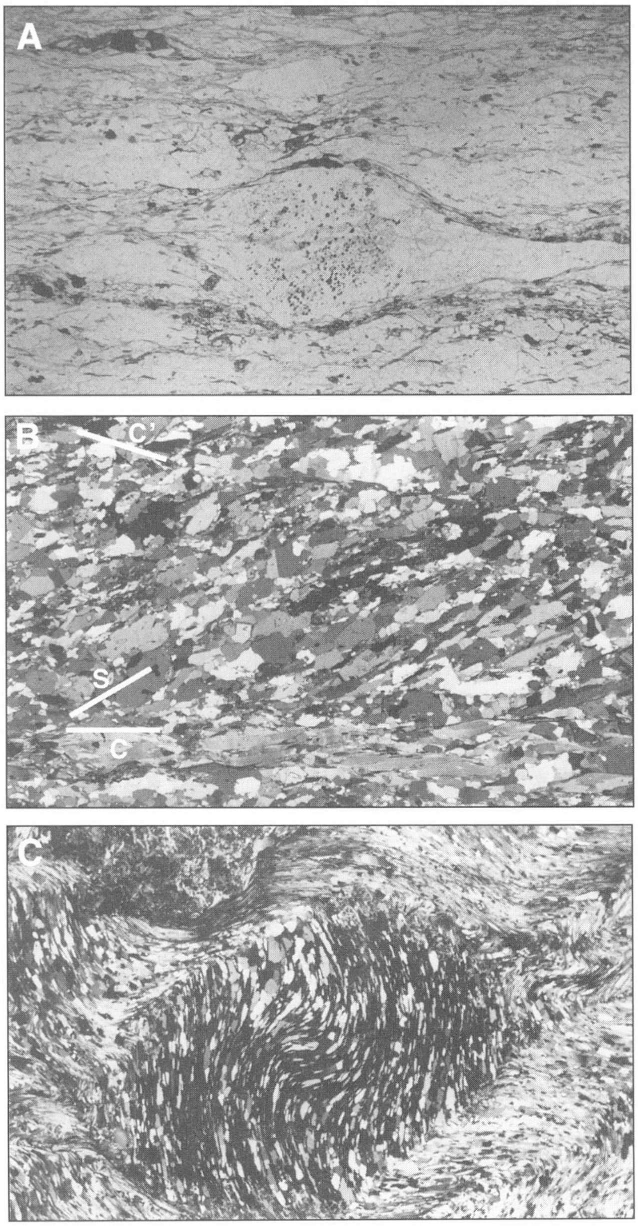


*Figure 32. Kinematic indicators in the Corner Brook Lake Belt. **A**) Asymmetrically tailed albite porphyroclast in a mylonitic albite schist of the South Brook Formation in the Humber Valley (g.r. 424241). The albite has pressure shadows and tails of recrystallized albite and quartz, indicating sinistral shear, which translates to top down to the northwest movement on a northwest-dipping foliation. The mylonite forms part of the overturned floor thrust of the Yellow Marsh fold nappe. Plane-polarized light. Width of the photo represents 5.4 mm. GSC 1995-149KKK **B**) Photomicrograph of a recrystallized S-C mylonite in gneiss of the South Brook Formation west of Northern Harbour (g.r. 531162). The orientations of C and S planes and a shear band (C') in the top of the photo are indicated. The angular relationships indicate dextral shear, which translates into top down to the southeast on an east-southeast-dipping foliation. Mica and quartz are post-kinematically recrystallized to an almost granoblastic texture. Cross-polarized light. Width of the photo represents 10.5 mm. GSC 1995-149NNN **C**) Rotated garnet in a garnet-muscovite schist of the Breeches Pond Formation south of Grand Lake Brook (g.r. 294985). The spiral-shaped inclusion trail suggests that the garnet has rotated dextrally over about 90° during growth. The matrix is subsequently crenulated in F_3 folds. Cross-polarized light. Width of the photo represents 14 mm. GSC 1995-149MMM*

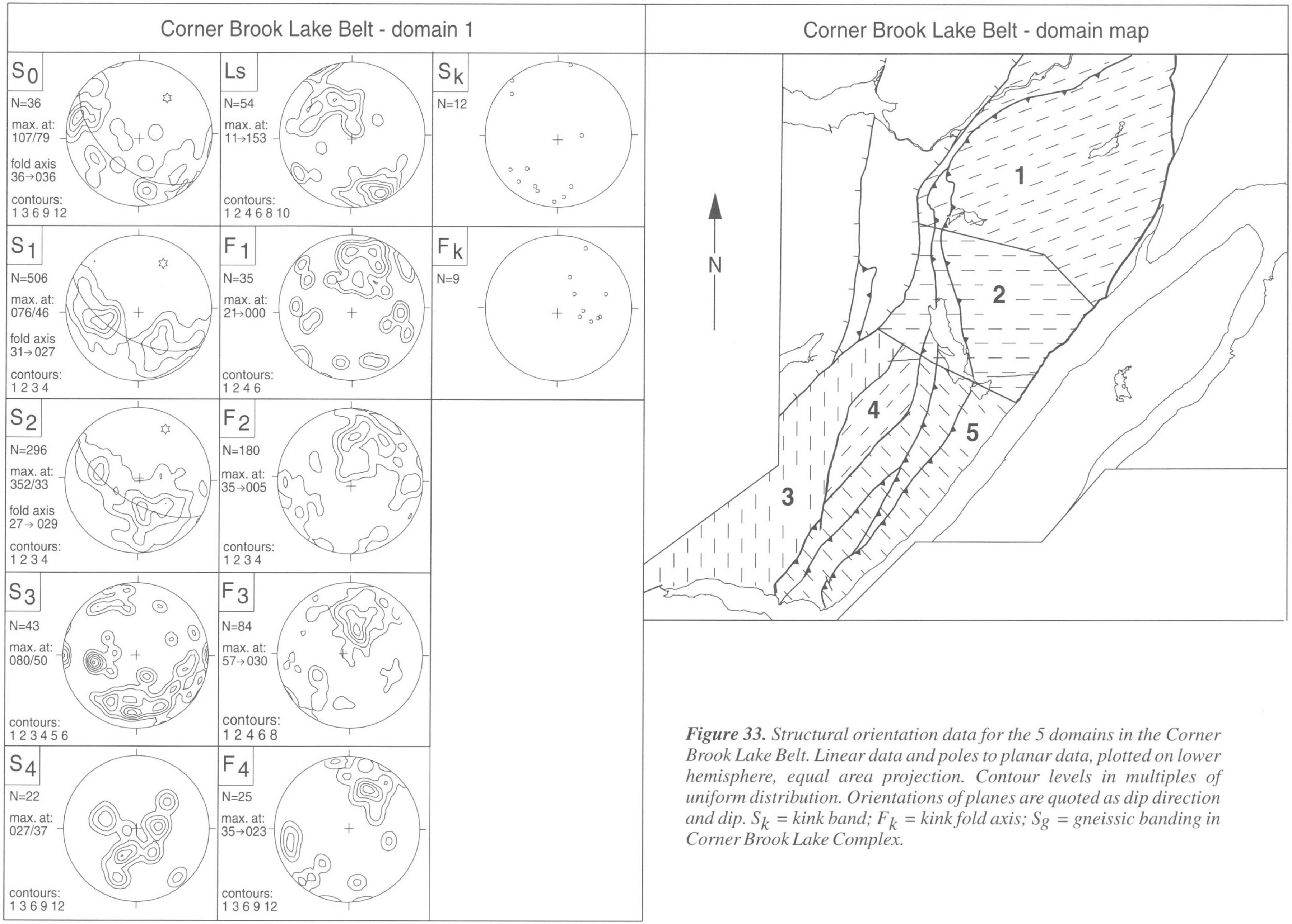


Figure 33. *Structural orientation data for the 5 domains in the Corner Brook Lake Belt. Linear data and poles to planar data, plotted on lower hemisphere, equal area projection. Contour levels in multiples of uniform distribution. Orientations of planes are quoted as dip direction and dip.* S_k *= kink band;* F_k *= kink fold axis;* S_g *= gneissic banding in Corner Brook Lake Complex.*

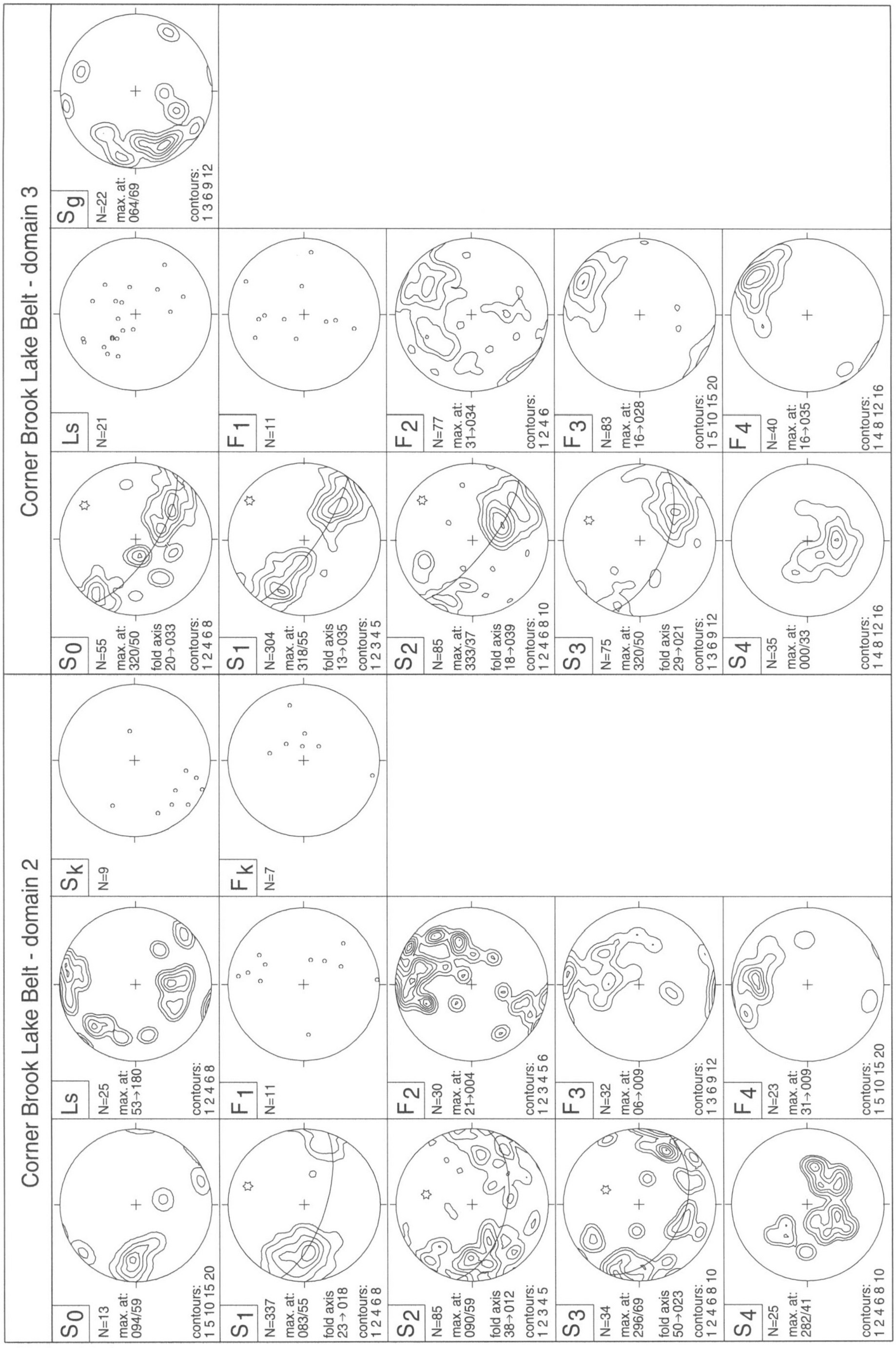


Figure 33 (cont.)

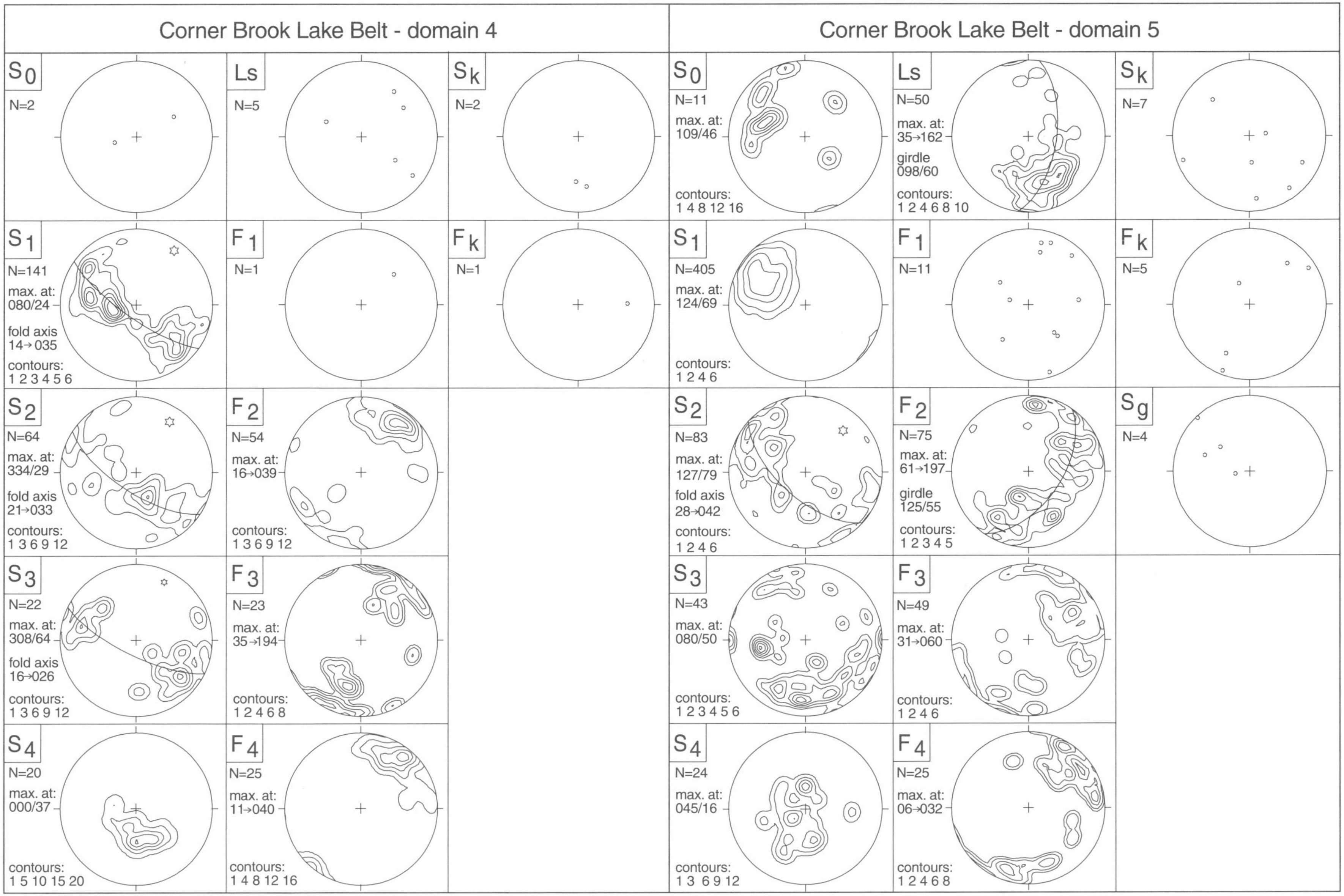


Figure 33 *(cont.)*

measured in areas with a constant southeast-dipping orientation of the foliation show a fairly constant south-southeasterly plunge (Fig. 33, L_S in domain 5).

D_1 structures in the basement rocks

The basement rocks show several styles of D_1 deformation, depending on lithology. A distinction can be made between two basement lithologies: layered gneiss and homogeneous quartzofeldspathic gneiss. The layered gneiss in the southwest of the map and in parts of the basement belt around Corner Brook Lake has a gneissosity or gneissic layering on the scale of centimetres to decimetres (Fig. 6A), and isoclinal, rootless folds are common. In the two largest basement blocks exposed along the southwestern arm of Grand Lake (g.r. 224920) mafic layers, including both layers parallel to the layering, which are commonly boudinaged, and diabase dykes of unknown age that truncate the layering, contain a schistosity that is at a small angle to the gneissosity and parallel to the regional orientation of the S_1 foliation. Towards the margins of the large blocks and in the thinner basement sheets, the S_1 foliation is more intense and parallel with the gneissosity as a result of transposition of the older structures towards the orientation of S_1. Open to tight F_1 folds in the gneissosity, with an S_1 axial planar cleavage, occur locally (Fig. 6D, 6A; Fig. 30, sketches C5 and G2).

The homogeneous quartzofeldspathic gneiss (including Lady Slipper Pluton) lacks an obvious pre-Paleozoic fabric. The fabric in the two smaller occurrences of basement rocks in the southwestern arm of Grand Lake, and in the homogeneous basement rocks east of Corner Brook Lake, is generally defined by isolated mica and elongated quartz and feldspar but with local differentiation of mica into thin, spaced (5 mm - 2 cm) cleavage planes, separated from quartz-feldspar rich domains (Fig. 6B, C; Fig 8). The homogeneous basement rocks lack marker planes to outline folds. Layering in rocks of the Lady Slipper Pluton shows very tight to isoclinal F_1 folds as well as more open younger folds.

D_2 structures

D_2 structures are variable in character. They include a penetrative schistosity, wide-spaced differentiated layering, spaced crenulation cleavage, spaced fracture cleavage, and folds ranging from isoclinal cleavage folds to chevron folds. The S_2 foliation locally forms the dominant foliation, and can completely overprint and obliterate S_1, and in such cases distinction from S_1 can be difficult. The relative ages of the foliations with respect to metamorphic porphyroblasts can locally be used to distinguish between S_1 and S_2. Where no evidence was found for the relative age of the principal foliation, it was mapped as S_1. In many areas where S_2 is penetratively developed, S_1 is still recognizable.

In the quartz-muscovite schist around Mount Musgrave, the S_2 foliation is a wide-spaced, irregular, differentiated cleavage that can locally be recognized as a crenulation cleavage (Fig. 34A, B; Fig. 30, sketches A1, B10, G1). Isolated kink-like bands, in which S_1 is folded in rounded rather than angular hinges, and which are bounded by mica-rich planes, are also typical S_2 features (Fig 34B, top-right corner). Less commonly S_2 develops as a wide-spaced fracture cleavage, in which the fracture planes are very thin muscovite-rich bands (Fig. 30, sketches A3, E5). This type of cleavage is axial planar to chevron folds. In zones of low D_2 strain, S_2 is absent or can be detected as a crosscutting foliation defined by oriented mica. These structures are well exposed near the top of the southern of the two major ski-lifts at the Marble Mountain ski resort (g.r. 388201) and along the powerlines that trend east from Breeches Pond.

Where S_2 forms the dominant fabric in mica-rich rocks it is a penetrative schistosity or closely spaced crenulation cleavage, with quartz stringers that are a product of differentiation. This schistosity is locally axial planar in isoclinal to tight folds (Fig. 30, sketches B11, E4, G3). In calc-schist S_2 is a schistosity, with slight differentiation of the carbonates and quartz from the micas. With increasing carbonate content of the rocks, S_2 is progressively weaker developed and tends to become a fracture cleavage. The overall intensity of S_2 varies with lithology and strain and can, within small areas (or even within an outcrop), vary from dominant foliation to nonexistent (Fig. 34A). No mappable areas of dominant S_2 were delineated, but is seems overall better developed in the psammite of the South Brook Formation in the Yellow Marsh fold nappe and in the schist in the southwest of the Corner Brook Lake Belt, whereas S_2 is rarely developed in rocks of the basement or in the massive marbles. Possibly, in locations where S_2 could not be identified, the D_2 caused coaxial overprinting of S_1 instead of forming a new fabric.

In thin section, S_2 is a crenulation cleavage showing better preserved hinges than S_1. The hinges usually form polygonal arcs in which the originally folded or bent mica is recrystallized and now outlines the microscopic folds in short straight segments (Fig. 34C). Recrystallization and transposition are not as progressive as in the S_1 foliation.

Locally in quartzofeldspathic rocks, S_2 can be recognized as a mylonitic foliation. Since the relative age of the foliation is hard to determine, there are only a few isolated outcrops in which the D_2 age of shear deformation can be proven. Relationships with porphyroblasts or younger folds were mainly used as criteria, as well as geometrical relationships of shear zones in the map pattern. The eastern part of the Yellow Marsh fold nappe is the only area where S-C mylonites fairly consistently yielded D_2 ages. This relative age was determined from 1) the foliation wrapping around albite porphyroblasts that locally had S_1 or F_1 inclusion trails, and 2) the shear foliation was related to an S_2 crenulation cleavage which was axial planar to folds that folded an S_1 differentiated layering. In the field, this area is characterized by two cleavages, intersecting at an angle of 20° to 45°, and locally cut by a third, weaker foliation at a very low angle. Detailed petrography showed that locally the two main cleavages represent S and C planes, with the third foliation formed by shear bands (Fig. 35A; c.f. Fig. 32B). At other locations, the dominant foliation is an asymmetric crenulation cleavage (S2), with the weaker, oblique foliation representing the older crenulated cleavage (S1; Fig. 35B). In these samples no third foliation is developed. Because the metamorphic grade was still high during and after D_2, the minerals that delineate these structures were recrystallized, which hampered the distinction between C-S fabrics and crenulations. Criteria used for the identification were: 1) occurrence of the third oblique foliation, representing shear bands; 2) angles between the two foliations that exceeded 45°, which is unlikely for C-S fabrics; 3) alternating asymmetries, suggesting parasitic F_2 folds; 4) inclusion trails of straight S_1 foliations inside albite porphyroblasts, which were parallel with the weaker, oblique foliation in the matrix, whereas the main foliation did not occur as inclusion trails, suggesting that the two are of different age, rather than concurrent S and C planes. Presumably lenses or bands with predominant F_2 crenulation and folding are surrounded by, or alternate with, zones of S-C mylonites. These structures are well exposed along the powerline, west of Northern Harbour and along the western shore of Grand Lake.

The sense of movement, determined from the angular relationship between S and C planes and asymmetric pressure shadows around porphyroblasts, indicate either reverse, top up to the northwest or oblique-normal, top down to the south-southeast, displacement. Indicators for oblique-normal displacement, with a dextral component, are far in the majority. Presently it has not been established whether the structures indicating reverse movement are relicts of D_1 shear, whether there were multiple stages and directions of D_2 shear, or whether these represent misinterpreted crenulations. The normal displacement inferred for the D_2 shearing in the eastern

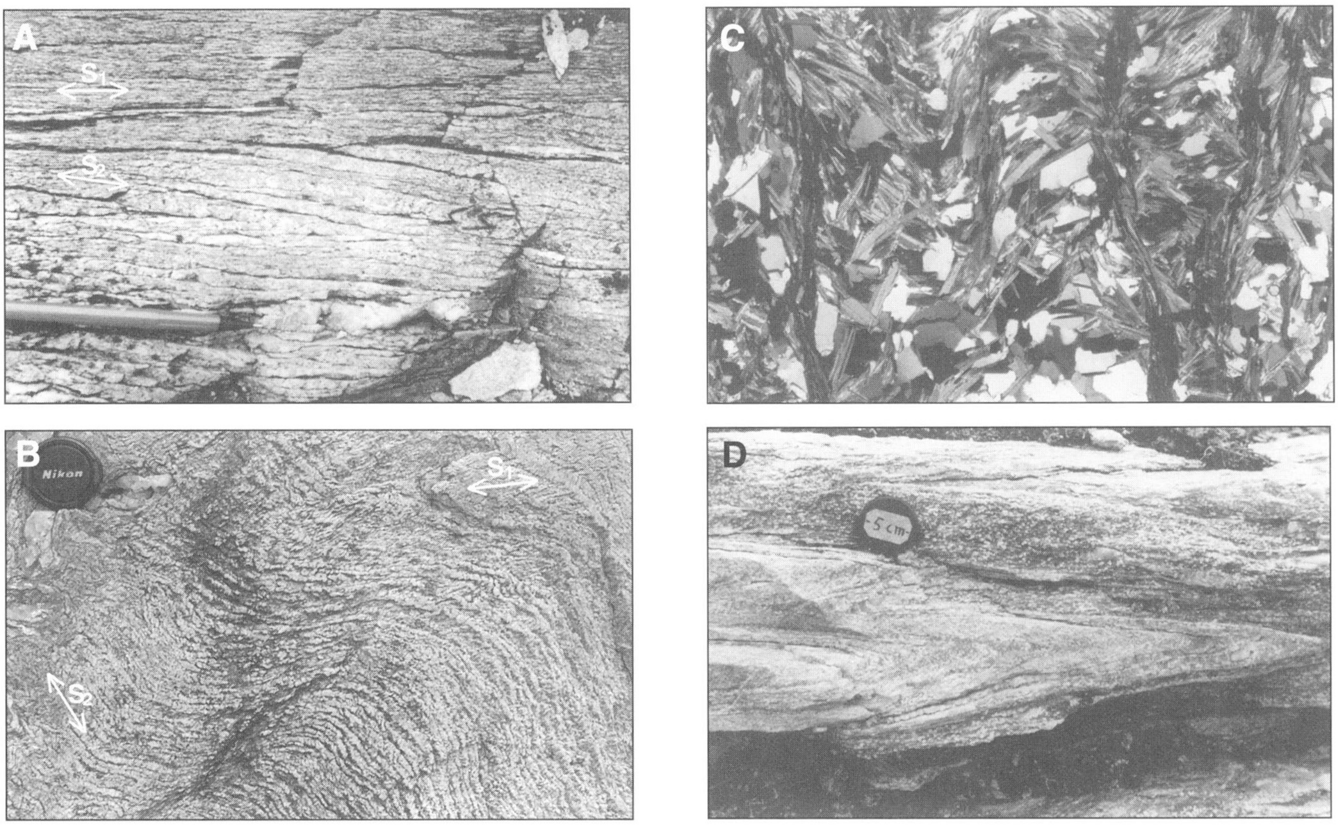


Figure 34. *D_2 structures in the South Brook Formation.* ***A****) Intersecting S_1 and S_2 foliations in quartz-muscovite schist south of Eastern Lake (g.r. 389141). The S_1 foliation in the top 1/3rd of the photo, is overprinted by S_2 in the lower 2/3rds of the photo. The S_1 foliation is close-spaced and regular, whereas the S_2, which is slightly oblique to S_1 and dips to the right, is wider spaced (0.5-1 cm) and more irregular. In parts of the leucosome about 5 to 8 cm above the pencil, S_1 is asymmetrically crenulated, in west-vergent crenulations. They have been enhanced with a marker pen on the right-hand side. View looking north. GSC 1995-149GGG* ***B****) Differentiated S_2 crenulation cleavage, in quartz-muscovite schist north of Steady Brook. In the top-right-hand corner S_2 consists of isolated bands which are wider spaced. S_2 is folded in an F_3 fold. (g.r. 434212). GSC 1995-149HHH* ***C****) Photomicrograph of an S_2 crenulation cleavage in mica schist north of Steady Brook Lake (g.r. 488234). It shows the folded S_1 as polygonal arcs and cross micas in the hinges of the crenulations. Cross-polarized light. Width of the photo represents 4.4 mm. GSC 1995-149CCCC* ***D****) F_2 fold in banded quartz-albite schist west of Northern Harbour (g.r. 517163). In the core of the fold the finely spaced S_1 differentiated layering can be recognized as the folded fabric. In the quartz-rich core of the fold no axial planar cleavage is developed, but in the mica-rich parts, S_2 forms a weak schistosity. GSC 1995-149VV*

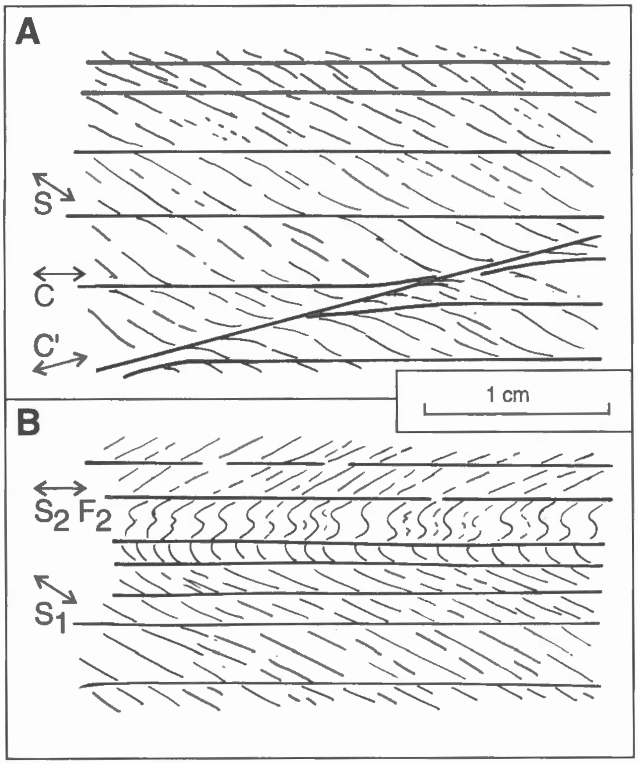


Figure 35. Sketches of intersecting cleavages formed by S-C mylonites (A), and asymmetric crenulation cleavage (B).

part of the belt is in accord with observations by Piasecki (1991) who reported apparent normal displacement in sheared rocks in the eastern part of the southwestern arm of Grand Lake. In the remainder of the area, D_2 shearing was consistent with northwest-directed thrusting.

F_2 folds are common and best developed in the quartz-muscovite schist of the South Brook Formation in the northern part of the belt, and in all schists in the southern part of the belt. They vary in style from isolated isoclinal folds (Fig. 34D; Fig. 30, sketches B11, E4, G4) to more open folds or fold sets (Fig. 30, sketches A3, E1) and trains of chevron folds (Fig. 30, sketches A3, E5). They are usually characterized by a well developed axial plane cleavage (Fig. 30, sketch G3), but also F_2 folds with a weak, widely spaced fracture cleavage (Fig. 30, sketches A3, E5) or no axial planar cleavage (Fig. 34D; Fig. 30, sketch G4) were observed. Some of these fold styles, specifically the chevron folds and other folds lacking an axial planar cleavage, are very similar to F_3 folds and can only be distinguished using overprinting relationships. A west-northwest-vergence of F_2 folds is dominant, but the asymmetry of the F_2 folds is not constant. No systematic variation of fold vergence was observed and no regional F_2 structures were mapped.

The orientation of the F_2 folds is shown in Figure 33. Highest concentrations of F_2 fold axes are in a shallow north-northeast-plunging orientation, but in most domains there is a wide spread of orientations. Only in the southeastern domain 5, where the bedding and foliations have a fairly constant southeastern dip and where there is a low intensity of later F_3 and F_4 folding, do F_2 fold axes define a great circle corresponding with the average orientation of the S_2 foliation. This distribution is a result of the overprinting of F_2 on strata already folded by F_1. In all other domains the F_2 axes show a spread that results from a combination of overprinting of F_2 over older structures and later folding of the axes by F_3 and F_4. Poles to S_2 foliations and axial planes consistently form concentrations in a steeply south-southwest-dipping π-girdle. Great circle distributions are most complete in the northern and central domains of the Corner Brook Lake Belt (domains 1 and 3 in Fig. 33). The distribution depends on the intensity of F_3 and F_4 folding and the location of the data on the larger structures (from west-dipping or east-dipping limb or from both limbs). In the western part of the Corner Brook Lake Belt moderate western dips are dominant; in the eastern part of the belt, moderate eastern dips are dominant.

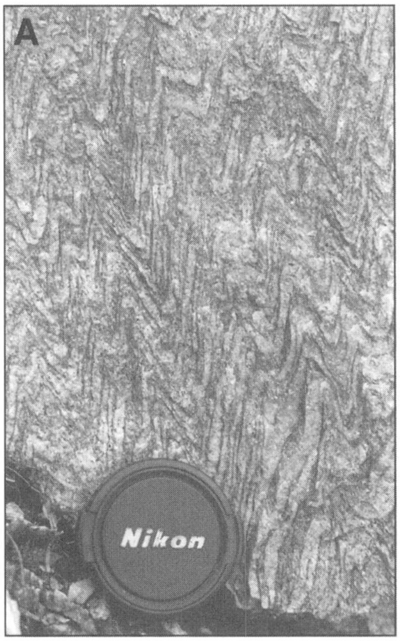
D_3 structures

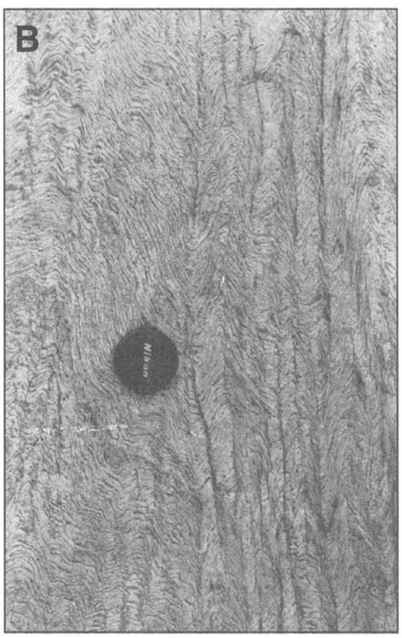
F_3 folds are the dominant expression of the third major deformation event. Most commonly they form chevron type folds without axial planar cleavage (Fig. 36A). F_3 folds vary in size from centimetre-scale crenulations to metre-size folds, and map-scale F_3 folds are also common (e.g. around Breeches Pond and in the southeastern part of the area). The chevron fold geometry is fairly consistent in all lithologies where they are developed (Fig. 30, sketches A1, B8, B10, C7, E1, E2). Fold limbs tend to be straight and interlimb angles are predominantly between 60° and 90°. In psammite, F_3 folds locally have a fracture cleavage (Fig. 30, sketch B9) or crenulation cleavage as axial planar fabric, and can look very similar to F_2 structures (Fig. 36B). In mica schist and psammite with a higher than average mica content, a vague schistosity of isolated oriented mica can occur.

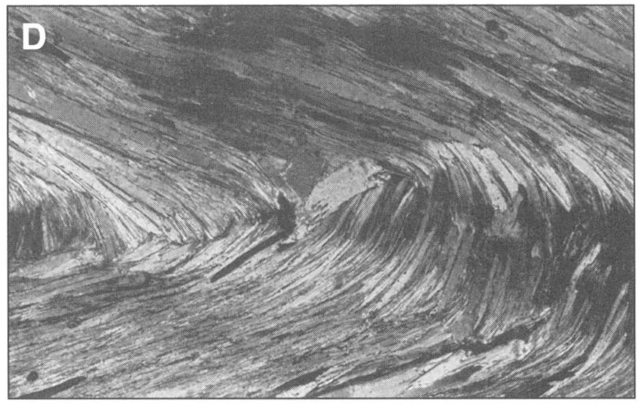
F_3 folds are most intensely developed near the western boundary of the Corner Brook Lake Belt. D_3 strain diminishes towards the east and F_3 folds are uncommon in the central and eastern part of the Yellow Marsh antiform and its southern extension, and in the Valley Lakes thrust stack. This may be an indication that the F_3 structures are related to the formation of the western boundary of the belt.

The overall vergence of the F_3 structures in the Corner Brook Lake Belt is to the west-northwest. On a smaller scale, the F_3 vergence varies, because large-scale F_3 folds exist particularly around Breeches Pond/Mount Musgrave (cross-section B) and in the southwestern part of the area (cross-section E). In the area around Grand Lake Brook, the hinge of a large-scale F_3 antiform is cut out by a late, high-angle fault, which juxtaposes two limbs of the antiform. The associated synform is exposed to the west in isolated outcrops along the east-west section of Grand Lake Brook. The map-scale F_3 folds are overturned to the east with a moderately to steeply west-dipping axial plane and a shallow plunge to the north-northeast (Fig. 33).

Overall, D_2 and D_3 structures are similar and are easily confused if overprinting relationships (as in Fig. 34B, 36C) are absent or structures cannot be correlated by orientation with those of known relative age in nearby outcrops. The







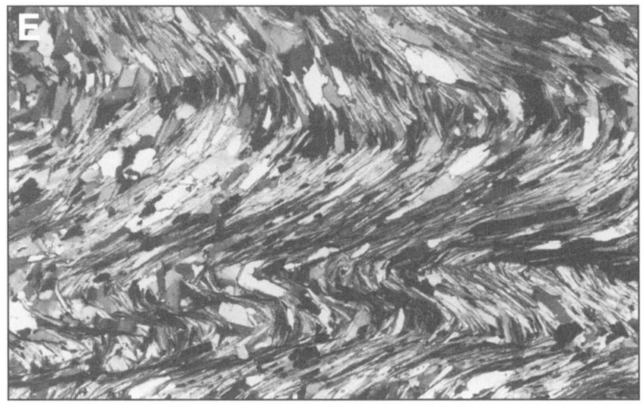


Figure 36. *D_3 structures in the Corner Brook Lake Belt.* ***A****) F_3 Chevron folds in quartz-muscovite schist of the South Brook Formation. The folds deform an S_2 foliation and are west vergent. Road outcrop beneath the new powerline on the new road east of the ski-resort. View looking south (g.r. 403214). GSC 1995-149EEE* ***B****) Wide-spaced S_3 crenulation cleavage in an area of intense F_3 folding in quartz-muscovite schist of the South Brook Formation. The folds crenulate a finely spaced S_1 differentiated layering. Looking down on gently inclined outcrop face, oblique to the fold axis. Burnover north of Steady Brook (g.r. 434212). GSC 1995-149DDD* ***C****) Overprinting relationships between S_1, S_2, and F_3 in quartz-muscovite schist of the South Brook Formation, north of Steady Brook (g.r. 434212). An F_3 fold, with "S" asymmetry (looking east), folds a centimetre-spaced S_2 crenulation cleavage. The crenulated S_1 is easily recognized and, in the short limb of the F_3 fold, it is aligned with the S_3 axial plane, and is possibly reactivated. GSC 1995-149LLL* ***D****) Photomicrograph of an F_3 fold with strain dependent recrystallization in muscovite schist of the South Brook Formation near Steady Brook Lake (g.r. 471214). In the tightly folded core on the left muscovite has grown across the hinge of the fold by strain induced recrystallization. In the outer arc of the fold, which is not as tight, the muscovites are bent with only minor recrystallization. Cross-polarized light. Width of the photo represents 8.9 mm. GSC 1995-149BBB* ***E****) Photomicrograph of F_3 crenulations in a muscovite schist of the Breeches Pond Formation south of Grand Lake Brook (g.r. 294985). Where the crenulations are tight a crenulation cleavage is formed. The folded fabric is an S_1 differentiated layering, in which the micas are recrystallized to form polygonal arcs. Cross-polarized light. Width of the photo represents 4.5 mm. GSC 1995-149AAA*

western vergence is dominant in both, and orientations of fold axes are overall similar (Fig. 33). These two generations of structures are interpreted to be genetically linked and to form subsequent stages of deformation in an evolving tectonic framework.

On microscopic scale F_3 folds or crenulations are formed by either bent mica or kinks that are partially or totally recrystallized (Fig. 36D). Different types of structures are present within a single thin section, and even within a single fold (Fig. 36D), indicating that these differences are not solely defined by metamorphic grade. The extent of recrystallization is dependent on metamorphic grade as well as strain and damage to the lattice of the mica, and varies from bent mica in open microfolds with no recrystallization, to kinked mica with slight kink band boundary migration and total overgrowth of recrystallized mica across crenulation hinges in tightly folded, higher strain, F_3 folds (Fig. 36D, E). Quartz grains show predominantly granoblastic textures indicating a total recrystallization during and after D_3. Intragrain deformation of quartz is common only in the lowest grade part of the belt in the west.

Data in Figure 33 show that the orientation of F_3 is fairly constant with a shallow to moderate north-northeastern plunge, which is similar to the F_2 orientation, but slightly less dispersed. S_3 planes are generally moderately to steeply dipping to the southeast, steeper than S_2. Data points are more concentrated in the southwest than in the remainder of the area, where orientations of S_3 are more dispersed, without forming true (statistically) great circle distributions. Comparison of relative numbers of measurements of D_3 and other structures in the southern domains shows that D_3 is less represented in the eastern domain 5. Field observations indicate that F_3 is progressively less intense from west to east in the whole belt.

In summary, D_3 structures represent a period of crustal shortening which was accomplished largely by westward vergent folding with only limited foliation development.

D_4 structures

D_4 structures are predominantly folds with gently dipping axial planes which overprint the F_3 folds and older structures (Fig. 30, sketches A2, B11, C4, E4, G1). F_4 folds are predominantly chevron type, similar folds, with fairly straight limbs, but more rounded, parallel fold geometries also exist, sometimes in the same outcrop (Fig. 37A, B). Kink band-like folds with rounded hinges and gently dipping axial planes are also attributed to D_4. Crenulations are common in the cores of larger F_4 folds, but a penetrative crenulation cleavage was not observed (Fig. 37C). An axial planar fracture cleavage is uncommon and only occurs in the more massive rock types, like the quartz-rich psammite of the South Brook Formation (Fig. 37B).

In thin section, F_4 structures look similar to F_3 microfolds, but recrystallization of mica in the cores of folds is less common. In most crenulations or microfolds the mica is kinked or bent without any trace of recrystallization (Fig. 37D), but a few kink band boundaries are irregular, suggesting minor recrystallization. In high strain parts of the folds, however, mica grows across the hinges. Quartz grains in F_4 hinges are internally strain-free in the higher metamorphic parts, but show minor strain in the western part of the belt.

Large-scale, open, recumbent, or gently dipping F_4 folds are found throughout the belt and are evident in the cross-sections. The dominant foliation fans from east-dipping to west-dipping throughout the belt, with the exception of the hinge zone of the Yellow Marsh antiform in the north, where the foliation swings around from a northwestern to a northeastern dip in a steeply inclined fold. These fans are located in the hinges of recumbent F_4 folds, which are indicated in the map where they were well defined (see map, cross-sections A, B, C, E). Since the F_4 folds plunge to the north-northeast, these structures have an effect on the map pattern. They cause the swing in the trend of the western part of the belt from northeast-trending (northwest-dipping) in the north, swinging into a north-trend near Breeches Pond (changing to east-dipping), and back to a northeast-trend near Corner Brook Lake (west-dipping near the western boundary). Towards the eastern part of the belt the F_4 effects diminish. Since the foliation is east-dipping in this low D_4-strain area, and since an eastern dip is prevalent in the more intensely folded part, the overall orientation of the foliation in the belt (with the exception of the hinge area of the Yellow Marsh antiform) is assumed to be moderately east-dipping. The shallow orientation of the axial planes suggests contraction along a steep axis, which effectively results in a flattening or collapse of the whole belt. The swing in the western boundary of the belt follows the changes in trend in the foliation in the Corner Brook Lake Belt caused by F_4 folding. This may suggest that the boundary itself is also folded by F_4, but since the fault surface is not exposed, this cannot be unequivocally established.

Orientation diagrams of D_4 structures (Fig. 33) show the fairly consistent gently to moderately south-dipping S_4 axial planes. The F_4 fold axis is about coaxial with older fold axes, plunging gently to moderately to the north-northeast. The decrease in the intensity of D_4 deformation to the east is witnessed by a relative decrease in D_4 data from the southwestern to the southeastern domain.

Kink folds and late faults

The youngest pervasively developed structures in the Corner Brook Lake Belt are brittle high-angle faults and kink folds, which are interpreted to be related to the Cabot Fault system.

The late faults in the area are poorly exposed. Small scale faults (Fig. 30, sketch, C9) occur throughout the belt and are common in the east. Where the actual fault plane is exposed, slickensides are common. No systematic preferred orientation was detected for either fault planes or slickensides (Fig. 33). Tension gashes that truncate all other structures are usually several decimetres to over a metre wide and up to 20 m long. They are filled with very coarse-grained, pure white quartz and are most common in the eastern part of the Corner Brook Lake Belt. These structures do not show a systematic preferred orientation and are also assumed to be related to the late faulting.

On a larger scale, truncations in the map pattern coincide with sharp lineaments that are visible either in the field or in air photos. The lineaments are generally straight, suggesting that they are high angle faults. They occur in many orientations, but most commonly are either parallel to the strike of the belt, east-striking or north-striking. The set of east-striking faults that offset a northeast-striking fault escarpment east of Corner Brook Lake, is visible on both the topographic map and air photos, and is the best example of these late faults. Their offset is interpreted to be between tens and hundreds of metres.

Kink bands (Fig. 30, sketch B13) appear throughout the belt but are most abundant in the east, near Grand Lake, adjacent to the Cabot Fault system. Kink bands are between 2 and 10 cm wide and are characterized by a moderate to steep north-northwest-dip, with approximately down-dip plunging kink fold axes (Fig. 33). No preferred asymmetry of the kink folds was detected.

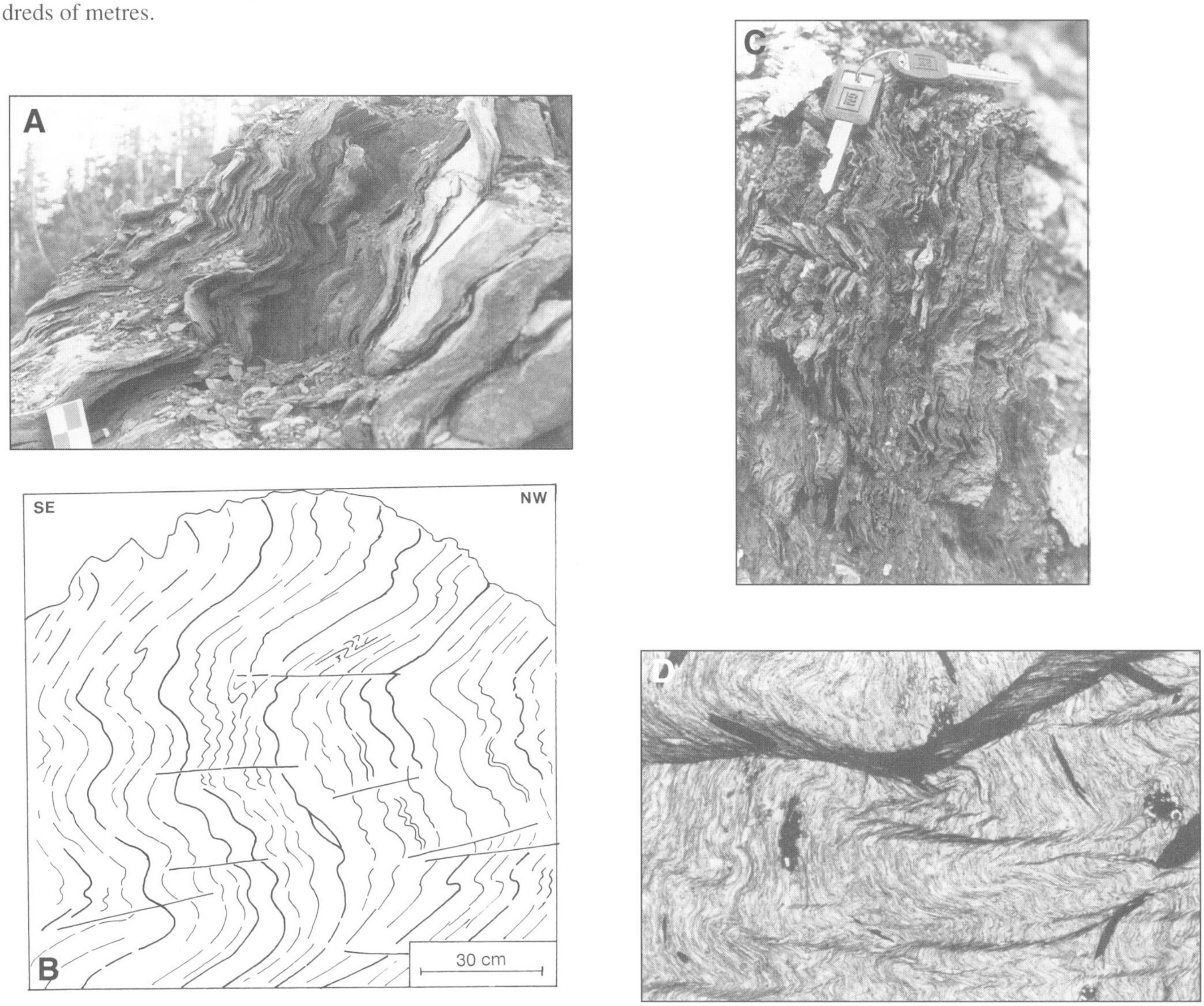


Figure 37. *D_4 structures in the Corner Brook Lake Belt.* ***A)*** *F_4 folds in calc-schist and thin-bedded marble of the Breeches Pond Formation in the southwestern part of the map area (g.r. 287950). GSC 1995-149ZZ* ***B)*** *Sketch of F_4 folds in quartz-muscovite schist of the South Brook Formation, north of Steady Brook Lake. These folds have a rounded geometry and very widely spaced fractures along the axial plane. They are located in the core of a large-scale hinge (g.r. 480236).* ***C)*** *F_4 crenulations with a subhorizontal axial plane in calc-mica-schist of the Breeches Pond Formation along Lady Slipper Road west of Corner Brook Lake (g.r. 377077). GSC 1995-149YY* ***D)*** *Photomicrograph of F_4 crenulations in a mica schist of the Breeches Pond Formation, near the bridge on Lady Slipper Road across Grand Lake Brook (g.r. 332992). No recrystallization of mica occurred in the hinges of the crenulations. Locally a fine crenulation cleavage is formed, which mesoscopically resembles a fracture cleavage. Plane-polarized light. Width of the photo represents 4.6 mm. GSC 1995-149CCC*

Valley Lakes thrust stack

Figure 10 is a detailed map of part of the Valley Lakes thrust stack east of Corner Brook Lake (see also Ryan, 1992). This area provides a good example of the tectonics of the Corner Brook Lake Belt and is relatively well exposed and easily accessible.

This part of the thrust belt consists of a stack of thin thrust sheets, which dip moderately to steeply to the east. The four thrust slices between the upper block, formed by the Yellow Marsh fold nappe, and the large basement-dominated block in the west of the detailed map, range in width from less than 100 m to about 300 m (Fig. 10). Each thrust slice contains all or part of the total stratigraphic sequence, which is extremely thin here, suggesting that these thrust sheets were sliced off a high in the basement. The sheets are laterally quite consistent and can easily be traced.

The repetition of the stratigraphic sequence and the thrust faults, which are only locally exposed, outline the thrust sheets. The three lower thrust sheets contain thin slices of basement rocks, which are no more than a few tens of metres thick, and are locally absent.

The thrusts are generally brittle features (discrete planes), but locally an intense mylonitic fabric is developed on the thrust contacts. The dominant fabric in the thrust stack (S_1) postdates the brittle thrust emplacement, but can locally be recognized as a mylonitic fabric, in which asymmetric fabric elements indicate northwest-directed non-coaxial flow and ductile thrusting, slightly oblique with respect to the eastern dip of the thrust planes. No large-scale folds are developed in the thrust stack, and outcrop-scale folds are uncommon.

Yellow Marsh fold nappe

The northeastern part of the Corner Brook Lake Belt is dominated by the Yellow Marsh antiform, which is interpreted as a hanging wall antiform in a fold nappe, formed by the large, continuous block of metaclastic rocks of the South Brook Formation in the east of Corner Brook Lake Belt. The antiform is expressed in the folding of the dominant foliation (S_1 and S_2) and the boundary between the two map units of the South Brook Formation (P€SB and P€SBa) in a gently north-northeast-plunging antiform with a half-wavelength of approximately 10 km. Orientation data of S_0, S_1, and S_2 in the northern part of the fold nappe (Fig. 33, domains 1 and 2) are all distributed on π-girdles that define a fold axis of the antiform plunging 20° to 30° towards about 025°. Cross-sections A, B, and C show that the antiform is overturned to the west. It dies out to the southwest. None of the small-scale folds have appropriate change of vergence across the antiform to be parasitic small scale folds, and this hampers the determination of the relative timing of the formation of the antiform. Since the S_2 foliation is folded in the antiform, the structure must postdate at least the main phase of D_2. D_4 structures seem to affect both limbs of the structure equally, and the orientation of S_4 axial planes is relatively constant through the antiform (Fig. 33) indicating that the antiform predates D_4. The S_3 axial planes do not appear to be affected by the folding of the antiform. Although their orientation is not constant, field observations in the core of the antiform show that S_3 axial planes here are steeply dipping, suggesting that they were not affected by the large-scale folding.

The inferred relative age of the Yellow Marsh antiform is between D_2 and D_3. A possible interpretation of the structure is that it forms part of a thrust nappe, which moved northwestwards on an out-of-sequence thrust at the base of the metaclastic rocks, cutting through, and overriding the previously formed thrust belt. The map pattern is interpreted as an oblique section through this fold nappe.

Humber Zone rocks on Glover Island

A thin slice of the Corner Brook Lake Belt, containing rocks of the Corner Brook Lake Complex and South Brook Formation, is exposed along the western shore of Glover Island (from g.r. 453035 to g.r. 411983), in the lower reaches of Keystone Brook (g.r. 429003), and in a few isolated outcrops in the hills along the western shore of the island. The structural development of these rocks is similar to the belt west of Grand Lake, but took place under lower metamorphic conditions, reaching only up to greenschist facies (Knapp, 1982). S_1 and S_2 are both crenulation cleavages, locally differentiated, and west-vergent F_2 and F_3 folds, similar to those to the west, as well as kink bands were observed. D_4 structures were not recognized.

Thrust faults within this slice, as reported by Knapp (1982), were not recognized. The existence of the large-scale F_2 syncline, based on a change in facing directions (Knapp, 1982), which is not reliable in such high strain rocks, is not supported by a change of vergence of small-scale F_2 folds.

These rocks appear in the core of the Cobble Cove antiform, in cross-section F. They are juxtaposed against rocks of the Glover Island Belt to the east by a high angle fault which is a splay of the Cabot Fault system, and they are overthrust by gabbro and minor ultramafic rock of the Grand Lake Complex to the southwest along a ductile shear zone. This latter boundary may represent a remnant of the Baie Verte Line.

Boundary between the Carbonate Belt and the Corner Brook Lake Belt: the Humber River Fault

The boundary between the platform carbonates of the Carbonate Belt and the predominately metaclastic rocks of the Corner Brook Lake Belt is formed by a steeply dipping fault, which largely follows the trend of the two belts, but locally truncates structures in both. The fault dips to the west-northwest along most of its strike, but is overturned, east dipping, between Breeches and Island ponds. This swing in the Humber River Fault parallels that in the rocks of the Corner Brook Lake Belt, where it is caused by D_4 deformation.

The structural nature of this lithotectonic boundary is enigmatic, because the fault plane is not exposed in the map area. The boundary was previously presented as a thrust, which was overturned and affected by the east-directed thrusting and high angle faulting (Williams and Cawood, 1986). The northern part of this fault was termed the Corner Brook Lake thrust (Walthier, 1949; Kennedy 1981), and its southern extension was named the Grand Lake thrust

(Kennedy, 1981). The term Corner Brook Lake thrust is a misnomer, because present data show that this fault does not reach Corner Brook Lake and there is no clear evidence for thrust movement on this structure. In the southwestern part of the area, the large basement inlier on the southwest arm of Grand Lake was interpreted to be in the hanging wall to platform carbonates along the Grand Lake thrust. This contact is, however, stratigraphic with a thin veneer of South Brook Formation (<15 m thick) unconformable on basement and conformably overlain by Breeches Pond Formation (g.r. 204932).

The Humber River Fault is considered to be an extensional structure because: 1) it dips predominantly to the northwest; 2) it forms a metamorphic break; 3) it cuts D_2 structures to the west; and 4) adjacent structures in the carbonate belt are not overturned. However, the interpretation that the fault is an overturned thrust cannot be excluded (cf. Williams et al., 1982, 1983, 1985; Williams et al., 1989b). In the northern part of the map area the fault lies adjacent to overturned thrusts in the Corner Brook Lake Belt, and the present extensional structure could partially follow the trace of, and would have reactivated, pre-existing overturned thrusts. Waldron and Milne (1991) and Knight (1992, 1994; Knight and Boyce, 1991) have described late-stage down to the west extensional movement in the vicinity of this contact in the Pasadina map area east of Old Mans Pond. In the southern part of the map area the Humber River Fault runs west of the main carbonate sequence. The occurrence of platform carbonates in both the Carbonate Belt and Corner Brook Lake Belt suggests that the two adjacent belts may have formed a continuous lithostratigraphic unit, divided into separate thrust sheets, and that final movement on the fault did not involve extensive displacement.

Glover Island Belt

Structural elements in the Glover Island Belt are steeply dipping. A regionally developed foliation is present, and locally there are multiple generations of foliations, folds, and faults. Overprinting relationships between the various structural elements show that the rocks have experienced a complex deformation history. The dominant structures of the belt are the large antiform in the west of Glover Island (Cobble Cove antiform), the faults cutting through the core of the antiform, the Kettle Pond shear zone, which separates the Grand Lake Complex from the Glover Formation, the broad warp of the foliation around the Glover Island Granodiorite, and several narrow slices of granitic and mafic rocks west of Grand Lake (see map, cross-sections A, B, C, D, F). The relatively low strain in most of the belt contrasts with the high strain rocks in the adjacent Corner Brook Lake Belt of the Humber Zone. Reconstruction of the structural development of the belt is hampered by poor exposure. Apart from the shores of Glover Island and a few brooks, outcrops are scarce and isolated, preventing tracing and detailed correlation of structures.

Main foliation – S_1

The dominant foliation (S_1) in the map area is the oldest structure recognized in most rocks, although flow banding and mineral alignment in the intrusive rocks, and sedimentary structures in the volcaniclastic rocks, constitute structures that locally existed prior to D_1. S_1 varies in style in the different rock types (cf. Knapp, 1982), but the continuity of its orientation across lithological boundaries suggests that the main foliation in all units is approximately equivalent. Folds related to the S_1 foliation are uncommon.

The greenschist and lower serpentinite body of the Grand Lake Complex (map unit OGGS) are highly foliated. The schistosity in the greenschist is defined by aligned phyllosilicates, predominantly chlorite and minor biotite, plus epidote and plagioclase. The serpentinites show aligned serpentine and talc, and locally flattened carbonate. The foliation locally forms an anastomosing pattern of a serpentine-rich matrix, surrounding lenses of less deformed and altered ultramafic material. These lenses form hills that stand out in the landscape. Gabbros in the complex are locally strongly layered (Fig. 15A) and show a weak primary mineral alignment (Knapp, 1982) that predates the regional S_1. They are cut by narrow shear zones, which vary in width from less than a centimetre to tens of metres, in which minerals are flattened and recrystallized to form a mylonitic foliation (S_1), with alteration to greenschist mineralogy. The shear zones are generally narrow and form isolated zones or anastomosing patterns, whereas the wall rock is virtually undeformed. The small scale shear zones become more ubiquitous towards the boundaries of the Grand Lake Complex and are probably related to the bounding shear zones.

The regional S_1 foliation in the Glover Formation is variably developed, but shows an overall decrease in intensity towards the east, away from the Kettle Pond shear zone. The foliation is not penetrative in all rocks of the formation. Fine-grained volcanic and volcaniclastic rocks have a well developed S_1 schistosity that is (sub-)parallel to original layering, suggesting a transposition of the layering towards the foliation. S_1 is defined by a preferred orientation of plagioclase and actinolite, locally with flattened carbonate. In the more massive and coarser grained diabase and basalt, a foliation is not always present. Finer grained phases of these rocks and narrow high strain zones generally show a strong schistosity or mylonitic foliation, defined by chlorite, flattened plagioclase (both phenocrysts and groundmass), and minor actinolite. Alteration to lower greenschist mineralogy is common throughout the belt, but is enhanced in the high strain zones. In pillow basalt S_1 is commonly defined by flattening of the pillows (Fig. 15C; Fig. 38, sketches B17 and C12), and a penetrative cleavage is not developed.

In the Glover Island Granodiorite, S_1 consists of a preferred orientation of aligned feldspar phenocrysts in a matrix of flattened quartz and isolated biotite or chlorite after biotite. The fabric is weak in the coarse-grained lithologies and in the core of the pluton, and is better developed in the medium-grained parts of the pluton and near its margins. The deformation of the pluton increased towards the east. In thin sections of granodiorite with a penetrative foliation, the plagioclase phenocrysts are aligned with no obvious internal deformation, but the crystals are intensely sericitized. The quartz and, to a lesser extent, the microcline in the matrix are deformed, and quartz is recrystallized or shows formation of subgrains and serrate grain boundaries. They form elongate clusters that define a foliation together with the chlorite, which replaces the original biotite. The foliation affects both the main granodiorite and late pegmatitic phases of the pluton (Fig. 38, sketch B16). Although the S_1 foliation in the granodiorite is approximately parallel to the S_1 foliation in surrounding rocks of the Glover Formation, some xenoliths of inferred Glover Formation contain a foliation which is truncated by, and at an angle to, the foliation in the engulfing granitoid. At one location xenoliths show a folded internal foliation, with the axial plane of this fold being approximately parallel to the foliation in the pluton (Fig. 38, sketch C11). The foliation in the xenoliths must predate the intrusion, indicating that the Glover Formation rocks were deformed before the time of intrusion. Presumably the folding of that older foliation took place after intrusion during deformation of the pluton.

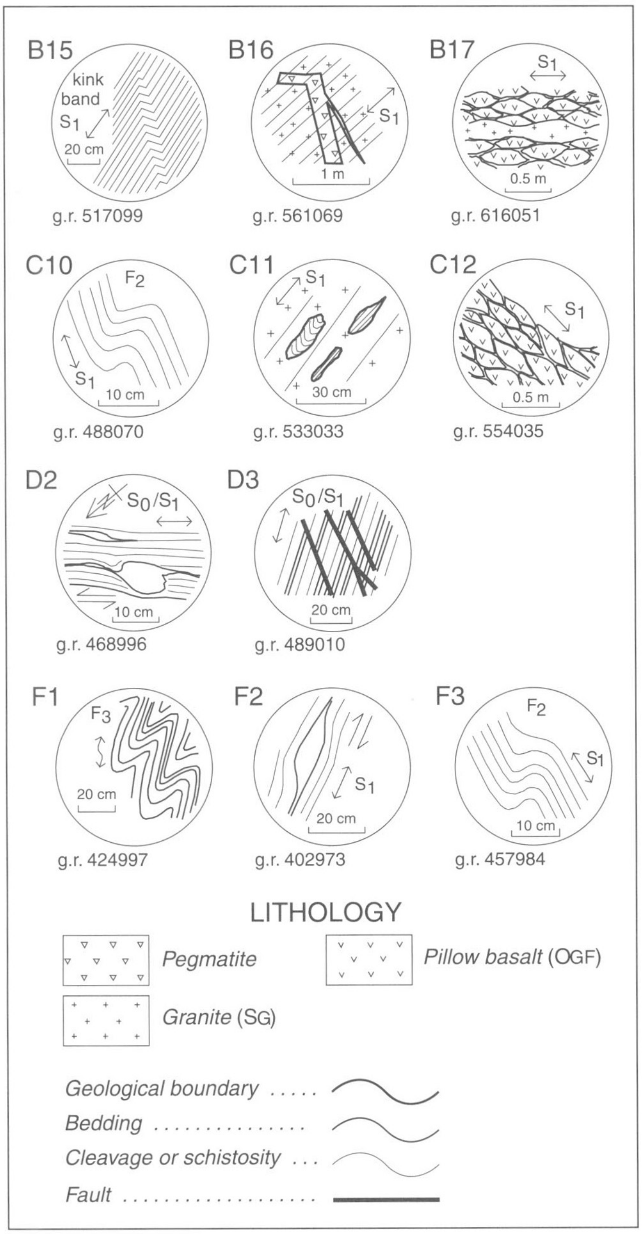


__Figure 38.__ Sketches of structural relations from outcrops in the Glover Island Belt. Letters and numbers refer to the cross-sections (Map 1893A, in pocket) and their relative positions in the sections from which they were compiled. All sketches are oriented with west to the left, except D2, which is in plan view.

These observations complicate the correlation of the S_1 foliation in the rocks of the Glover Formation with the S_1 in the Glover Island Granodiorite. The parallel orientation of the S_1 in both rock types suggests that they were formed in the same stress regime or that the older fabric was transposed. The relationships between the foliation in the xenoliths and the granodiorite indicate that the foliation in the pluton postdates an earlier event. The latter is confirmed by the fact that the foliation in the pluton is associated with chlorite retrogression, whereas in the Glover Formation the foliation is formed by the higher metamorphic grade actinolite. Presumably the deformation in the pluton took place during a later pulse of the D_1 event under lower metamorphic conditions, which had only a minor effect on the surrounding rocks of the Glover Formation.

Shear zones

Shear zones are common on all scales in the Glover Island Belt. The two main ductile dislocations are the Kettle Pond shear zone, which separates rocks of the Grand Lake Complex and the Glover Formation, and the one separating the rocks of the Grand Lake Complex from those of the Fleur de Lys Supergroup, previously named the Keystone shear zone (Cawood and van Gool, 1993), and corresponding with the Humber-Dunnage boundary in this region.

Kettle Pond shear zone

The Kettle Pond shear zone (see map, cross-sections C, D, F) consists of highly strained and interleaved lithologies from the adjoining Grand Lake Complex to the west and the Glover Formation to the east. The shear zone is steeply dipping and is on average 500 m wide. All rocks in the shear zone are retrogressed to either greenschist, sericite schist, quartz-porphyroclastic mica schist and serpentinite or talc-carbonate schist, representing deformed and retrogressed mafic rocks, felsic volcanic rocks, trondhjemite, and ultramafic rocks, respectively. Kinematic indicators are not consistent throughout the zone. Both dextral and sinistral movement occurred, suggesting a complex movement history. Kinematic indicators include C-S fabrics, shear bands, and asymmetrically tailed porphyroclasts. Mineral lineations in the steeply dipping, north-northeast-striking part of the shear zone plunge shallowly to the south-southeast, suggesting predominantly a strike-slip movement. Figure 39A shows a central part of the shear zone with a δ-shaped pod of calc-silicate, which, together with other kinematic indicators in this outcrop, suggests sinistral displacement. Kinematic indicators in the western, northeast-striking part of the shear zone indicate a reverse, dip-slip movement in their present orientation (Fig. 38, sketch F2). The foliation in the main, north-northeast-striking part of the Kettle Pond shear zone varies in orientation between steeply west-northwest-dipping and steeply east-southeast-dipping. The southern part of the shear zone is folded into a southwest-dipping orientation in the kilometre-scale Cobble Cove antiform.

Orientations of the foliation in the Kettle Pond shear zone are similar to S_1 in the wall rock. However, in the shear zone, amphibole that defines the foliation in the wall rock is retrogressed to chlorite, suggesting that, as in the Glover Island Granodiorite, the shear deformation took place during retrograde metamorphism at a later stage of deformation. The early deformational event could be related to the original emplacement of the ophiolitic complex and the rocks of the Glover Formation. The later shearing and associated retrogression occurred during subsequent juxtaposition of the two lithological units. Because of the similarity in orientation, both foliations were mapped as S_1.

Humber-Dunnage boundary

A south-east-trending shear zone separates Humber Zone rocks from Dunnage Zone rocks on Glover Island. It is predominantly developed within the basal Dunnage Zone rocks of the southwestern segments of the Grand Lake Complex. The shear zone is best exposed on the western shore of the island south of Cobble Cove, where steeply west-dipping, strongly foliated gabbro and serpentinite overlie and are tectonically mixed with psammite of the South Brook Formation. Kinematic indicators, including asymmetric quartz lenses and shear bands, show that the Dunnage Zone rocks moved up and over the psammite to the northeast. Towards the east, the shear zone is truncated by a set of steep late faults in the core of the anticline, which form the main boundary between Humber and Dunnage Zone rocks on Glover Island. These late faults are brittle, rectilinear structures and are considered to be splays of the Cabot Fault system. The ultramafic rocks immediately adjacent to the easternmost of these faults are variably sheared and metasomatized, suggesting that the late faults partially follow the trace of the shear zone.

D_2 deformation – F_2 folds

Folding of the S_1 foliation is not common and is only observed in the volcanic rocks of the Glover Formation, in the greenschists of the Grand Lake Complex, and in the shear zones. F_2 folds generally have no axial planar cleavage and usually form parallel folds (Fig. 39B). In sheared volcanic rocks south of Cobble Cove, F_2 crenulations and an S_2 crenulation cleavage occur locally, overprinting the S_1 foliation. Fold axes have a wide range of orientations, but the majority have a moderate southern plunge (Fig. 40). The axial planes are steep and have orientations at small angles to S_1, usually forming strongly asymmetric folds. The vergence of the F_2 folds on the two limbs of the Cobble Cove antiform generally agree with the expected east-vergence on the western limb and west-vergence on the eastern limb, suggesting that the antiform may be an F_2 structure. However, data were not consistently collected through the core of the antiform, so neither the vergence or the orientation of the D_2 structures in the core of the fold is well known. The stereographic plots of the S_2 data (Fig. 40) are inconclusive, and overall insufficient data are available to categorically establish the relative timing between the F_2 folding and formation of the Cobble Cove antiform.

D_3 structures

D_3 structures consist of open crenulations (F_3), which lack a well developed crenulation cleavage, and have shallow to moderate, north- or south-plunging axes (Fig. 40). The axial planes are commonly at a high angle to the S_1 foliation. In interbanded mafic and felsic volcanics south of Cobble Cove, overprinting relations between F_2 and F_3 folds were noted. Here the orientations of both sets of fold axes were variable. F_3 fold geometry is more open and lacks an axial planar cleavage. Where overprinting relationships are absent, the similar style of F_2 and F_3 folds hampered distinction between the two fold sets.

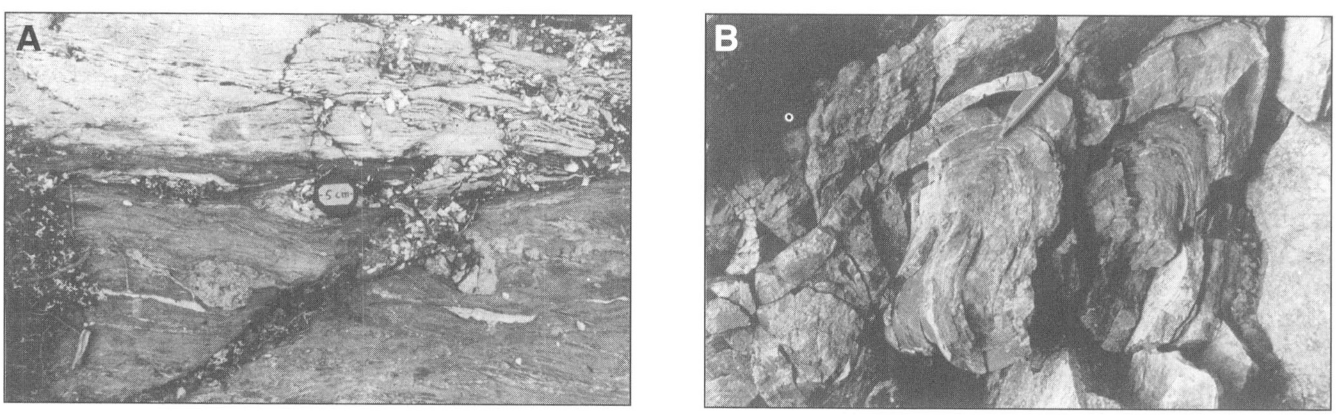


Figure 39. *Structures in the Glover Island Belt.* ***A)*** *High strain mafic schist, containing asymmetric calc-silicate pods and quartz-feldspar stringers (lower part of photo), in contact with intensely foliated trondhjemite (top part of photo) in the Kettle Pond shear zone on Glover Island. The calc-silicate pod to the lower left of the lens-cap has asymmetric tails that give it a d-shape. The shape of the pod and the shear bands in the outcrop indicate a sinistral sense of shear. Looking down on horizontal rock face (g.r. 468996). GSC 1995-149UU* ***B)*** *F_2 folds in interbedded mafic and felsic volcanic rocks of the Glover Formation on the west shore of Glover Island south of Cobble Cove. View looking northeast (g.r. 400970). GSC 1995-149BBBB*

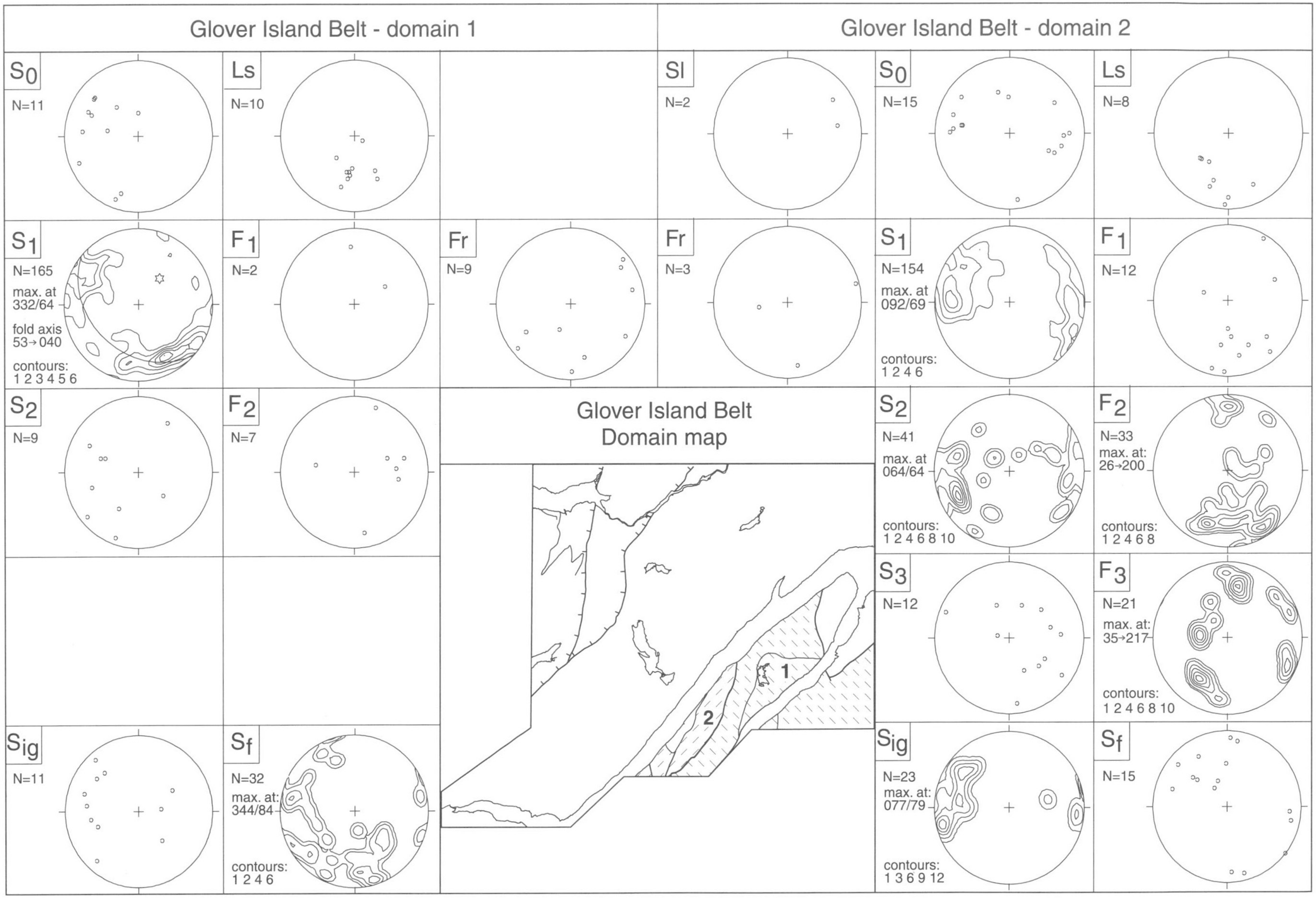


Figure 40. *Structural orientation data from the two domains in the Glover Island Belt. Linear data and poles to planar data, plotted on lower hemisphere, equal area projection. Contour levels in multiples of uniform distribution. Orientations of planes are quoted as dip direction and dip. Fr = fractures;* S_f *= fracture cleavage;* S_{ig} *= igneous layering; Sl = slickensides.*

Fracture cleavage

In the massive mafic rocks of the Glover Island Belt, the diabase and basalt of the Glover Formation and the gabbro of the Grand Lake Group, a fracture cleavage overprints the S_1 foliation. This cleavage is distinct from the late fracturing, in that no fracture fill exists and that only one dominant set occurs. East of Grand Lake, where the cleavage is most pervasive and best exposed, it has a fairly consistent northeastern strike, but elsewhere there is no consistent orientation (Fig. 40). The relative age of this fracture cleavage with respect to the F_2 and F_3 folding is uncertain. It may be the response in the massive rocks to the same stress field which caused the folding in the more schistose units. The brittle fracturing postdates the fracture cleavage.

Brittle faulting

The rocks along the shores of Grand Lake and in the thin slices of the Glover Island Belt to the north and west of Grand Lake are intensely fractured (e.g. Fig. 15A). The fracturing is most intense in the massive lithologies. Fractures are generally filled with chlorite and\or epidote. In most outcrops many fracture sets occur, most of them containing slickensides. Orientations of the fracture sets were not systematically measured and there is no clear preferred orientation (Fig. 40). In the slices of Glover Island Belt rocks north of Grand Lake, in the northeastern part of the map, fracturing and shearing are so intense that the rocks are locally pervasively chloritized and retrogressed to chlorite schist and original rock types are difficult to recognize. This fracturing, which is most intense along the trace of the main strands of the Cabot Fault, is interpreted to be of the same age as the Carboniferous faulting.

Deer Lake Basin

The Deer Lake Basin is a strike-slip basin resulting from dextral movement on the Cabot Fault system (Hyde et al., 1988). In most places the Carboniferous rocks are in fault contact with the adjoining lower Paleozoic rocks, except north of Northern Harbour, where rocks of the North Brook Formation unconformably overlie gneiss of the South Brook Formation (see map, cross-sections A, B).

The massive Carboniferous rocks of the Deer Lake Basin generally show few signs of penetrative outcrop-scale deformation, apart from occasional small faults. On a regional scale, rocks of the older Anguille Group, northwest of Grand Lake, are folded in a set of upright, shallowly north-northeast-plunging folds with wavelengths of tens to hundreds of metres (see map and cross-section A). Folding within the Carboniferous sequence is well exposed along the northwestern shore of Grand Lake (Fig. 41A). The fold hinges are commonly faulted, and fold limbs are straight. Outcrop-scale folds, with amplitudes less than 10 m, are rare, and only occur in thin-bedded sequences, so that fold axes are rarely measured in the field. Fold axis orientations were generally reconstructed from the intersection of the two limbs.

An S_1 foliation at a small angle to bedding is occasionally present in the shale and occurs locally as a very weak fracturing in the more massive rocks. S_1 is steeply oriented and is axial planar to the large scale folds (Fig. 42). Younging directions in the Carboniferous sedimentary rocks were determined from sedimentary structures. They indicate that the folds are facing upwards. In siltstone which is interbedded with sandy layers, southwest of Northern Harbour, close to the fault that bounds the western margin of the Carboniferous rocks, a steeply northwest-dipping fracture cleavage forms a second fabric (Fig. 43, sketch B14). Outside this area by Northern Harbour, no second cleavage was observed in the rocks of the Deer Lake Basin.

The rocks of the younger Deer Lake Group along the eastern margin of the map area, are folded in an open kilometre-scale syncline (cross-sections A, B). No outcrop-scale folds were found in these rocks, and parasitic folds are only indicated by slight variations in the orientation of the bedding on the scale of tens to hundreds of metres.

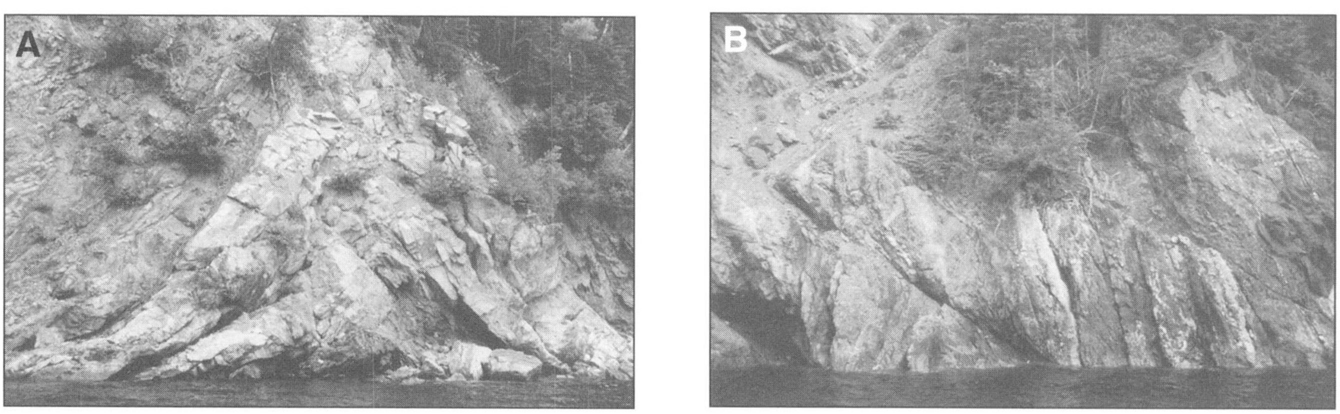


***Figure 41.** Structures in the Deer Lake Basin. **A**) Antiform in thick-bedded sandstone of the Carboniferous Saltwater Cove Formation, along the western shore of Grand Lake, east of Northern Harbour. The rocks are faulted in the fold hinge. View looking north (g.r. 636192). GSC 1995-149FFF **B**) Reverse faults (dipping to the right) in a thick-bedded sequence of the North Brook Formation, along the shore of Grand Lake in the northeastern corner of the map area. In the left of the photo, beds are dragged towards the orientation of the fault plane. View looking north (g.r. 623191). GSC 1995-149III*

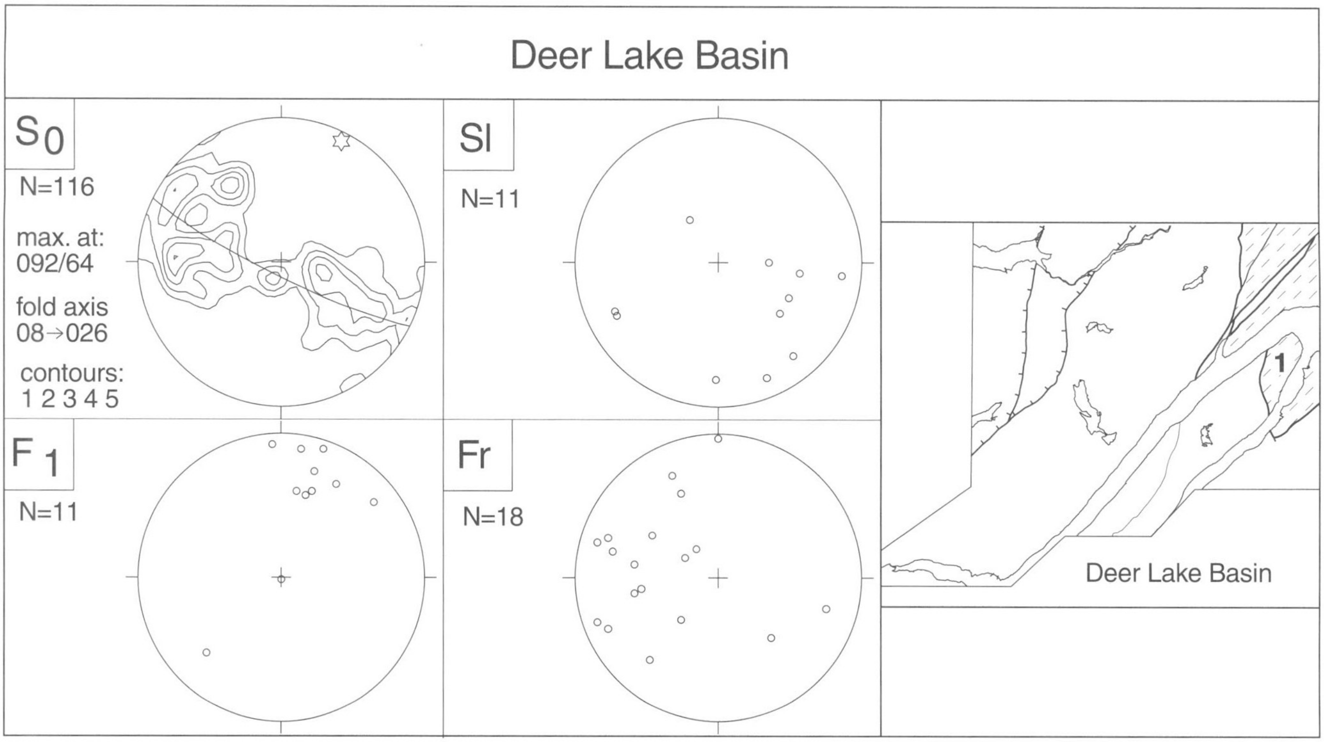


Figure 42. Structural orientation data from the Carboniferous rocks of the Deer Lake Basin. Linear data and poles to planar data, plotted on lower hemisphere, equal area projection. Contour levels in multiples of uniform distribution. Orientations of planes are quoted as dip direction and dip. Fr = fractures; Sl = slickensides.

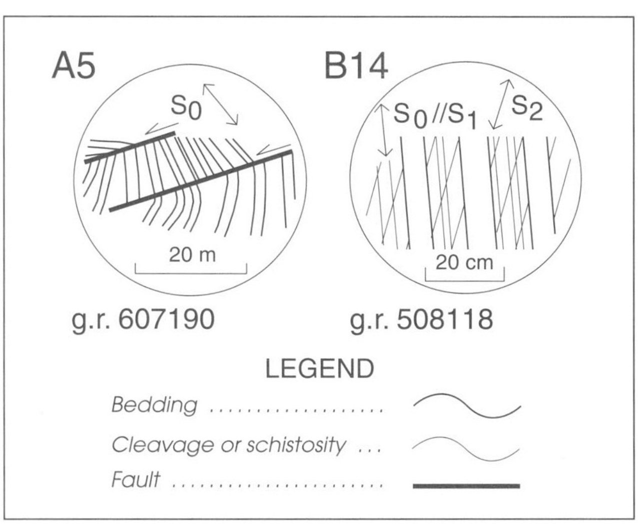


Figure 43. Sketches of structural relations in outcrops from the Deer Lake Basin. Letters and numbers refer to the cross-sections (Map 1893A, in pocket) and their relative positions in the sections from which they were compiled. The sketches are oriented with west to the left.

Reverse, normal, strike-slip and oblique faults with offsets on the scale of several metres to several tens of metres cut the sedimentary rocks (Fig. 43, sketch A5; Fig. 41B). Some of the smaller faults can be related to space problems in the hinges of the folds, but many are direct expressions of local compression or extension. These brittle faults and fractures were not measured systematically, but the available data show that the majority of the faults have moderate easterly dips with slickensides that vary between moderately east-plunging and gently south-plunging (Fig. 42).

Figure 42 shows that the poles to the bedding planes are oriented in a great circle, which defines the shallowly north-northeast-plunging fold axis, consistent with the orientation of the few measured fold axes.

Hyde et al. (1988) explained the occurrence of compressional structures (reverse faults and folds) in the Deer Lake basin by a period of transpression that followed original basin subsidence during transtensional strike slip. An alternative explanation for the extensional and compressional structures is that in a pattern of anastomosing strike-slip faults convergence and divergence causing local areas of compression and extension (Reading, 1980) which are not constant in the fault zone. A true change in stress field or movement pattern of the fault zone is then not required, and both extensional and compressional structures are the result of the strike-slip regime.

Cabot Fault system

The Cabot Fault is an orogen-scale northeast-trending, high-angle, strike-slip fault cutting through western Newfoundland. It is late Paleozoic and locally trails the Dunnage Zone-Humber Zone boundary. In the Corner Brook Lake-Glover Island region, fault blocks within the strike-slip system show significant relative vertical movement, indicating oblique components of displacement. The presence of the large sedimentary basin (Deer Lake Basin) at the northern end of the area suggests that the Cabot Fault system had an overall transtensional setting.

The western arm of Grand Lake is the site of a major strand of the Cabot Fault system, which runs on-shore 3 km southwest of Northern Harbour and continues mainly through Carboniferous strata in a northeast trend. Apart from the sliver of basement gneiss and South Brook schist on the west side of Glover Island, the fault corresponds with the Humber-Dunnage boundary, and metamorphic grade changes from amphibolite grade west of the fault zone to greenschist grade in the east. Locally, smaller faults splay from the main zone, such as at Northern Harbour and on Glover Island. Along the eastern arm of Grand Lake, rocks of the North Brook Formation show no apparent offset indicating no late Carboniferous movement on any fault on this side of the lake. However, the restriction of the Glover Island Granodiorite to the island, and the absence of similar rocks directly east on the eastern shore of the lake, apart from the small Red Indian Brook Granodiorite which is apparently offset from the Glover Island body, may indicate an early Carboniferous or older fault in this region. High-angle, brittle faults in the map area occur many kilometres away from the main locus of the fault, and most of these are attributed to the Cabot Fault system. They are specifically ubiquitous in the southwestern part of the map area. The majority have a northeastern trend, largely parallel to the main orientation of the fault, but east- and north-trending faults occur as well.

Fracturing due to fault movement is concentrated along the western arm of Grand Lake and in the slice of Dunnage Zone rocks northwest of Northern Harbour. Fractures are of variable orientation and no consistent pattern was noted. The majority of poles to the fractures in the Carboniferous rocks plot in the western half of the stereographic plot in Figure 42, representing moderate eastern dips. Slickensides are common, but do not indicate consistent orientations of displacement.

More ductile shearing associated with the movement on the Cabot Fault system was observed along the contacts between gneiss of the Corner Brook Lake Belt and the thin slices of Glover Island Belt rocks at the western shore of Grand Lake (g.r. 373965), on the contact between Deer Lake Basin rocks and Glover Island Belt rocks, which is well exposed on the western shore of Grand Lake 3 km southwest of Northern Harbour (g.r. 508118), and in a small quarry on the logging road northeast of Northern Harbour (g.r. 546179). At the latter outcrop, a steeply east-dipping fault juxtaposes conglomerate of the Saltwater Cove Formation in the west against sheared and chloritized mafic volcanic rocks of the Glover Formation to the east. The orientation of the fault and of the moderately northeast-plunging mineral lineation on fault surfaces near the bounding fault suggest a reverse-sinistral component of fault displacement, but elsewhere on the outcrop other orientations of the mineral lineation occur. The mafic volcanic rocks are transected by a dense anastomosing network of small shear zones, which consist predominantly of oriented chlorite. Farther east in this slice, granitoid rocks of the Island Pond Pluton are cataclastically deformed, but here chloritization is again so intense that recognition of the protolith is often hampered. In outcrops on the shore of Grand Lake, southwest of Northern Harbour, a zone several hundred metres wide contains tectonically mixed rocks of mafic volcanics, gneiss, shale, and sandstone (of both early Paleozoic and Carboniferous age), graphitic schist, thin coal seams, carbonate, and carbonate-conglomerate, all intensely foliated, chloritized, cut by calcite veins and cataclastically deformed. Kinematic indicators show mixed movement senses, with a predominance of eastern block-up displacement. The lowest, westernmost unit of what is continuous Saltwater Cove Formation, consists of a pebble to cobble conglomerate, containing predominantly bluish carbonate clasts (similar to the Breeches Pond Formation) with minor quartz, schist, and volcanic pebbles in a foliated chlorite-quartz-calcite-epidote matrix (Fig. 44A). Cataclastic zones, including ultracataclasite and pseudotachylyte, cut the shear fabric. The ultracataclasite to pseudotachylyte forms thin irregular bands of no more than 5 mm width, that consist of poorly sorted, rounded to angular clasts of the wall rock (predominantly feldspars are preserved) in a submicroscopic matrix, which is often brown, but any original glassy texture is no longer preserved (Fig. 44B). Thin chlorite and carbonate-filled cracks cut the pseudotachylyte.

In the thin slices of granitoid rocks of the Little Paddle Point pluton on the west shore of Grand Lake and on the contact with the gneiss to the west, very thin shear zones (less than 10 cm wide) form an anastomosing pattern. The shear bands are overprinted by fractures, which are generally filled with chlorite and epidote, but which locally form thin, black pseudotachylyte veins. The majority of the shear bands and fractures have a steep northwest strike. The plot of S_1 planes in Figure 45 represents measurements of the foliation in these shear zones related to the Cabot Fault system.

The ductile and brittle features found in the Cabot Fault zone may indicate that faulting occurred at different conditions, presumably during a protracted time or separate pulses of deformation.

No direct evidence was found for the sense of displacement of the Cabot Fault. Kinematic indicators in foliated zones or on fault planes are scarce and often ambiguous. Within one outcrop indications could be found for movement in many orientations, suggesting a very inhomogeneous movement pattern. Hyde et al. (1988) used indirect evidence to show that displacement on the fault zone is dextral. Figure 46 shows the orientation of the folds in the rocks of the Anguille Group with respect to the trace of the Cabot Fault. Assuming that the folds are related to the same stress field as the fault, and that the principal stress axes are orthogonal to the orientation of the folds, the σ_1 stress direction is resolved onto the

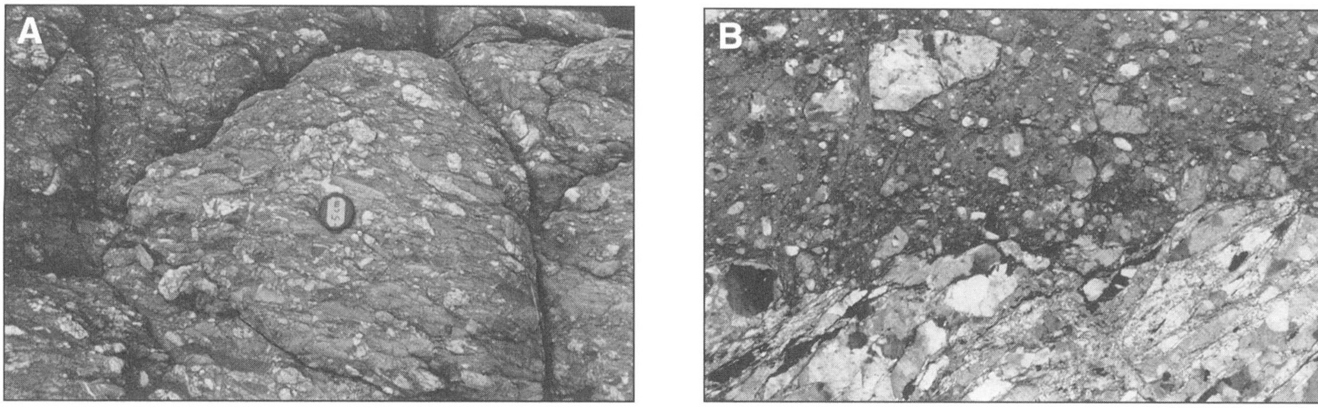


***Figure 44.** Structures related to the Cabot Fault. **A**) Deformed pebble to cobble conglomerate of the Saltwater Cove Formation (map unit Cs) consisting of elongated, predominantly carbonate clasts in a foliated, chlorite-rich matrix. A second cataclastic fabric cuts the main foliation at an angle of about 25°. Looking down on slightly inclined rock face (g.r. 508118). GSC 1995-149QQ **B**) Photomicrograph of an ultracataclasite to pseudotachylyte band in a gneiss of the South Brook Formation in an outcrop on the western shore of Grand Lake, southwest of Northern Harbour, near the western branch of the Cabot Fault (g.r. 503114). Angular and sub-rounded fragments sit in a dark brown matrix, which is partially recrystallized. Polarizors at 70° angle. Width of the photo represents 7.5 mm. GSC 1995-149YYY*

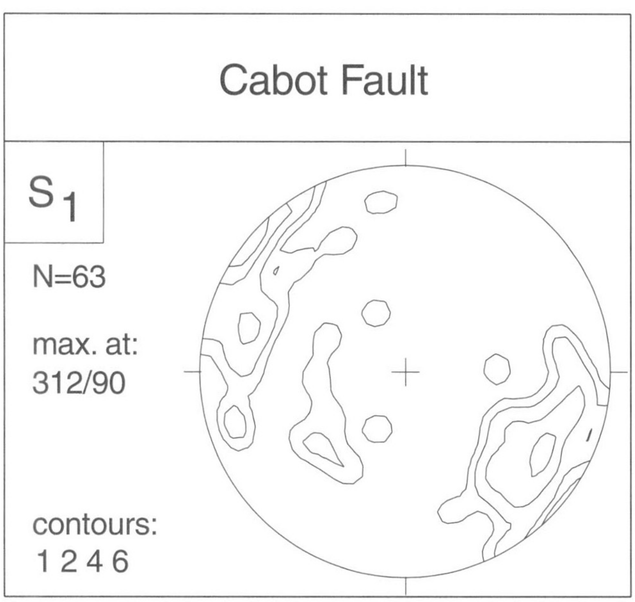


***Figure 45.** Stereographic plot showing orientation of S_1 along the Cabot Fault in the vicinity of Northern Harbour. Poles to planes, plotted on lower hemisphere, equal area projection. Contour levels in multiples of uniform distribution.*

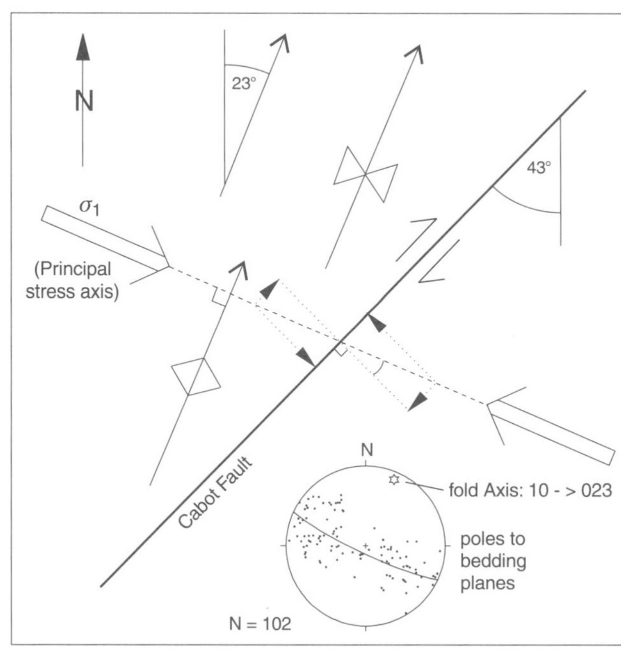


***Figure 46.** Schematic representation of the orientation of folds in sedimentary rocks of the Anguille Group, northwest of Grand Lake, and the orientation of the Cabot Fault. Assuming that the folds are approximately perpendicular to the principal stress axis, this stress orientation resolves onto the fault plane as a dextral stress couple. The stereo plot shows the orientations of the fold axes in the sediments.*

fault plane with a dextral shear stress couple. The angular relationship between folds and fault plane is therefore consistent with dextral displacement on the Cabot Fault.

Correlation of the structures

Each of the lithotectonic belts has a distinct structural history indicating that they were in different tectonic settings during at least part of their tectonic development, but sufficient similarity does exist between generations of structures in the different belts to enable correlation. However, not all generations of structures recognized in the map area can be correlated across lithotectonic boundaries, and hence, D_1 in one belt is not necessarily the same age as D_1 in the other belts.

Figure 47 is a schematic representation of proposed correlation of structures across the lithotectonic belts. The final juxtaposition of the belts along their present boundaries is indicated by the numbered squares, but the similarity of structures before this time suggests that the four lower Paleozoic belts were in proximity during part of their development. Initial juxtaposition of the Humber Arm Allochthon with the Carbonate Belt, and the Glover Island Belt with the Corner Brook Lake Belt, occurred during the Ordovician Taconian orogeny. These two original thrust contacts, have been largely masked by subsequent events.

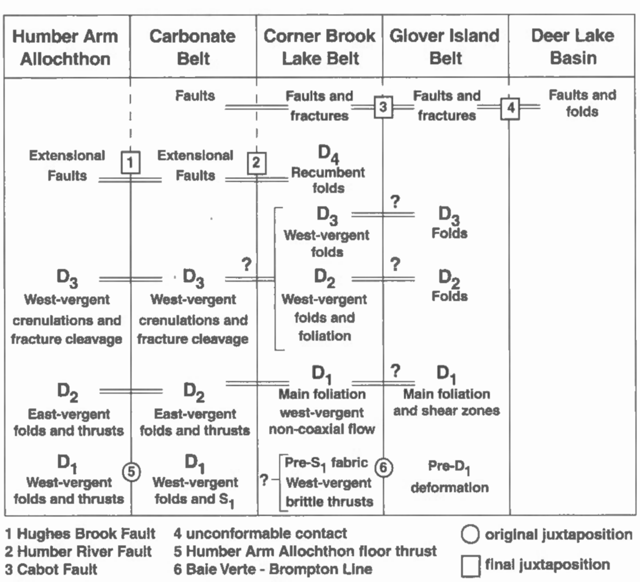


Figure 47. *Table showing the correlation of structures across the boundaries of the five lithotectonic belts. Circles indicate the time of initial docking of the lithotectonic units with respect to the adjacent ones. Squares indicate the final juxtaposition of two adjoining belts in their present relative position. Horizontal double lines indicate a correlation of generations of structures. Dashed vertical lines between columns indicate that the belts are juxtaposed and have a common structural history. Solid vertical lines indicate separation of the belts.*

The overall style and sequence of deformation in the Humber Arm Allochthon and the Carbonate Belt are quite similar. Both show early west-vergent folds related to thrusting, followed by east-vergent folding and thrusting and the development of west-vergent F_3 crenulation and fracture cleavage. The westward vergence of the D_1 deformation in the Humber Arm Allochthon and the recumbent nature of at least some of the folds (Fig. 19) suggest it may be related to west-directed emplacement of the allochthon onto the shelf, which is represented by the Carbonate Belt. Allochthon emplacement is a Taconian age event (Stevens, 1970) and correlation of D_1 in the allochthon with west-vergent D_1 folds and associated cleavage in the Carbonate Belt imply a similar age for this event. However, in the Humber Zone, penetrative deformation associated with allochthon emplacement is considered to be limited to the allochthon and does not extend into the footwall, such as the rocks of the Carbonate Belt (Cawood and Williams, 1988). Penetrative deformation of the carbonate sequence elsewhere in the Humber Zone is considered to be Silurian or Devonian (Cawood and Williams, 1988; Cawood, 1993; Cawood and Dunning, 1993). This apparent conflict can be explained by one of the following alternatives: the D_1 events in the Humber Arm Allochthon and Carbonate Belt do not correlate; age assignment for either or both of the D_1 structures is wrong; deformation associated with allochthon emplacement extended to deeper structural levels towards the orogenic hinterland (cf. Cawood and Williams, 1988). There are insufficient data to resolve which, if any, of these alternatives is correct. However, there is no direct control on the age of the D_1 structures in either belt, and evidence against a Taconian age for the D_1 fabric in the allochthon is the absence of any mélange fabric associated with D_1 structures. Mélanges are a characteristic of Taconian deformation elsewhere within the zone (Stevens, 1970; Williams and Stevens, 1974).

The D_2 structures in the Humber Arm Allochthon and the Carbonate Belt are correlated on the basis of their similarity in geometry, orientation, and relative timing. It is assumed that by this time the Humber Arm Allochthon overlay the carbonates and that they were together affected by the east-vergent deformation. The D_3 structures in both belts are also of a similar nature and can be correlated. These two belts were juxtaposed by late extensional faulting on the Hughes Brook fault, which lowered the rocks of the Humber Arm Allochthon to those of the Carbonate Belt, in a top-down to the west movement. This caused additional extensional faulting throughout the two belts.

Pre-D_1 brittle thrusting and the pre-D_1 foliation development in the Corner Brook Lake Belt are possibly related to a single event. The west-directed thrusting is either correlated with the west-directed thrusting of the Humber Arm Allochthon, and is therefore a Taconian event, or it is the start of the Silurian tectonic development, and the onset of the build-up of a thrust wedge which subsequently was deformed by ductile processes (van Gool and Cawood, 1994).

The D_1 deformation in the Corner Brook Lake Belt is mainly a result of west-directed movement. It is tentatively correlated with the D_2 event in the two western belts because of the similarity of the nature of the deformation. D_2 in the west and D_1 in the east are foliation-forming events, both strongly asymmetric, but with opposite polarities. In that case the west-vergent and east-vergent deformation must converge in a triangle zone, presumably close to the present location of the Carbonate Belt. The steeper dips of the S_2 foliation in the Carbonate Belt may be a result of the convergence of the two opposite-directed movements. D_2 and D_3 in the Corner Brook Lake Belt are progressive developments of the overall west-vergent deformation, and can be correlated with the west-vergent D_3 deformation in the Humber Arm Allochthon and Carbonate Belt. The last deformation in the Corner Brook Lake Belt is D_4 folding, which results in lateral extension of the belt. Because of the extensional nature of this deformation, it is tentatively correlated with the extensional faulting in the two western belts, and also related to the extensional Humber River Fault, but with continued D_4 folding also subsequently deforming the fault.

The earliest deformational features in the rocks of the Glover Island Belt are found in xenoliths in the Glover Island Granodiorite, and are tentatively linked with the oldest foliation in the Glover Formation. This fabric must predate the 440 Ma age of the granodiorite and may be related to the Taconian orogeny. Formation of the main shear zones postdates the age of the granodiorite, and may be similar to the 430 Ma age of deformation in the Corner Brook Lake Belt. The D_1 deformation in these two belts is correlated, based on these age data and on the apparent similarity of the S_1 foliation in the slice of Humber Zone rocks on Glover Island and the surrounding Glover Island Belt rocks. Consequently, the two belts were juxtaposed by the end of the D_1 deformation, and F_2 and F_3 folding are therefore also correlated. Final juxtaposition of the two belts to their present positions occurred during the movement of the Cabot Fault. This Carboniferous faulting, which caused the deposition and deformation of the sediments in the Deer Lake Basin, is presumed to be responsible for the latest brittle faults in all lithotectonic units, with the exception of the Humber Arm Allochthon, where Carboniferous deformational features are not recognized.

Summary

Based on the structural field observations in combination with data from literature from the surrounding regions, a simplified tectonic history can be reconstructed. The deformation started in the Glover Island Belt by thrusting of the Dunnage Zone rocks (pre-D_1) onto the continental margin rocks of the Humber Zone, and was associated with emplacement of the Humber Arm Allochthon onto the continental shelf (? D_1 in allochthon). Earliest thrusting in the rocks of the Corner Brook Lake Belt (pre-D_1) and the Carbonate Belt (D_1) either occurred at this time or at the beginning of the major mid-Paleozoic orogenic pulse. D_1 in Glover Island Belt resulted in juxtaposition of the Grand Lake Complex and the Glover Formation along the Kettle Pond shear zone, and the contemporaneous juxtaposition of this belt against the rocks of the Corner Brook Lake Belt. This resulted in the shearing and west-directed ductile thrusting in the Corner Brook Lake Belt (D_1), as well as the east-directed thrusting and folding in the two western lithotectonic units (D_2 in Humber Arm Allochthon and Carbonate Belt). After the docking of the lithotectonic units, progressive crustal shortening resulted in ductile, west-vergent folding of all rocks (D_2 and D_3 in Glover Island Belt and Corner Brook Lake Belt; D_3 in Humber Arm Allochthon and Carbonate Belt). The Yellow Marsh fold nappe formed during late D_2 in the Corner Brook Lake Belt, and its emplacement represents a late stage of ductile thrusting in the area that culminates in the floor thrust of this fold nappe. The end of plate convergence resulted in the drop of the compressive forces and extensional collapse of the orogen was accomplished by recumbent folding in the Corner Brook Lake Belt (D_4) and extensional faulting in the western units. Transtensional plate movements along the Cabot Fault zone resulted in the creation of the Deer Lake Basin, and the folding and faulting of the sediments deposited in it. The basement rocks to these sediments, and the adjoining lithotectonic units, underwent brittle strike-slip faulting as a result.

METAMORPHISM

The Upper Proterozoic to Lower Paleozoic rocks in the map area have been affected by one regional metamorphic event, which ranged in the rocks of the Humber Zone from lower greenschist facies in the west of the map area, to intermediate-pressure amphibolite facies east and south of Corner Brook Lake (Cawood and van Gool, 1992). The rocks in the Glover Island Belt are in greenschist facies (Knapp, 1982). The thermal imprint on the rocks of the Deer Lake Basin was of sub-greenschist grade and reflects depth of burial (Hyde et al., 1988). In the Humber Zone and Dunnage Zone rocks, metamorphic grade was determined from metamorphic mineral assemblages, and for the Deer Lake Basin from a combination of analyses published by Hyde et al. (1988). At the time of regional metamorphism in the Middle Paleozoic, the Humber Arm Allochthon, Carbonate Belt, and the Corner Brook Lake belt were juxtaposed, although not in their final relative position, and were affected by a common metamorphic event. The metamorphic histories of these three lithotectonic units are treated together whereas the development of the Glover Island Belt and the Deer Lake Basin are described separately.

The Proterozoic metamorphism of the gneiss of the Corner Brook Lake Complex is largely overprinted by the Paleozoic metamorphic event. Currie and van Berkel (1992b) reported the occurrence, south of Grand Lake, of two-pyroxene granulite grade gneiss in the Disappointment Hill complex, and in migmatite in an unnamed granitoid-gneiss complex, which is correlated with rocks of the Corner Brook Lake Complex, indicating that metamorphic grade reached in situ melt conditions. Mineral assemblages in basement rocks in the Corner Brook Lake Complex are attributed to Paleozoic metamorphism.

Metamorphism of the Humber Zone

The Middle Paleozoic metamorphic event resulted in a metamorphic gradient which is assumed to be continuous in the rocks of the Humber Zone, with the exception of breaks at the extensional Humber River and Hughes Brook faults and Carboniferous high-angle faults. Determination of the metamorphic field gradient in the Humber Zone was based on variations in metamorphic mineral assemblages in pelite, semipelite, and psammite. In the area with the highest metamorphic grades near Corner Brook Lake, a sequence in the growth of minerals was established, which could be interpreted as a qualitative P-T path. Metamorphic mineral assemblages in the calc-schist and mafic rocks were not studied in detail, but are summarized for comparison.

Mineral assemblages and reactions in pelitic to psammitic rocks

Most sedimentary rocks in the Humber Zone, except the massive carbonate units, contain sufficient pelitic and semipelitic rocks to provide a reasonable coverage of the area for the reconstruction of a metamorphic map, and the interpretation of the variations in equilibrium mineral assemblages. The data presented here are predominantly from a petrographic study, but were supplemented with field observations, especially in the southwestern part of the area, where petrographic data are scarce. Gaps in the database formed by the massive platform carbonate rocks and the quartzofeldspathic gneiss of the Corner Brook Lake Complex caused only localized uncertainties in the position of isograds on the map.

Based on the distribution of mineral assemblages in pelite, semipelite, and psammite, five metamorphic zones were defined. These are from west to east, in order of increasing metamorphic grade: 1) muscovite-chlorite; 2) biotite; 3) garnet; 4) staurolite; 5) biotite-garnet-kyanite. The mineral assemblages that are diagnostic for these metamorphic zones, and the main reactions that represent the transitions between them, are shown in Table 3 and are schematically presented in AFM diagrams in Figure 48. The regional distribution of the observed mineral assemblages and the isograds that bound the metamorphic zones are plotted on the map in Figure 49. Zone 1 covers the Humber Arm Allochthon and the Carbonate Belt, whereas the Corner Brook Lake Belt covers the four higher grade zones. The boundary between zones 1 and 2 is formed by the Humber River Fault, rather than an isograd, and hence no isograd is indicated in the map. The upper stability of chlorite is also indicated on the map, but the position is not based on a specific reaction and its location in the field is too uncertain to

Table 3. Mineral assemblages in the metamorphic zones of the Humber Zone and separating mineral reactions.

	ASSEMBLAGE	ISOGRAD REACTION	
	Zone 1 - muscovite - chlorite		
1a	Ms + Chl ± Qtz		
1b	Ms + Chl + Pl ± Qtz		
	Zone 2 - biotite		
		Kfs + Chl = Bt + Ms + Qtz + H_2O	R1
2a	Ms + Chl + Bt + Pl ± Qtz		
2b	Ms + Chl + Cld + Pl		
	Zone 3 - garnet		
		Ms + Chl = Grt + Bt + Qtz + H_2O	R2
3a	Ms + Chl + Grt + Bt + Pl + Qtz		
3b	Ms + Chl + Grt + Pl + Qtz		
3c	Ms + Chl + Grt ± Bt + Pl		
3d	Ms ± Chl + Grt (with Cld inclusions) + Pl + Qtz		
	chlorite-out		
3e	Ms + Grt ± Bt + Pl + Qtz		
	Zone 4 - staurolite		
		Ms + Chl + Grt = Bt + St + Qtz + H_2O	R3
4	Ms + Grt + Bt + St + Pl + Qtz		
	Zone 5 - garnet - biotite - kyanite		
		St + Ms + Qtz = Grt + Bt + Ky + H_2O	R4
5a	Ms + Grt + Bt + Ky + Qtz ± Pl		
5b	Ms + Grt + Bt + Ky + St ± Pl		
		Qtz + Ms = Kfs + Ky + H_2O	R5
	Mineral name abbreviations after Kretz (1983).		

be used for the subdivision into another metamorphic zone. Most of these zones have several diagnostic assemblages, some of which can also occur as non-diagnostic assemblages in higher grade zones, depending on bulk composition. This is reflected in the AFM diagrams of Figure 48 by a darker shading of the fields representing diagnostic observed mineral assemblages, and a lighter shading for assemblages that are not diagnostic for a metamorphic zone. The numbering of the assemblages is such that they indicate the lowest grade zone in which they are observed. The petrogenetic grid in Figure 50 shows estimated locations in P-T space of the assemblages, representing the metamorphic field gradient, and the isograds that bound the zones. The mineral

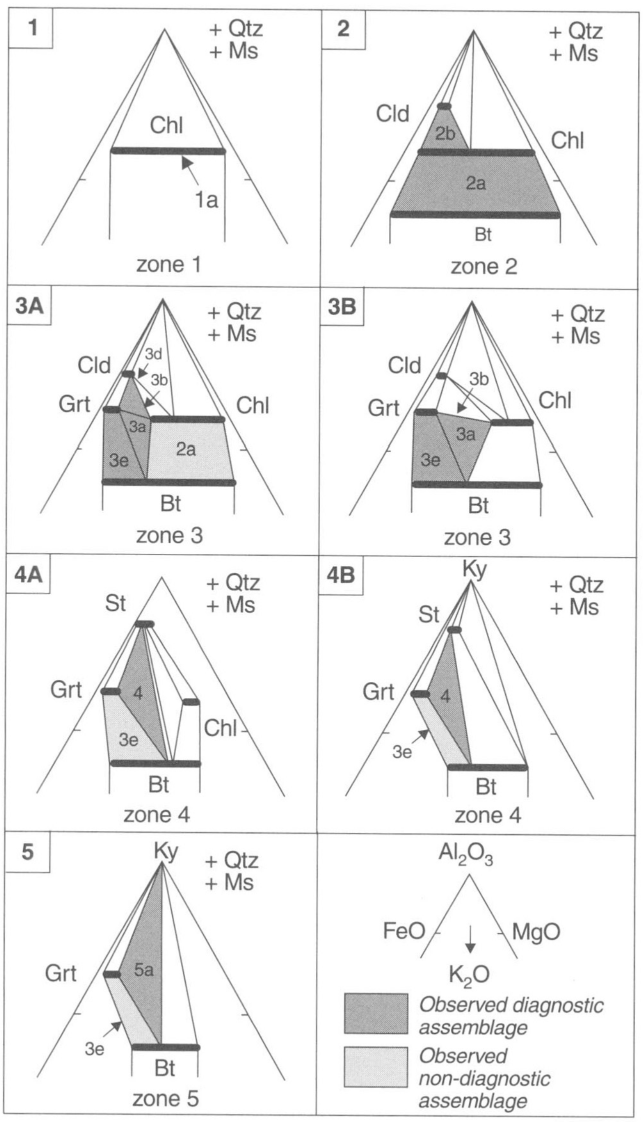


Figure 48. *AFM diagrams representing the possible mineral assemblages in the five metamorphic zones in the pelite and psammite of the Humber Zone. Observed mineral assemblages are indicated by the shaded areas, and the numbers refer to Table 3. Abbreviations for mineral names according to Kretz (1983).*

assemblages, petrographic observations, and inferred mineral reactions are described from the lowest to the highest metamorphic zone.

Zone 1: muscovite + chlorite

Metamorphic zone 1 contains the Humber Zone Allochthon and the Carbonate Belt in the map area, and is characterized by the occurrence of the assemblages white mica+chlorite± quartz (1a), which occurs predominantly in the western part of the zone, and white mica+chlorite+quartz+plagioclase (1b), which is found in the eastern part of the zone. Both assemblages occur in the absence of biotite or chloritoid, and 1b is assumed to represent slightly higher grades than 1a. They plot on an AFM diagram on the chlorite line (Fig. 48). These assemblages occur in the slate and sandstone of the Summerside and Irishtown formations, near Corner Brook, and in slate of the Pinchgut and Reluctant Head formations. Slate consists of finely intergrown white mica, chlorite, and quartz, together with graphite, plagioclase and fine-grained opaque oxides (pyrite). Siderite, tourmaline, and calcite/dolomite are accessories. Plagioclase was only recognized in samples in the eastern part of the zone (assemblage 1b), but this could be because at the lowest metamorphic grades it is too fine grained to be optically distinguished from quartz. K-feldspar is not common. White mica and chlorite form a schistosity which is intensely crenulated in most rocks (Fig. 22B). In the sandstone, quartz and plagioclase form up to 80% of the clasts, and K-feldspar, magnetite, epidote, muscovite, and tourmaline form a minor part of the clasts. Very fine-grained quartz, muscovite, chlorite, probably plagioclase, and locally minor calcite/dolomite form the matrix. Assemblage 1b occurs also in higher grade zones as a nondiagnostic assemblage. An overall increase of the grain size of the newly grown minerals, and the occurrence of plagioclase in the eastern part, indicates an increase in metamorphic grade towards the east.

Zone 2: biotite

The occurrence of biotite together with muscovite and chlorite (assemblage 2a), or chloritoid+white mica+chlorite+albite in the absence of biotite and quartz (assemblage 2b), characterizes the biotite zone. Biotite and chloritoid first occur at, or very close to, the Humber River Fault, suggesting that the western boundary of zone 2 is not a reaction isograd, but a structural boundary. Beside the change in mineral assemblages, the fault also marks a notable change in overall grain size of micas from a few micrometre thick to 10 or 20 μm thick. Biotite is scarce in the rocks of the Corner Brook Lake Belt, presumably as a result of their generally aluminous bulk rock compositions, and the biotite zone is defined on very few occurrences. Assemblage 2a occurs also in the garnet zone.

Samples of pelite with assemblage 2a contain a matrix similar to that of zone 1, consisting of finely intergrown muscovite and chlorite, commonly with quartz and plagioclase, and porphyroblasts of coarser white mica, albite, biotite, and rare chlorite, all of which overgrow a finer white mica-chlorite

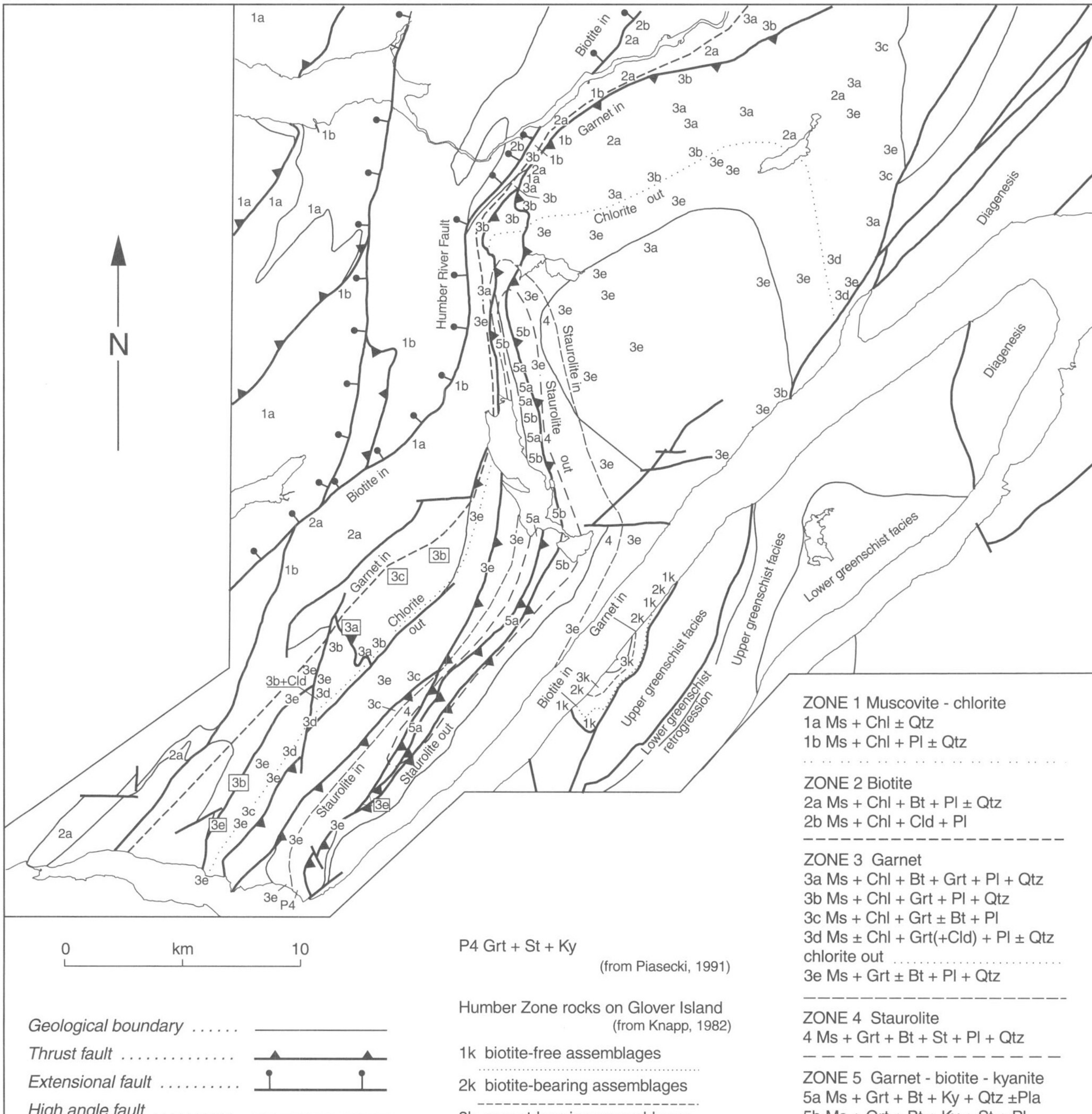


Figure 49. *Map of observed metamorphic mineral assemblages and reconstructed isograds, which bound the metamorphic zones that are defined in the text. The first occurrence of biotite is immediately east of the Humber River Fault and is not defined by an isograd. Number-letter combinations refer to the mineral assemblages in Table 3. Assemblages surrounded by a rectangle are from field observations. P4 is the tentative location of a Grt-St-Ky occurrence reported by Piasecki (1991). 1k, 2k and 3k on Glover Island are respectively biotite absent, biotite bearing, and garnet bearing assemblages from Knapp (1982).*

schistosity. Zoisite/epidote, siderite, magnetite, ilmenite, and tourmaline are accessory minerals. In a few samples of assemblage 2a, quartz is only present in minor amounts or was totally absent.

Locally the white mica porphyroblasts grow along the axial planes of F_1 or F_2 crenulations to form a recrystallized crenulation cleavage (Fig. 51A), but overgrowth of randomly distributed white mica porphyroblasts, either in a preferred or in a random orientation, was also observed (Fig. 51B). The newly grown white mica locally obliterates the pre-existing foliation indicated only by trails of opaque dust, and forms a new coarser matrix. These textures are also observed in higher grade rocks, where they are more common.

Albite locally forms small porphyroblasts up to 2 mm in diameter, but generally are less than 1 mm. They have well developed inclusion trails of fine-grained quartz, white mica, and locally graphite (Fig. 51C). These albite porphyroblasts are common in the pelite of the northwestern part of the Corner Brook Lake Belt north of Steady Brook. Biotite occurs in the pelite of metamorphic zone 2 and higher grade rocks, predominantly as isolated, fairly large, subhedral to anhedral porphyroblasts, up to 5 mm in diameter, which overgrow the pre-existing white mica-chlorite foliation (Fig. 51D). Partial retrograde replacement of biotite by chlorite is present in many samples.

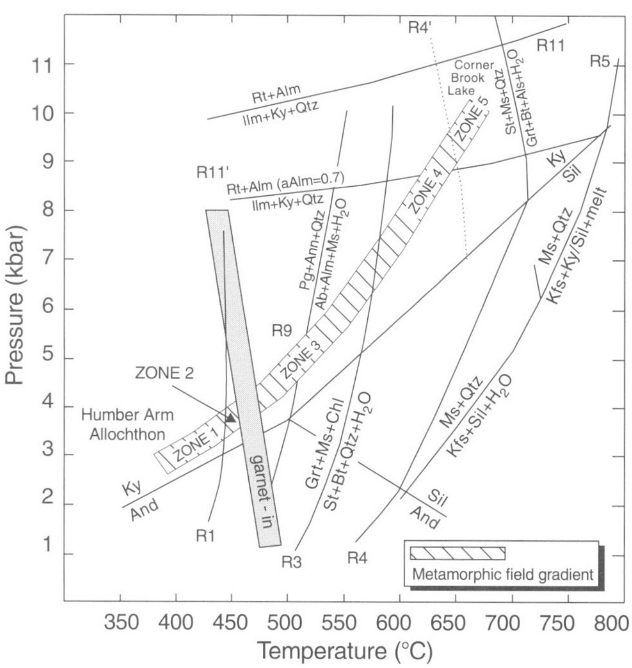


Figure 50. *Schematic petrogenetic grid with the metamorphic field gradient of the Humber Zone and the most important mineral reactions. All reactions are shown for end-member compositions, except R11', in which a_{Alm} equals 0.7 and R4', which is an estimate. Alumino-silicate stability fields from Holdaway (1971), reaction R1 from Rivers (1983), R3 and R4 from Spear and Cheney (1989), R5 and garnet-in isograd from Yardley (1989). The remainder were constructed using the computer program TWQ by Berman (1991), and the thermodynamic database by Berman (1988).*

Biotite is more common in semipelite, and psammite than in pelite, and it occurs either as part of a coarser grained matrix, intergrown with white mica (Fig. 51C), or as large subhedral to anhedral porphyroblasts. Albite porphyroblasts similar to those in the pelite are common but not as abundant as in the pelite, and plagioclase occurs as a matrix mineral. K-feldspar is present as relicts of clasts in a few samples.

In pelite of the South Brook Formation west of Humber River, and in the lowest outcrops on the Marble Mountain ski-slopes, chloritoid occurs together with chlorite, white mica, and plagioclase (assemblage 2b). The chloritoid is associated with plagioclase, and in a sample from west of the river it occurs in a radiating pattern around the plagioclase. The association between the plagioclase and chloritoid and the absence of quartz in these samples suggest that albite and chloritoid formed by the reaction:

$$\text{paragonite + chlorite + quartz = chloritoid + albite} + H_2O \quad \text{[R7]}$$

(Miyashiro, 1973). In an AFM diagram this assemblage tentatively plots in the chlorite-chloritoid field (Fig. 48), although it lacks quartz.

Reaction R1 is only one of several possible biotite-producing reactions.

$$\text{K-feldspar + chlorite = biotite + muscovite + quartz} + H_2O \quad \text{[R1]}$$

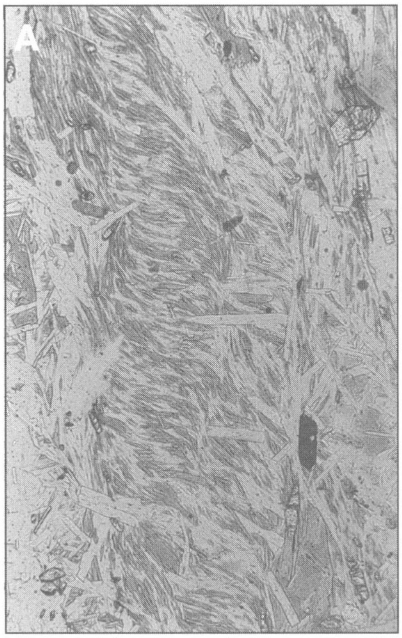
The formation of biotite in the presence of K-feldspar is responsible for the first appearance of biotite in the psammite and is used as the lower boundary of the biotite field in the petrogenetic grid (Fig. 50). The reaction

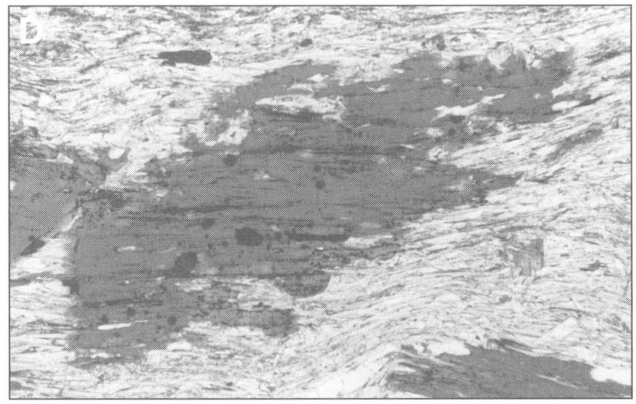
$$\text{chlorite + phengite + quartz = biotite + muscovite} + H_2O \quad \text{[R6]}$$

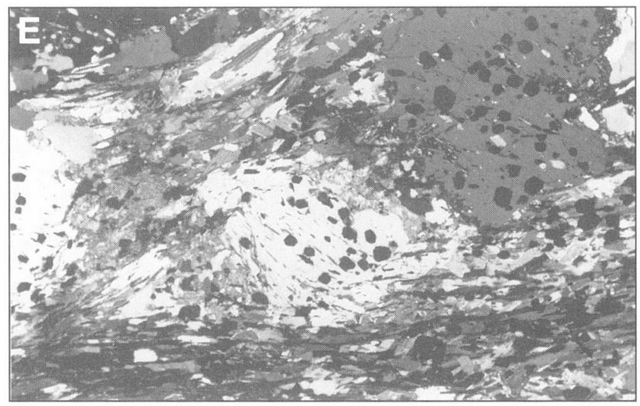
presumably produces biotite in the pelite. It would be responsible for the consumption of chlorite, which explains why chlorite is not abundant in this zone together with biotite. The position of R1 in P-T space is uncertain because of its dependence on mineral compositions and the activity of H_2O in the fluid phase.

Zone 3: garnet

The western boundary of the garnet zone is drawn along the westernmost occurrences of garnet-bearing assemblages. In the southwest of the map area, field observations were used to construct the garnet-in isograd, to supplement scarce petrographic data. The chlorite-out isograd is drawn within the garnet zone. Its position is poorly delineated, because a wide zone exists in which assemblages both with and without chlorite occur. The occurrence of chlorite in these rocks depends both on bulk rock composition and on metamorphic grade. Five diagnostic garnet-bearing mineral assemblages occur in the garnet zone (Table 3). Assemblage 3c cannot be plotted in the AFM diagram in Figure 48 because of the lack of quartz. Assemblage 3b plots on the chlorite-garnet tieline.







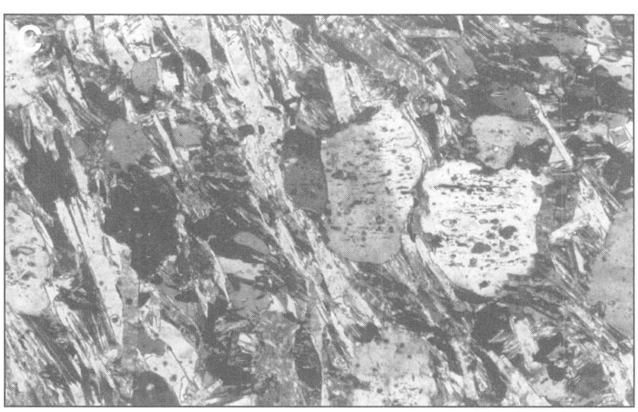


Figure 51. *Photomicrographs of the relationships between foliations and porphyroblasts.* ***A)*** *F_1 crenulations in which a pre-D_1 schistosity, consisting of a fine intergrowth of chlorite and white mica, is crenulated, and larger white mica grows parallel to the S_1 cleavage plane. Sample from Marble Mountain ski slopes (g.r. 494206). Plane-polarized light. Height of the image is 1.35 mm. GSC 1995-149RR* ***B)*** *Isolated muscovite porphyroblasts (colourless white bands, at a high angle to the foliation) have grown across a schistosity consisting of a fine intergrowth of paragonite and chlorite (darker parts of the photo). Graphite trails outline the foliation inside the neoblasts. Sample from garbenschiefer of the Breeches Pond Formation, north of Grand Lake Brook (g.r. 332992). Plane-polarized light. Height of the image is 1.35 mm. GSC 1995-149ZZZ* ***C)*** *Albite-mica schist of the South Brook Formation. Two albite porphyroblasts on the right show an internal fabric of fine-grained inclusions (pre-S_1) which is at an angle to the S_1 foliation in the matrix, which postdates the albite. Sample from north of Humber River (g.r. 434265). Cross-polarized light. Width of the image is 1.35 mm. GSC 1995-149RRR* ***D)*** *Anhedral biotite porphyroblast overgrowing an S_1 foliation, defined by muscovite, chlorite, and graphite. Sample of garnetiferous mica schist of the South Brook Formation, south of Corner Brook lake (g.r. 336984). Plane-polarized light. Width of the image is 4.7 mm. GSC 1995-149XX* ***E)*** *Photomicrograph of albite porphyroblasts that grew between D_1 and D_2 in garnet zone in albite schist of the South Brook Formation, west of Northern Harbour (g.r. 512164). The albite contains garnet (black euhedral grains), biotite, and zoisite inclusions which define an internal fabric S_1. The foliation in the matrix (S_2) wraps around the albite. Cross- polarized light. Width of the image is 9.9 mm. GSC 1995-149XXX*

Pelite in the garnet zone contains abundant white mica in the matrix, with one or more of chlorite, quartz, biotite, and plagioclase. Zoisite/epidote is a common accessory, but forms a main phase in some samples. Magnetite, ilmenite, and graphite are other accessories. Porphyroblasts are garnet, biotite, albite, and locally muscovite. Two samples contained chloritoid porphyroblasts.

Pelite commonly contains subhedral, poikiloblastic garnet several millimetres in diameter, but locally up to 2 cm across. Most garnets have abundant inclusions of quartz, zoisite, ilmenite, and locally chloritoid. Chloritoid inclusions occur only in rocks without biotite (assemblage 3d). Biotite is not common in the pelite and occurs as subhedral to anhedral porphyroblasts (Fig. 51D) and rarely as part of the matrix. Above the chlorite stability it can occur in a fine intergrowth with white mica, where it presumably replaced chlorite that was originally intergrown with the white mica. This fine schistosity is not commonly preserved, but it is locally present in rocks that contain little quartz and abundant graphite, which can prevent the grain boundary migration needed for grain growth. Usually the white mica in the matrix is recrystallized to a larger grain size, forming either a decussate texture or a new foliation. The recrystallization of white mica is pervasive only in rocks with abundant quartz. Microprobe analyses were performed on a pelite from south of Corner Brook Lake, which contained the overgrowth textures of coarse white mica over finer grained white mica, intergrown with biotite (c.f. Fig. 51B). The results indicated that the original white mica in the fine-grained matrix consists of paragonite, whereas the coarser micas are muscovites. Representative analyses of the two micas are shown in Table 4.

Chlorite is not common and occurs only as matrix mineral, locally interlayered with white mica. In rocks of certain bulk compositions, mainly quartz-deficient rocks, the chlorite can occur up to the staurolite isograd. Chlorite commonly pseudomorphically replaces garnet and biotite during retrograde metamorphism. Quartz is generally present in minor amounts in the matrix in the pelite, but is absent in some samples. Albite porphyroblasts occur in most pelite samples and can be up to one centimetre in diameter. In hand specimen they occur usually as white, rounded porphyroblasts with abundant inclusions, but in graphite-rich micaceous rocks they occur as dark grey, stubby minerals in which no inclusions can be recognized with the naked eye. Most of the albite porphyroblasts are slightly to strongly zoned, with increasing anorthite content towards the rim, resulting locally in oligoclase in the rims.

In the semipelite and psammite of the South Brook Formation, similar assemblages were found, but garnet porphyroblasts are smaller, usually less than 0.5 mm in diameter, and biotite is more abundant, both as a matrix mineral and as porphyroblasts. Garnets are either poikiloblastic or inclusion free, and have subhedral to euhedral shapes. Inclusions are quartz, zoisite, albite, and rarely biotite or white mica. In psammite from the northern part of the garnet zone, garnet has fairly large albite inclusions, of the same size as albite porphyroblasts in the matrix. Albite is abundant in the psammite. It occurs as large, unzoned porphyroblasts (locally over 2 cm in diameter) in the core of the Yellow Marsh antiform (see map, unit P€SBa). They have different inclusion patterns than the albite in the pelite. Garnet, zoisite/epidote, biotite, and white mica are the most common inclusions, which can be fairly large and show a weak preferred orientation. Inclusions of garnet in albite porphyroblasts from the central and eastern part of Yellow Marsh (Fig. 51E) are in contrast with the occurrence of albite as inclusions in garnet in pelitic rocks and in psammite in the northern, external part of the garnet zone. The relative timing of the growth of the minerals is not constant throughout the area.

The character and position of the garnet-in isograd in the P-T diagram are uncertain. Several reactions can be responsible for the first occurrence of garnet, and their position is strongly dependent on the composition of the participating phases. Therefore, the garnet-in isograd is shown as a wide zone in the P-T diagram (Fig. 50), representing several reactions and reflecting the uncertainty of the position. One of the main garnet-forming reactions in pelite is:

$$\text{chlorite} + \text{muscovite} = \text{garnet} + \text{biotite} + \text{quartz} + H_2O \qquad [R2]$$

Table 4. Representative white mica compositions[1] in sample 92-C806 (g.r. 336984).

Oxide weight %	Paragonite in matrix	Muscovite porphyroblast	Cations (22 ox.)	Paragonite in matrix	Muscovite porphyroblast
SiO_2	45.39	46.63	Si	5.86	6.19
TiO_2	0.16	0.50	Ti	0.02	0.05
Al_2O_3	38.75	35.18	Al^{IV}	2.14	1.81
			Al^{VI}	3.76	3.70
FeO	1.59	1.20	Fe[2]	0.17	0.13
MgO	1.40	1.03	Mg	0.27	0.20
CaO	0.56	0.00	Ca	0.08	0.00
Na_2O	5.47	1.59	Na	1.37	0.41
K_2O	2.28	8.51	K	0.38	1.44
Total	95.61	94.64	Total	14.04	13.93
			X_{par}	0.79	0.22
			X_{mus}	0.21	0.78

[1] Single analyses
[2] Total iron as Fe^{2+}

which represents the transition from AFM diagram 2 to 3A in Figure 48. With rising temperatures this continuous reaction will deplete chlorite from the matrix and enrich the remaining chlorite in Mg, while forming almandine-rich garnets. This results in a shift of the tie lines in the AFM diagram to the right, expanding the garnet field and decreasing the chlorite field, so that progressively less bulk rock compositions fall in the chlorite field (Fig. 48, diagram 3B). The disappearance of chlorite from the observed mineral assemblages in the garnet zone is not a result of a terminal reaction and does not represent a true isograd.

The breakdown of chloritoid could possibly likewise be caused by shifting tie-lines. Chloritoid can be broken down to form garnet by the reaction:

chloritoid + 2 chlorite + 6 quartz = 5 garnet + 13 H_2O [R8]

which would also result in a shift to the right of the tie-lines. This breakdown of chloritoid presumably takes place before chlorite is consumed in most rocks, because no chloritoid was observed within the chlorite-absent zone. Many other garnet-forming reactions are published, some of which may have occurred in the Corner Brook Lake Belt.

The reaction that causes the replacement of fine-grained matrix paragonite by coarser grained muscovite, also produces albite. The reaction:

annite + paragonite + quartz = albite + muscovite + almandine + H_2O [R9]

was proposed by Jamieson and O'Beirne-Ryan (1991), to be the main albite-producing reaction in a study of metamorphic mineral assemblages in pelitic rocks of the Fleur the Lys Supergroup in Baie Verte (Fig. 2), where albite is abundant as a late metamorphic phase. The observation that, in the rocks of the Corner Brook Lake Belt, the muscovite growth occurs only in the presence of quartz, is consistent with this reaction. The reaction can be responsible for the recrystallization of the fine-grained matrix, as well as the depletion of quartz or biotite in some rocks, and the abundance of garnet and albite.

Zone 4: staurolite

Only four samples have a stable staurolite assemblage, all four containing the same minerals: muscovite, garnet, biotite, staurolite, plagioclase, and quartz (assemblage 4). The garnet-biotite assemblage 3d and other nondiagnostic staurolite-free assemblages are common. The eastern and southern extent of the zone are poorly defined because of the lack of suitable lithologies. A garnet-staurolite-kyanite occurrence on the western arm of Grand Lake reported by Piasecki (1991) is assumed to be near the bend in Grand Lake (tentatively positioned at the location marked P4 in Fig. 49), and is taken as an assemblage of the staurolite zone, but might be of even higher grade.

In the pelite of zone 4, staurolite forms small euhedral porphyroblasts, and in three out of four samples it was not identifiable in hand specimen. Staurolite can be poikiloblastic and commonly contains quartz inclusions, defining an internal foliation. In quartz-poor rocks inclusions are absent. Biotite is common in this metamorphic zone and is generally in contact with staurolite. It forms relatively large (up to 2 mm across) anhedral porphyroblasts which randomly overgrow the foliation. In graphite-rich rocks, biotite can occur in a fine intergrowth with muscovite to define a microscopic schistosity. Albite and garnet have similar features, as in the pelite of the garnet zone, with generally larger grain sizes. Albite is zoned, with a rim of increased anorthite content surrounding a core of constant composition. This zoning is locally visible in the field as a white rim around a grey core. Muscovite is the main matrix mineral, together with quartz, feldspar, rare biotite, and in the Breeches Pond Formation also calcite/dolomite. Tourmaline, titanite, apatite, ilmenite, and rutile form accessories.

Psammite occurring in the staurolite zone is similar to that in the garnet zone, with the exception that albite inclusions in garnet are absent.

The observation of staurolite together with biotite in thin section suggests that the two are formed together. The staurolite-in isograd is located in a series of road outcrops of the South Brook Formation, 10 km south of Corner Brook Lake (g.r. 336984 to 340981). Samples from west of the isograd have the assemblage Ms + Grt + Chl + Bt, which changes to the east into Ms + Grt + St + Bt + Qtz, with biotite and staurolite closely related, suggesting that staurolite is formed by:

muscovite + chlorite + garnet = biotite + staurolite + quartz + H_2O [R3]

This reaction, together with the chloritoid breakdown reaction to form staurolite and garnet, form the transition from AFM diagram 3B to 4A (Fig. 48). The absence of chloritoid and chlorite in the rocks of the garnet zone near the staurolite zone, may explain why staurolite assemblages are rare, lacking two important staurolite-forming reactants. In the staurolite zone the reaction:

muscovite + staurolite + chlorite = biotite + kyanite + quartz + H_2O [R10]

represents the transition from AFM diagram 4A to 4B, and kyanite could be formed (Fig. 48). No kyanite-bearing assemblages were found in the staurolite zone, suggesting that the bulk rock compositions were too rich in iron to form kyanite (Fig. 48, diagram 4B).

Zone 5: garnet+kyanite+biotite

The stable occurrence of the breakdown products of staurolite defines the zone with the highest metamorphic grades in the Corner Brook Lake Belt. The zone forms a narrow belt around the floor thrust of the Yellow Marsh fold nappe. Two mineral assemblages are diagnostic for this zone, Ms+Grt+Bt+Ky+Qtz±Pl (assemblage 5a) and Ms+Grt+Bt+Ky+St±Pl (assemblage 5b; Fig. 52A). The non-diagnostic assemblage 3e occurs in a few locations (Fig. 48, AFM diagram 5). The most common pelitic rock is a quartz-muscovite schist with abundant albite and small garnet porphyroblasts, but this zone contains a very coarse garnet-kyanite-staurolite-albite schist. This pelitic unit is abundantly exposed on the slopes east of Corner Brook Lake (Fig. 9D, F) and in isolated outcrops elsewhere. It commonly contains high amounts of graphite, which occurs concentrated in bands in the rock, that may represent original bedding. Ironically, large staurolite crystals, up to 7 cm long, occur only in zone 5, above the staurolite stability where they always occur together with the staurolite-breakdown products, garnet, kyanite and biotite. The staurolite is anhedral and can be poikiloblastic. Euhedral staurolite occurs in quartz-free rocks, together with garnet and either kyanite or biotite, or with all three (Fig. 52A). Kyanite occurs as up to 15 cm long euhedral crystals, and locally pseudomorphically replaces staurolite. Anhedral kyanite is less common and is poikiloblastic. Garnet is abundant in most pelite and is up to 5 cm in diameter. It has similar features as in the lower grade zones. In a few samples chloritoid inclusions occur in garnet, besides the more common quartz, epidote, and ilmenite. Biotite is common only in rocks that contain staurolite or its breakdown products, and forms relatively large porphyroblasts as in the lower grade zones. Albite can form porphyroblasts up to 5 cm in diameter in the graphite-rich garnet-kyanite schist (Fig. 9F) and is also strongly poikiloblastic, but contains predominantly graphite, muscovite, and minor quartz or biotite inclusions, which define an internal foliation (S_1). Zoning patterns in albite are similar to those in albite schist at staurolite grade. The matrix of the pelite consists predominantly of muscovite, with minor amounts of quartz, plagioclase, and biotite. Quartz is absent in some rocks, and in these quartz-deficient samples, white

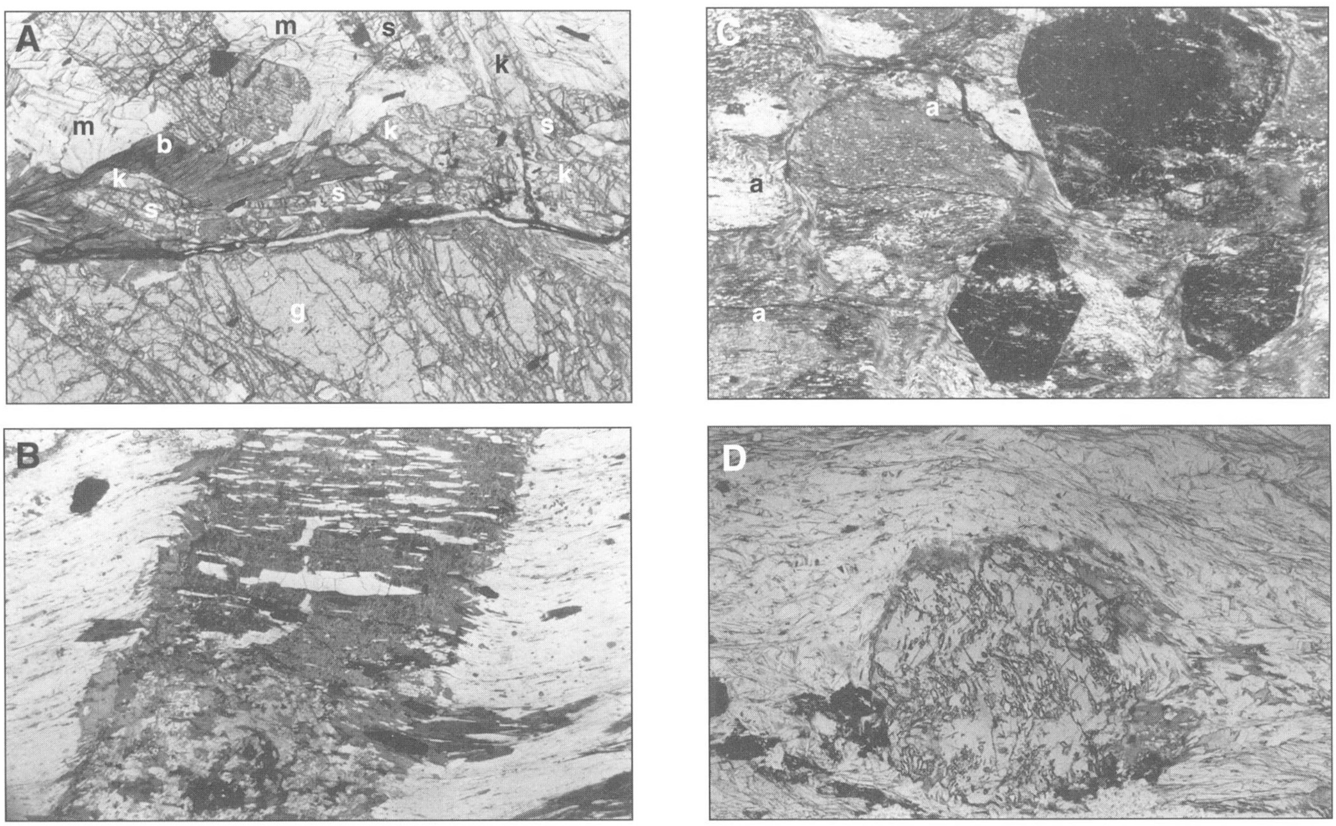


Figure 52. *Photomicrographs of pelitic schist.* ***A)*** *Coexisting garnet (g) + biotite (b) + kyanite (k) + staurolite (s) + muscovite (m) (assemblage 5b) in a pelitic schist of the South Brook Formation, east of Corner Brook Lake (g.r. 393089). Width of the image is 5.0 mm. GSC 1995-149GG* ***B)*** *Poikiloblastic hornblende has grown over S_1 in a garbenschiefer of the Breeches Pond Formation, southwest of Corner Brook Lake on Lady Slipper Road (g.r. 365019). The amphibole is pseudomorphically replaced by biotite. The internal fabric consists predominantly of quartz. Post peak metamorphic deformation caused minor buckling of the foliation along the sides of the amphibole. Plane-polarized light. Width of the image is 8.9 mm. GSC 1995-149I* ***C)*** *Garnet and albite porphyroblasts with a straight internal fabric (S_1) in an F_2 crenulated matrix. The black euhedral crystals are garnet, albite (a) is extremely poikiloblastic and is recognized in the photo as white and grey areas with a straight foliation. The matrix consists predominantly of crenulated muscovite. Garnetiferous albite-mica schist of the Breeches Pond Formation, west of Corner Brook Lake (g.r. 377059). Cross-polarized light. Width of the image is 15 mm. GSC 1995-149OO* ***D)*** *Garnet porphyroblast with inclusion trails that delineate F_1 crenulations. Crenulations in the matrix have the same axial plane, but are much tighter as a result of D_2 coaxial overprinting, and they are recrystallized, so the hinges are not commonly preserved. Garnetiferous quartz-muscovite schist of the South Brook Formation, 1 km northeast of Breeches Pond (g.r. 394194). Plane-polarized light. Width of the image is 5.4 mm. GSC 1995-149H*

mica in the matrix is extremely fine grained. Accessory minerals are tourmaline, titanite, monazite, zoisite, and rutile. Textures and mineral relations in the psammite are similar to those in the lower grade zones. None of the higher grade mineral assemblages were observed in these rocks.

Reaction R4 forms the reaction isograd between zones 4 and 5.

$$\text{staurolite + muscovite + quartz} = \text{garnet + biotite + kyanite} + H_2 \quad [R4]$$

No kyanite was found outside this metamorphic zone, although it can be formed at conditions lower than the staurolite breakdown. The occurrence of staurolite together with its breakdown products, suggests that these rocks were near the staurolite-out isograd, which is true for all rocks in this zone that is no wider than about 3 km. Also, the lack of quartz in some of these rocks plays an important role in the prevention of total breakdown of staurolite, since quartz is one of the reactants in reaction R4, and its absence will stop the breakdown.

The appearance of rutile together with kyanite, quartz, and garnet, and the absence of ilmenite in the matrix, suggests that the reaction:

$$\text{ilmenite + kyanite + quartz} = \text{almandine + rutile} \quad [R11]$$

has taken place, which requires that the rocks have experienced fairly high pressures (Fig. 50). This reaction took place relatively late in the metamorphic development, because inclusions in garnet are ilmenite, as they are in the matrix of lower grade rocks.

Metamorphic field gradient and spatial distribution of the isograds

The petrogenetic grid of Figure 50 shows the main mineral reactions and a proposed metamorphic field gradient in P-T space. The locations of the reactions do not represent the exact conditions for the examined rocks, because most are end-member reactions. True compositions and activities of the minerals or phases were not determined, and most reactions will shift when mineral chemistry data of the rocks in the area are used. Furthermore, most reactions involve water, and water activities lower than 1.0 will also cause a shift of the position of the reactions. However, the grid gives a semiquantitative impression of the variation of the metamorphic conditions throughout the map area. It should be kept in mind that the pressures quoted here are the minimum pressures, using mainly the kyanite-sillimanite isograd as a lower limit. The maximum pressure is not constrained, and a wide margin will be given for the minimum pressures.

Zone 1, representing the Humber Arm Allochthon and the Carbonate Belt, experienced pressures and temperatures that were lower than about 3 to 4 kbar and 440°C, presumably decreasing to the west. The western margin of the Corner Brook Lake Belt forms the biotite zone, and the P and T were in the order of 3-4 kbar and around 460°C.

The majority of the rocks of the Corner Brook Lake Belt are in the garnet zone, which includes most of the Yellow Marsh fold nappe and a 5 km wide zone in the southwest of the map. In the field, the garnet isograd is situated at, or in the footwall of, the lowest recognized continuous thrust in the Corner Brook Lake Belt. The maximum temperature in the garnet zone is defined by the staurolite-in isograd, which represents conditions of about 5 to 7 kbar at 570 to 580°C. The chlorite-out isograd and the boundary of the albite porphyroblastic lithologies in the South Brook Formation (map unit P€SBa), which is also a metamorphic boundary, are both folded in the Yellow Marsh fold nappe.

The western boundary of the staurolite zone in the field, defined by the R3 isograd, is located within a 12 km long slice of South Brook Formation lithologies, southwest of Corner Brook Lake. The remainder of this isograd, from the northern tip of this slice to Eastern Lake, is located in basement lithologies, where it cannot be accurately positioned due to the lack of suitable pelitic sequences. The eastern part of the trace of the staurolite-in isograd is also poorly constrained in this predominantly psammitic thrust sheet. It is positioned along the easternmost observed occurrences of stable staurolite, but it could actually be located farther east, if the assemblage 3e is regarded as a subassemblage of zone 4.

In the petrogenetic grid the staurolite zone is probably too large. The maximum P-T conditions, defined by the R4 isograd as shown in Figure 50 would be about 8-10 kbar and 700°C, which seems too high for the observed assemblage. Furthermore, this would put the upper part of the staurolite zone above the minimum melt curve in the presence of water. The lack of indications of in situ melting in the map area suggests that either the melt curve was never reached, or the water activity and bulk rock composition were unsuitable for melting. If the R4 isograd were constructed using mineral compositions from the map area (e.g. garnet with lower activity of almandine instead of pure almandine), it presumably would shift considerably to lower temperatures, as is shown in Figure 50 in the location of reaction R4'. A more realistic estimate for the upper limit of staurolite stability is about 650°C, which would bring pressure estimates down to 7-9 kbar.

Zone 5 assemblages are restricted to the base of the Yellow Marsh fold nappe and the Valley Lakes thrust stack. The western boundary of zone 5 assemblages is located within the basement slice west of the thrust stack, and the eastern boundary is constrained by the occurrence of stable staurolite + quartz + muscovite, which is found in a few locations east and north of Corner Brook Lake. South of the lake, the eastern boundary of zone 5 is uncertain.

The peak metamorphic pressures and temperatures in zone 5 are constrained by the staurolite breakdown reaction R4' and the kyanite isograd, which are assumed to provide a minimum of about 650°C and 7-9 kbar. The observed assemblage Rt+Grt+Ky+Qtz within zone 5, indicates pressures and temperature on the high pressure side of R11. The position of this reaction was recalculated using an activity of Alm of 0.7, and is shown as R11' in the petrogenetic grid. Combining R4' and R11' gives a minimum P and T for the rocks in zone 5 of about 9 kbar and 650°C (Fig. 50). The occurrence of staurolite with quartz and the staurolite breakdown products in a few of the samples suggests that the staurolite breakdown conditions were not far exceeded. An absolute maximum to the metamorphic conditions is provided by R5, the upper stability limit of quartz and muscovite. Preliminary thermobarometry by Ryan (1992) on a sample from zone 5 east of Corner Brook Lake provided P-T estimates for one Grt-Bt-Ky-Qtz-Pla assemblage of 8.5 kbar and 578°C, which was derived from a garnet rim. Compared with the petrogenetic grid, this is a slight underestimate of maximum conditions, which may be explained by slight retrogression of the analyzed garnet at the rim, providing P-T conditions during the retrograde part of the P-T path. Alternatively, the location of reaction isograd R4' is located at too high temperatures.

The isograds in the slice of Humber Zone rocks on Glover Island were taken from Knapp (1982). The isograds trace the shape of the bedding in the Cobble Cove antiform and must be folded. The shape and location of the isograds in these rocks is not continuous with the metamorphic zonation in the area west of Grand Lake, which is another indication that the Cabot Fault postdates the metamorphism in the belt.

The overall pattern of isograd distribution in the Humber Zone shows that the metamorphic zonation is parallel with the trend of the thrust faults, and that the isograds are closest in areas of intense thrust stacking, especially in the area north of Corner Brook Lake, where the transition from biotite grade rocks (lower greenschist facies) to rocks that are above the staurolite stability (upper amphibolite facies) takes place within 1500 m across strike, suggesting that the isograds have been telescoped by the thrusting. Folding of isograds is only obvious in the Yellow Marsh antiform, where the "chlorite-out" and the "albite porphyroblast-in" boundaries follow the folded trend of the foliation, and in the Cobble Cove antiform.

The telescoping of the isograds suggests that a causal relationship exists between thrusting and metamorphism. Presumably metamorphism resulted from crustal thickening by the thrust stacking, and the isotherms were deformed during ensuing thrust movement.

Metamorphism of other lithologies

The variation of the metamorphic conditions throughout the area is well constrained in the pelites and semipelites. The metamorphic zones are also reflected in the mineral assemblages in amphibolites and calc-schists, but these rock types are not very useful for the establishment of metamorphic zoning, because they are not as widespread as the pelites and semipelites, and buffering of reactions that involve both CO_2 and H_2O strongly influences the P-T conditions at which reactions occur. Therefore, only a cursory comparison is presented of observed mineral assemblages in these rock types with the established metamorphic zoning. The marbles in the area are restricted to a narrow belt that shows predominantly a variation in recrystallized grain size with increasing metamorphic grade. Assemblages in the micaceous marbles are similar to those in the calc-schists, and are not described separately.

Metamorphism of amphibolite

The distribution of amphibolite is restricted to the slices of Precambrian basement rock, mafic bands of the Lady Slipper Pluton, and isolated mafic dykes in the South Brook Formation. Most outcrops of amphibolite are in the garnet zone, and a few are in the staurolite and kyanite zone east of Corner Brook Lake. The mineralogy of the amphibolites is very consistent and the main variation is in relative abundances of minerals, grain size, and intensity of retrogression. All amphibolites contain hornblende, plagioclase, quartz, epidote, and titanite. In some rocks, biotite is abundant, magnetite is commonly present, and in few rocks chlorite or dolomite occurs. A few samples contain garnet.

Hornblende forms stubby or elongate subhedral porphyroblasts up to 0.5 cm long. They have bluish-green – dark green – yellow pleochroism. Plagioclase is commonly a matrix mineral, but locally occurs as porphyroblasts (Fig. 8D) with amphibole and epidote inclusions. Plagioclase is rarely sericitized. Titanite is fine grained, subhedral to euhedral, and is dispersed in the matrix in most rocks, but locally occurs as relatively large (up to 2 mm long) elliptical grains. Quartz is common but occurs in minor amounts in the matrix. Epidote is a minor phase in most samples and occurs as small subhedral grains in the matrix. Rarely it forms larger porphyroblasts. Biotite is common but does not occur in all samples. It is either brown or brownish green and usually forms large subhedral porphyroblasts that overgrow S_1 in random orientation, but is aligned with the hornblende where S_2 forms the dominant foliation. Chlorite is rare apart from local pseudomorphic replacement of biotite. Magnetite forms small euhedral grains in most rocks. Dolomite was observed as euhedral crystals in the matrix in few samples. Garnet occurs in two samples from basement amphibolite at higher metamorphic grades (staurolite zone and kyanite zone). These samples are quartz-rich and contain moderate amounts of biotite. Garnet is anhedral and is possibly a relict of a previous metamorphic event. A quartz-rich amphibolite from the small basement slice in the north of the map area, contains pale green and blue actinolite, rather than hornblende. The sample contains plagioclase with abundant sericite inclusions, abundant green biotite, and minor chlorite, epidote, magnetite, and titanite. This rock from the western edge of the garnet zone, is of greenschist grade, instead of the amphibolite grade of the remainder of the amphibolite studied.

Metamorphism of calc-silicate rocks

Calc-schist, micaceous marble, thin-bedded limestone, and calc-slate are abundant in the western part of the map area. They appear in the Pinchgut, Reluctant Head, and Breeches Pond formations across all metamorphic zones. An added complexity of the reactions involving calc-silicate is the dependence on the activity of H_2O and CO_2 and the mobility or immobility of the fluid phase (Greenwood, 1975). This can make calc-silicate assemblages unreliable to use as P-T indicators. Furthermore, these rocks have a wide variation in composition, depending on the original relative abundance of quartz, carbonate, and silicate minerals.

In the map area, a general trend was observed in the variation of mineral assemblages, which is consistent with that in the pelites. No detailed petrographic study was carried out, and with the exception of the amphibole isograd, which is mainly defined from field observations, no isograds were mapped. Below is a short description of the main assemblages observed in each of the metamorphic zones defined from the pelitic assemblages.

In the muscovite-chlorite zone in the Humber Arm Allochthon and the Carbonate Belt, the calc-schists have a constant assemblage of quartz-calcite/dolomite-sericite-chlorite-plagioclase. Siderite and opaque oxides, predominantly pyrite, occur locally. The most westerly sample was from the town of Corner Brook, which is within the zone where plagioclase occurs in the pelites. All phases are fairly fine grained, but an overall increase of grain size to the east was observed. In most of the biotite zone, the same assemblage is dominant. In one sample from east of Pinchgut Lake, phlogopite occurs as small pale, yellowish-brown to yellow pleochroic grains. Near the garnet isograd in the east of the biotite zone, zoisite first appears as small grains in the matrix. Other samples from the eastern part of the biotite zone lack chlorite. The latter assemblage, calcite/dolomite-quartz-white mica-plagioclase is common at this grade and is found locally up to the highest grade area.

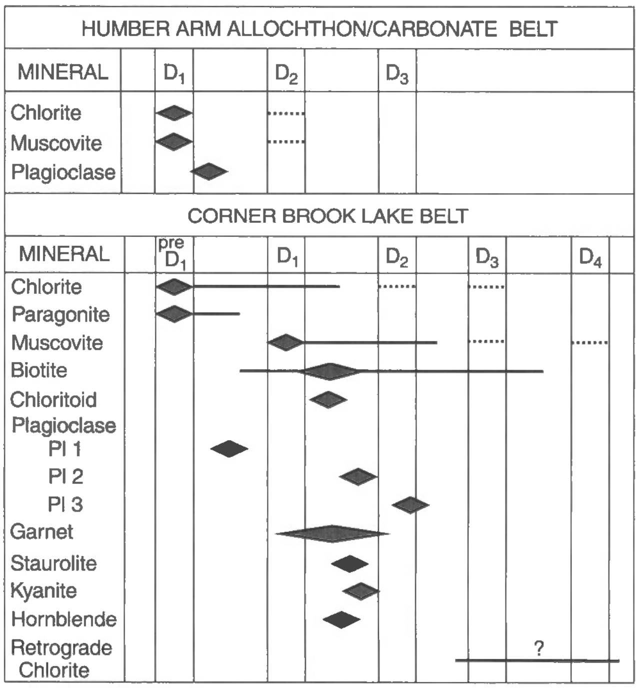


Figure 53. *Relative timing of growth of porphyroblasts and phases of deformation in the pelitic and psammitic rocks of the Humber Zone. Diamonds indicate main periods of porphyroblast growth, solid lines indicate periods of minor growth, and dashed lines indicate localized, strain-induced recrystallization.*

Garnet appears in the garnet zone in calc-schist with a predominantly pelitic composition. About 1 km east of the garnet isograd, amphibole first occurs in garbenschiefer with large hornblende crystals that form bow-tie structures (Fig. 12B). The hornblende overgrows S_1 (Fig. 52B). The bow tie structures indicate a static growth in a low stress regime. S_2 postdates the hornblende and wraps around the porphyroblasts. Pseudomorphic replacement of the hornblende by biotite or chlorite is common, especially near the hornblende isograd. Hornblende occurs in garbenschiefer commonly together with garnet, which has the same relative timing as the hornblende. These rocks contain muscovite+hornblende+plagioclase+quartz+zoisite+calcite/dolomite±biotite±ilmenite±pyrite±chlorite (retrograde). In rocks that lack hornblende, phlogopite is commonly abundant. Plagioclase can form post-S_1 porphyroblasts.

At the highest grades, a wide variety of phases starts to occur. Biotite is common in the calc-silicate, and white mica is rare. Other phases that were observed in the field include calcite, quartz, hornblende, tremolite, wollastonite, zoisite, and some of the rocks contain talc, judging from a greasy feel. Zoisite locally forms prisms up to several centimetres long. In samples from east of Corner Brook Lake, the following assemblages were observed: calcite/dolomite+biotite+talc+plagioclase+muscovite + quartz + minor epidote+minor apatite; biotite+calcite+plagioclase+quartz+talc+wollastonite; hornblende, phlogopite, calcite/dolomite+zoisite+plagioclase+quartz+diopside+K-feldspar+opaque oxides.

Relative timing of metamorphism and deformation in the Humber Zone

By studying the microstructural relationships between porphyroblasts, structures in the matrix, inclusion patterns in the porphyroblasts, and overgrowth relationships between different minerals, the relative timing of the metamorphic and structural events was determined. This was done for the Humber Zone rocks, specifically the Corner Brook Lake Belt, since the medium- to high-grade metapelites with their well developed foliations and folds and large porphyroblasts lend themselves best for such a study. All of the described observations are from thin sections, except where indicated otherwise. The relationships between generations of structures and porphyroblasts are summarized in Figure 53 and are interpreted as a relative timing of the phases of deformation and metamorphism. It shows that the peak of metamorphism occurred between D_1 and D_2.

Humber Arm Allochthon and Carbonate Belt

White mica, chlorite, and plagioclase are the only metamorphic minerals in these two lithotectonic units, and they are all very fine grained, which hampers an exact determination of the growth-deformation relationships. In slate of the Summerside, Irishtown, and Reluctant Head formations, and in the sandstone in the Irishtown and Summerside formations, the white mica and chlorite define the S_1 foliation. The mica is crenulated and partially recrystallized during D_2 to form a new S_2 crenulation cleavage, but no significant new growth of micas occurred during or after the crenulation. In one sample from east of Pinchgut Lake, plagioclase has an internal fabric of quartz and mica inclusions. Plagioclase is assumed to have started growing later than white mica and chlorite.

Although the exact timing of the metamorphic peak with respect to D_1 is not obvious in these low-grade and fine-grained rocks, the described relationship indicates that the growth of metamorphic minerals occurred during and after D_1 and pre-D_2. This puts the metamorphic peak syn- to post-D_1.

Corner Brook Lake Belt

The oldest structures recognized in the Corner Brook Lake Belt are the brittle thrusts, which predated the Paleozoic metamorphic events. Fine-scale intergrowths of chlorite and white mica define the pre-S_1 schistosity in the few samples where it is preserved (Fig. 51A, B). Microprobe analysis of mica in a sample from near the staurolite-in isograd, indicated that the fine-grained white mica consists of paragonite (Table 4). It is assumed that throughout the Corner Brook Lake Belt the earliest white mica was paragonite, and it is labelled as such in Figure 53. The onset of regional metamorphism must have occurred during or prior to this pre-D_1 deformation event. Larger white mica, presumably muscovite (Table 4), overgrows this earliest foliation, either statically (post S_1), forming locally a decussate texture, or dynamically, forming a new recrystallized S_1 crenulation cleavage (Fig. 31A, 51B), or a foliation of isolated crystals with a preferred orientation (Fig. 51A). The new growth of muscovite started at the latest during D_1, but similar textures of newly grown muscovite along S_2 were also observed. Strain induced recrystallization in the hinges of F_2, F_3, and F_4 microfolds is common. Recrystallization of mica and quartz is pervasive in F_2 folds (Fig. 34C), less common in F_3 folds (Fig. 36D, E), and rare in F_4 folds (Fig. 37D). The recrystallization also diminished towards the lower metamorphic grade rocks. Figure 36D is a photomicrograph of an F_3 microfold, showing evidence of strain-induced recrystallization. Muscovite has overgrown the hinge in the core of the fold, where the fold is tight, but in the more open outer arc, most mica is bent and shows only minor recrystallization. This indicates a decrease of the metamorphic grade after D_2, with conditions being barely high enough to allow minor grain boundary migration in mica during D_4.

The main growth phase of the other porphyroblasts occurred between D_1 and D_2. Figure 52C show microstructural relationships of garnet and albite overgrowing a straight S_1 (c.f. Fig. 51E). Where the internal S_1 fabric in the porphyroblasts is straight, it is commonly at an angle to the external fabric as a result of deformation (D_2 or later) after the growth of the porphyroblast (Fig. 51E). Poikiloblastic kyanite and staurolite show similar relations with respect to S_1 and S_2. Albite outlasted the growth of the other minerals which locally can be seen by the inclusion of garnet, kyanite, and staurolite in the albite (Fig. 9F). Biotite grows predominantly between D_1 and D_2 (Fig. 51D), but younger porphyroblasts also occur. Hornblende in the calc-schist of the Breeches Pond Formation shows a similar straight internal foliation formed by S_1, which is deformed by D_2 in the matrix (Fig. 52B). In rocks where an S_2 crenulation cleavage is developed, it is commonly at an angle to the internal fabric in the porphyroblasts (S_1), and where D_2 strain is high, it wraps around the porphyroblasts. At several locations these relationships could also be observed in the field. Where large garnets have an internal fabric, it is either continuous with the S_1 fabric in the matrix (Fig. 9D), or it is consistently oriented through parts of an outcrop, at a high angle to the S_2 crenulation cleavage in the matrix (Fig. 54). Figure 55 shows the sequence of developments that caused these relationships that were observed on mesoscopic and microscopic scales. In some samples F_1 microfolds are preserved in the inclusion trails in garnet (Fig. 52D).

Plagioclase shows a complex history of growth, and three phases are recognized and numbered Pl1, Pl2 and Pl3. Plagioclase growth is assumed to have started before, but to have outlasted the growth of most other minerals. In the northern part of the Corner Brook Lake Belt, approximately north of Steady Brook, Pl1 porphyroblasts predate garnet. In samples from this part of the area, plagioclase overgrows the fine-grained pre-S_1 foliation, but predates the S_1 (Fig. 51C). In

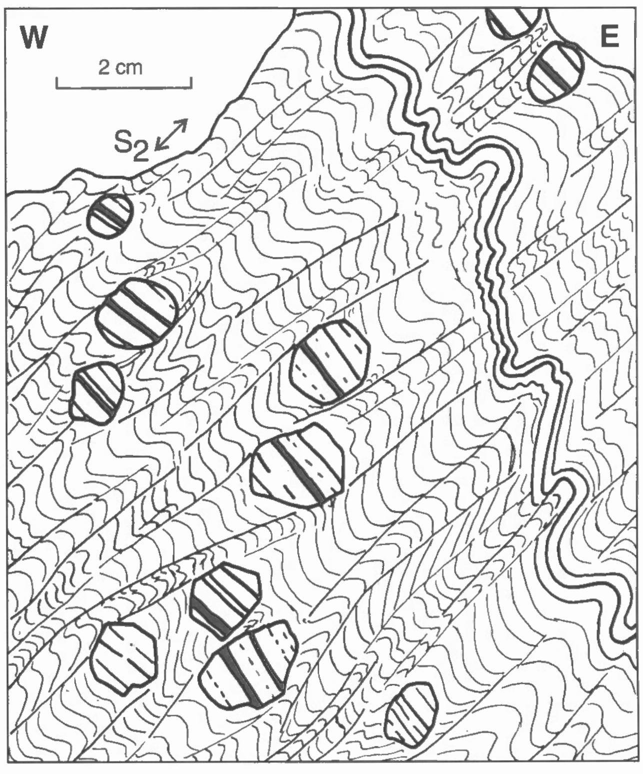


Figure 54. *Field sketch of garnets with an internal S_1 fabric, which is at a high angle to the S_2 crenulation cleavage in the matrix. In parts of the outcrop where D_2 strain is lower, the original orientation of the S_1 is preserved in the matrix as well. Garnet-muscovite schist of the South Brook Formation, north of Grand Lake Brook (g.r. 308007).*

other samples Pl1 porphyroblasts are included in garnet. In Figure 56 a garnet contains plagioclase porphyroblasts (Pl1) that have a straight, consistently oriented internal foliation. The inclusion trails in the garnet are folded in F_1 crenulations which postdate the included plagioclase porphyroblasts. The main foliation in the matrix is an S_2 crenulation cleavage, which is at an angle to the axial plane of the crenulations in the garnet and wraps around the garnet.

Pl2 is the most abundant plagioclase population in the area and consists of albite. In the psammite in the core of the Yellow Marsh antiform it is the dominant mineral (Fig. 9E), but it also occurs in pelite throughout the area. It overgrows garnet, muscovite, biotite, and zoisite/epidote (Fig. 51E). In the pelite, the Pl2 porphyroblasts have a fine-grained internal foliation defined by usually straight inclusion trail of graphite, zoisite/epidote, quartz, or fine-grained white mica (Fig. 52C). Where developed, the S_2 in the matrix wraps around the Pl2 porphyroblasts and is at an angle to the internal foliation (Fig. 51E).

The latest growth of plagioclase (Pl3) consists mainly of overgrowth rims on albite and rarely of newly formed grains. The rims are locally visible in hand specimen. Pl3 is more calcic than Pl2, having an albite to oligoclase composition. It occurs predominantly in the pelitic rock types in the higher grade metamorphic zones (garnet grade and up). Pl3 porphyroblasts locally have overgrown F_2 crenulations.

Biotite also has a complex growth history. In some rocks, biotite is part of the matrix minerals that form S_1, but more commonly biotite forms porphyroblasts that postdate the S_1 foliation, as shown in the example in Figure 51D. With very few exceptions, the biotite porphyroblasts are affected by F_2 folds and therefore predate D_2, but post-D_2 biotite does also occur.

Chlorite shows new growth during the pre-D_1 deformation as well as recrystallization during D_1, forming part of the recrystallized S_1 differentiated layering or crenulation cleavage. In the garnet zone and higher metamorphic zones, chlorite was subsequently broken down. During retrograde metamorphism, chlorite replaced biotite and garnet. Chlorite was observed as an overgrowth of an F_4 crenulation, indicating that at least part of the retrograde chlorite growth took place after D_4.

Most of the larger porphyroblasts have grown during static periods, and the majority do not show variations of the inclusion pattern within one crystal that could suggest deformation during growth. However, locally overlaps exist

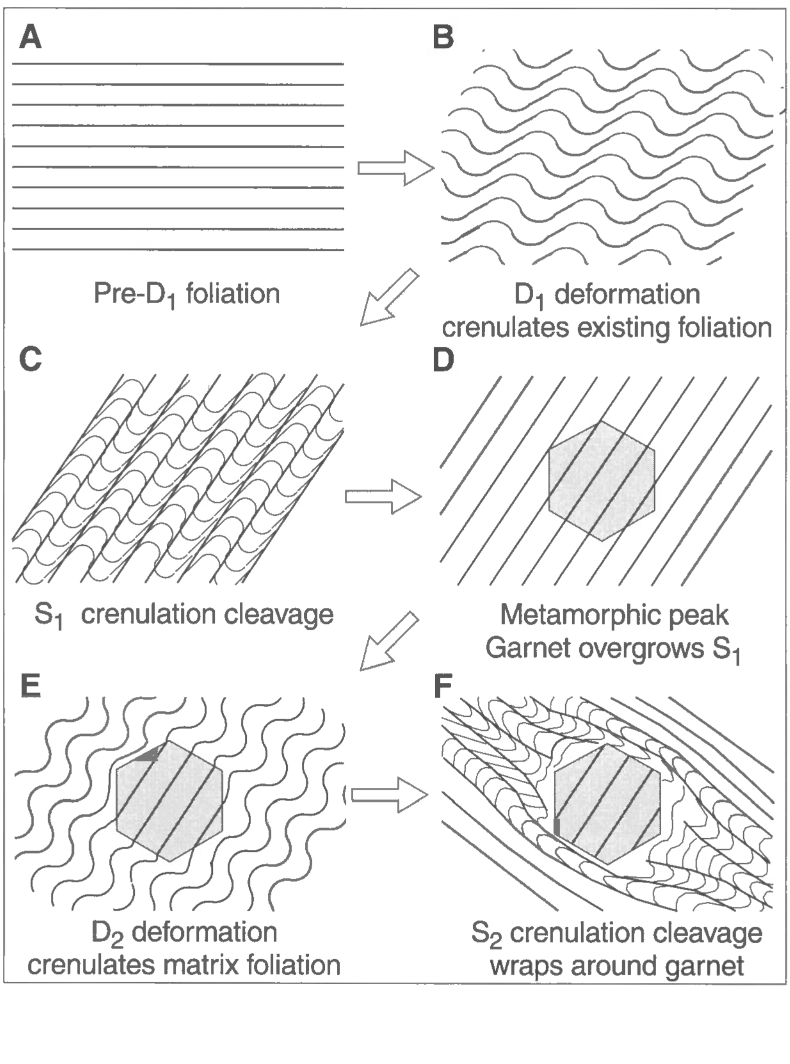


Figure 55.

Development from A to F of the observed inclusion patterns and matrix foliations on a microscopic scale.

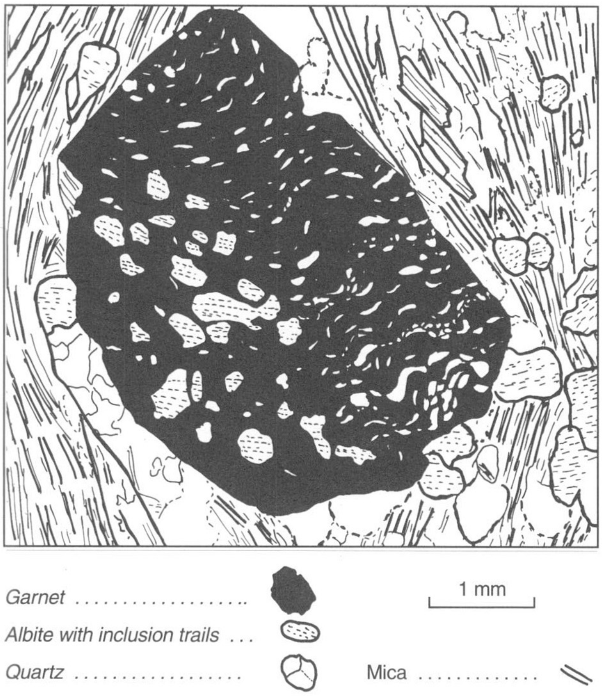


Figure 56. *Sketch of a garnet porphyroblast, in thin section, containing albite inclusions with straight inclusion trails. Quartz inclusions in the garnet outline F_1 crenulations that postdate the included albite. Garnetiferous psammite of the South Brook Formation at Steady Brook (g.r. 434205).*

between metamorphic mineral growth and D_1 or D_2. Spiral inclusion trails within both garnet and albite (Pl2, seldom Pl3) suggest that they grew while rotating during simple shear deformation (Fig. 32C). Most of the rotated garnet and albite (Pl2) grew during D_1, but a few porphyroblasts appear to have grown during D_2 deformation, suggesting that D_2 also has a component of non-coaxial flow.

All these observations show an overall static growth of the main metamorphic minerals between D_1 and D_2 in a specific order, as is outlined in Figure 53. Local deviations from this sequence of events show that deformation and mineral growth did not occur homogeneously throughout the area. Some of these variations can be explained by differences in bulk rock compositions, which locally caused certain minerals to grow at different times. However, other differences are presumed to have been caused by small variations in the P-T paths, e.g. differences between the hanging wall and foot wall of thrust faults, or by inhomogeneous deformation and moving deformation fronts, which localized the strain in parts of the belt at different times. Detailed studies are required to better define and understand these variations.

To summarize: the metamorphic grade in the area varied considerably during the different phases of deformation, as schematically presented in Figure 57. After the earliest brittle thrusting, the metamorphic grade started increasing and resulted in the growth of white mica and chlorite during the pre-D_1 deformation, suggesting low greenschist facies conditions. D_1 occurred when in the high grade part of the area garnet started growing at upper greenschist facies conditions, but farther west the rocks were still in lower greenschist facies during D_1. The final growth stage of the porphyroblasts presumably coincides with the (temperature) peak of metamorphism. This was closely followed by D_2, D_3, and D_4 under retrograde conditions. Temperatures were barely high enough during D_4 to enable strain induced recrystallization of mica in F_4 crenulations.

The observations described here suggest that the metamorphism in the Humber Zone was caused by the early brittle thrust stacking and subsequent ductile thrusting. The rise of the metamorphic grade after initial thrusting and peaking after D_1, as well as the geometry of the metamorphic zones in belts parallel to the traces of the thrust surfaces, the culmination of the metamorphism at the base of the largest thrust sheet in the Humber Zone, and the medium pressure metamorphic conditions, all indicate that the metamorphism was caused by the thrusting, rather than by a local heat source. The northwest-directed thrusting progressively buried the rocks deeper and became progressively more ductile and less concentrated in narrow zones with increasing metamorphic grade. West-directed thrusting continued after D_1 and after the peak of the metamorphic event, but deformation occurred predominantly as folding during D_2 and D_3. However, D_2 locally caused out-of-sequence thrusting, which resulted in considerable telescoping of the metamorphic gradient in the footwall of the Yellow Marsh fold nappe, specifically north of Corner Brook Lake. In spite of the D_2 and D_3 deformation causing shortening/thickening of the crust, the rocks were exhumed, as indicated by the decrease of metamorphic grade. Finally, the rocks were uplifted into a zone of extensional

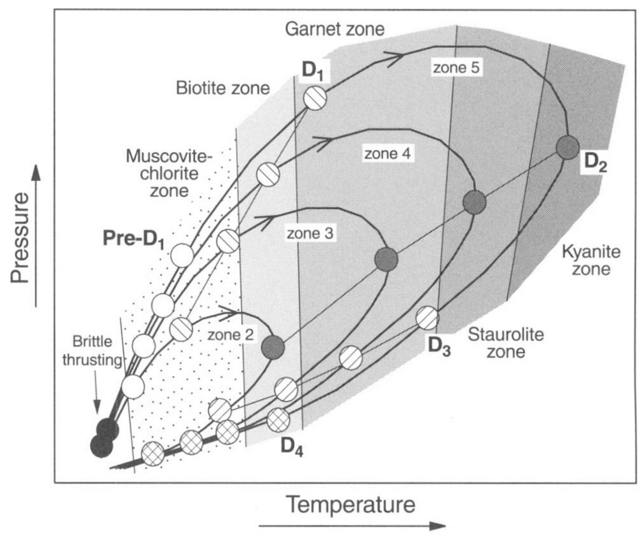


Figure 57. *Schematic P-T loops for rocks in the four highest metamorphic zones in the Humber Zone. The figure shows the relationships between P-T variations and the different phases of deformation in the various parts of the Corner Brook Lake Belt. Circles represent deformational events, and the thin lines connecting them are approximate time lines.*

deformation which, in the western lowest grade part of the map area, resulted in extensional faulting (Humber River and Hughes Brook faults and many small scale faults in the Humber Arm Allochthon). In the Corner Brook Lake Belt, where metamorphic grades were still slightly higher, the extensional deformation was expressed as recumbent folding during D_4.

Metamorphism of the Glover Island Belt

The deformation and metamorphism in the Glover Island Belt are less intense than in the Humber Zone, and the metamorphic minerals and assemblages relate a less complex history. No detailed petrographical study was done on the rocks of the Glover Island Belt, and our data were supplemented with those of Knapp (1982). No petrographic data are available for the rocks east of Grand Lake.

Metamorphic mineral assemblages and textural data indicate that the rocks from the Dunnage Zone were affected by a single regional metamorphic event. The metamorphism resulted in the development of a greenschist facies mineralogy in the Grand Lake Complex and Glover Formation which, in the former, retrogressed an early phase of amphibolite facies metamorphism, assumed to have occurred while the rocks formed part of an ocean floor. In the Kettle Pond shear zone, lower greenschist facies conditions dominated during the main deformation event. Near the margins of the Glover Island granodiorite, minor greenschist facies retrogression occurred.

Grand Lake Complex

In mafic gabbro of the Grand Lake Complex, the oldest foliation is formed by a weak preferred orientation of hornblende. Besides hornblende, plagioclase, epidote, oxides, titanite, and minor quartz occur, and locally the hornblende replaces original igneous clinopyroxene. Knapp (1982) identified the amphiboles as pargasitic hornblende and interpreted this to indicate a high temperature metamorphic event of amphibolite grade, which occurred while the gabbros formed part of an ocean floor (that is prior to obduction). In the ultramafic rocks, the original igneous minerals and textures are overall well preserved. An early foliation in the ultramafic rocks defined by antigorite is also interpreted to be related to events predating obduction of the ophiolite. Later growth of antigorite defines a younger foliation and together with a partial re-equilibration of igneous pyroxene to metamorphic composition, is interpreted to represent upper greenschist conditions (Knapp, 1982). This metamorphic and structural imprint on the rocks is weak, and is only locally observed.

During the main shearing event, which produced a network of narrow shear zones in the rocks of the Grand Lake Complex, the amphibolite facies rocks were overprinted by lower greenschist grade retrogression. The shearing resulted in replacement of the amphibole by chlorite and epidote, and locally by calcite/dolomite. Igneous plagioclase is pseudomorphically replaced by albite and clinozoisite. In the ultramafic rocks the increase of fluid migration during the shearing caused talc-carbonate metasomatism. The greenschist of the Grand Lake Complex contains an assemblage of chlorite, epidote, plagioclase, carbonate, and opaque oxides. These rocks are assumed to be intensely deformed, metasomatized and totally retrogressed mafic igneous rocks, but no relict high grade minerals remain (Knapp, 1982).

Glover Formation

The metamorphic mineral assemblages in the rocks of the Glover Formation are fairly consistent, in spite of the variety of rock types that occur. All mafic rocks contain the assemblage actinolite+plagioclase+zoisite/epidote+opaque oxides +titanite, with calcite from amygdules occurring in the basalt, and rare chlorite or biotite. This assemblage is typical for the greenschist facies in mafic rocks. Acicular blue-green – green – yellow pleochroic actinolite and plagioclase form up to 60 and 30% respectively, of most rocks, and are fine grained. The amphibole porphyroblasts generally have a weak preferred orientation which defines a foliation plane, suggesting that they grew during deformation. Plagioclase occurs in the fine-grained matrix and as relict phenocrysts. The latter contain many tiny actinolite and zoisite inclusions. Epidote occurs predominantly as a fine grained part of the matrix, but can occur as relatively large porphyroblasts.

Near the western boundary of the Glover Formation, Knapp (1982) attributed the occurrence of chlorite-rich assemblages, which lacked actinolite, but contained quartz, sericite, and calcite, to carbonate metasomatism that also caused the occurrence of carbonate-talc assemblages in the ultramafic rocks of the Grand Lake Complex. This change in assemblage towards the west is assumed to be related to retrograde metamorphism and metasomatism during the deformation, and increased fluid flow in the Kettle Pond shear zone.

Kettle Pond shear zone

All rocks in the Kettle Pond shear zone were synkinematically retrogressed to lower greenschist facies mineral assemblages. Deformation in the shear zone caused a lower grade overprint of the amphibolite and greenschist grade mineral compositions of the two adjoining lithological units, and caused a significant grain size reduction. All mafic rocks have a fine-grained matrix of chlorite and plagioclase, which usually occurs with epidote, opaque oxides, and minor carbonate. Quartz and sericite can occur as minor phases, and a few samples contain abundant actinolite. Where relict plagioclase is preserved as porphyroclasts, it is to a large extend replaced by fine-grained zoisite. The preferred orientation of the chlorite defines the main foliation. In low strain zones the original mineralogy can be partially preserved. In a sample from the western side of the shear zone, olive-green – brown – yellow pleochroic hornblende is overgrown by bluish-green – pale-green – pale yellow actinolite.

Rocks of felsic composition are mainly derived from the trondhjemite of the Grand Lake Complex and from minor felsic or silicified volcanic and/or volcaniclastic rocks. The strongly foliated matrix consists predominantly of quartz and aligned sericite and chlorite, with minor plagioclase, siderite, and opaque oxides. Fuchsite occurs at several locations in the shear zone, indicating high chromium concentrations. In one sample, pyrite overgrows the foliation. Observations of both brittle and ductile deformation features might indicate that deformation occurred at different metamorphic grades, ranging from possibly actinolite stability conditions (middle-upper greenschist) down to chlorite (lower greenschist) conditions, and the lower possibly sub-greenschist facies. This suggests a protracted deformation history of the shear zone.

Glover Island Granodiorite

Metamorphic effects in the Glover Island Granodiorite were only observed in the zones that were affected by moderate deformation. In the core of the pluton, where no effects of deformation are visible on outcrop scale, minor sericite and zoisite occur in plagioclase, and rims of chlorite surround biotite. At the edge of the pluton, the rocks are more deformed, resulting in flattening of the quartz and a weak preferred orientation of the mica. Sericitization of plagioclase is intense, and locally the plagioclase is partially broken down to a quartz-plagioclase-sericite matrix. Biotite is replaced by chlorite, and zoisite/epidote is common. This deformation occurred below the biotite stability, at lower greenschist facies conditions. A weak igneous mineral alignment in the pluton may suggest that it was intruded into a stress field during a tectonic deformational event. No thermal effect of the intrusion on the wallrock has been recognized.

Relative timing of metamorphism and deformation

The metamorphic signature of the rocks in the Glover Island Belt is overall of greenschist grade, but each lithological unit in the belt (Fig. 49) shows a distinct history. The amphibolite facies conditions in the Grand Lake Complex are assumed to predate the regional metamorphism. The antigorite and partial re-equilibration of pyroxenes in this western metamorphic zone represent a minor retrogression to upper greenschist conditions. It is separated from the prograde actinolite greenschist zone in the Glover Formation by the Kettle Pond shear zone at lower greenschist grade. The fact that antigorite forms a foliation in the Grand Lake Complex, and the alignment of actinolite in the Glover Formation, are interpreted to indicate that these minerals grew during deformation, suggesting that middle to upper greenschist facies conditions prevailed during the earliest deformation. The dominant mylonitic foliation in the Kettle Pond shear zone, defined by chlorite, but with remnants of higher grade conditions, postdated the metamorphic peak and the weaker foliations in the adjoining units. This means that the Grand Lake Complex and Glover Formation were juxtaposed after the peak of metamorphism. Although both the peak metamorphic foliation and the shear-related foliation were mapped as S_1 in the field, petrographic relations indicate that they did not form at the same time. They might be related to one progressively developing phase of deformation, as suggested by their parallel orientation. All foliation development in the area could be related to the protracted history of the Kettle Pond shear zone, which might have been active under peak metamorphic to subgreenschist conditions, thus preserving the chlorite assemblages.

The Glover Island Granodiorite contains foliated inclusions of the Glover Formation and was intruded after the onset of foliation development. It was deformed under metamorphic conditions that were similar to those prevailing in the Kettle Pond shear zone, suggesting that intrusion took place before the retrograde chlorite conditions were reached.

No significant new growth of minerals was observed in D_2 and D_3 structures. Locally, chlorite or sericite occurred along crenulation cleavage planes, where they grew as a new retrograde phase, or recrystallized dynamically. At the time of D_2 and D_3 folding, the metamorphic grade was estimated to have subsided to lower greenschist or subgreenschist grade.

These observations tentatively result in a tectonic history which postdates the amphibolite grade, metamorphism of the Grand Lake Complex. The onset of deformation occurred near the metamorphic peak, at upper greenschist facies conditions. Deformation outlasted the peak metamorphic conditions, which were followed by the intrusion of the Glover Island granodiorite. The final juxtaposition of the units along the Kettle Pond shear zone occurred at lower grade conditions, which, outside the shear zone, had only local effect on the wall rocks and caused minor deformation and retrogression of the granodiorite.

Thermal maturation in the Deer Lake Basin

Hyde et al. (1988) studied the effects of the Carboniferous thermal event in the Deer Lake Basin. They compared the data from vitrinite reflectance, clay mineral assemblages, and illite crystallinity. The results of their study, which apply to the whole Deer Lake Basin, are summarized in this section.

Hyde et al. (1988) determined vitrinite reflectance measurements from kerogen contained within samples of the Anguille and Deer Lake groups. The mean maximum reflectance values ($\overline{R}_{omax}$) from rocks of the Anguille Group around Grand Lake (mainly north of the map area) ranged from 0.75 to 1.91%, with all but two anomalous values being above 1.16%. Values for the Deer Lake Group are distinctly lower, with ($\overline{R}_{omax}$) ranging from 0.59 to 1.07%, indicating a lower maturation level. The conversion of these values to temperatures is not straightforward, because besides heat, time of burial and pressure are controlling factors. Several methods can be used for the estimation of the temperature, all of which indicate that the temperatures in the Anguille Group were about 100°C higher than in the Deer Lake Group. The average values for the temperature estimates are about 200°C for the Anguille Group and about 100°C in the Deer Lake Group.

Clay mineral assemblages in sandstone and clay of the Anguille Group are dominated by illite and Fe-rich chlorite, with minor amounts of kaolinite. The abundance of illite and chlorite suggests a 'high-rank' diagenetic assemblage (Hyde et al., 1988). Assemblages in the Rocky Brook Formation of the Deer Lake Group are more diverse. The samples contain illite and, to a lesser extent, chlorite, but also show the presence of smectite and a mixed-layer chlorite-smectite. In these rocks especially, the presence of smectite indicates a 'low--rank' assemblage in which transformation to anhydrous clays has not been completed. The transition from mixed-layer clays to a stable illite-chlorite assemblage is estimated at about 200°C, providing a minimum temperature for the rocks of the Anguille Group. Smectite is transformed to inter-layered illite-smectite at about 100°C, which is also the temperature given as the lower stability of mixed-layer chlorite-smectite. This gives a narrow temperature bracket for the Rocky Brook Formation around 100°C.

Illite crystallinity measurements by Hyde et al. (1988) on the same samples resulted in an average crystallinity index value of 0.53°$\Delta 2\theta$ for the Anguille Group and 0.68°$\Delta 2\theta$ for the Rocky Brook Formation. With standard deviations of 0.10° and 0.12°$\Delta 2\theta$, respectively, there is a large overlap of the measurements and the difference is minimal, but statistically significant. Samples collected near faults consistently showed lower values.

All these data are in agreement with a higher maturation level for the rocks of the Anguille Group with respect to the rocks of the Deer Lake Group. Approximate diagenetic temperatures are estimated to be around 200°C for the former and 100°C for the latter. Hyde et al. (1988) interpreted these data to most likely reflect deeper burial of the stratigraphically lower rocks of the Anguille Group, but also suggested that higher heat flow in the earlier stages of the development of the basin could have played a role.

DATING OF METAMORPHISM AND DEFORMATION

Humber Zone

U-Pb and Ar-Ar data for a number of metamorphic mineral phases, plus U-Pb data for the age of crystallization of the unnamed granite pegmatite body at Corner Brook Lake, provide important constraints on the timing of metamorphism and deformation within the Humber Zone of the Corner Brook Lake region.

Coarse-grained to pegmatitic granitic rocks locally occur within the high grade metamorphic segments of the thrust belt. These granitoid rocks are variably deformed with the coarsest grained pegmatitic phases showing only a weak S_1 foliation, whereas the finer grained phases are in places moderately to strongly deformed (Fig. 17B). This difference in strain could be an effect of either the later timing of intrusion of the pegmatitic phase, suggesting that the granitoid body as a whole intruded during deformation, or of a competency contrast between granite and pegmatite. The pegmatitic phase locally created strain shadows in which the finer-grained phase was less deformed, indicating that the intrusion predated at least part of the D_1 deformation.

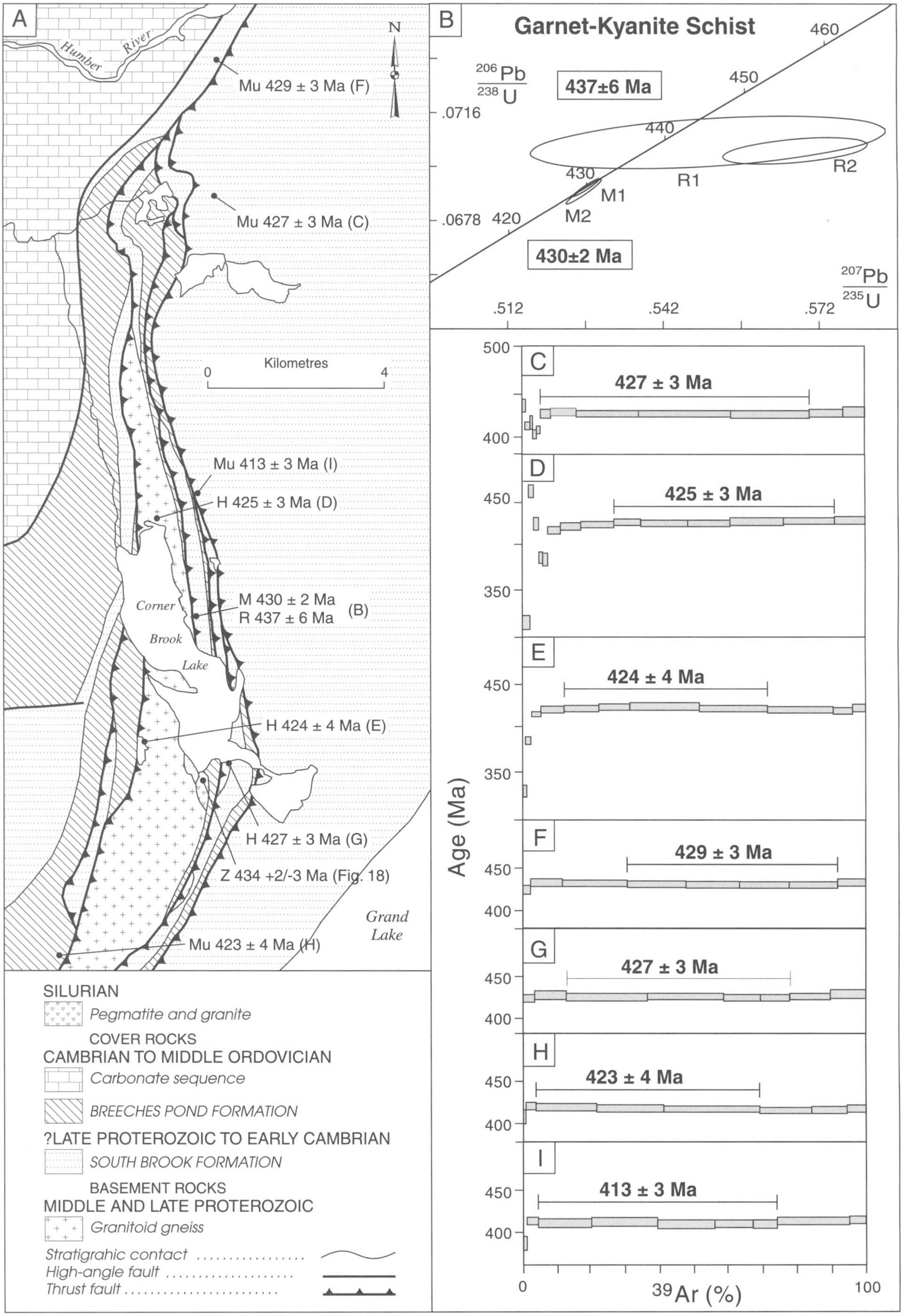
The metamorphic mineral phases monazite, zircon, rutile, hornblende, and muscovite were separated from a variety of basement and cover lithologies for isotopic dating. Most samples were collected from lithologies which had undergone upper greenschist or amphibolite facies metamorphism. Figure 58A shows the location and age of the samples, while Figures 58B and 58C to I respectively show the U-Pb concordia and Ar-Ar release spectra. Table 1 contains details of the U-Pb analytical data. Attempts at Ar-Ar biotite dating of samples from this region were unsuccessful.

Zircon from a pegmatite intrusion into the basement-cover sequence defines an age of crystallization of 434 +2/-3 Ma (Fig. 18). A sample of garnet-kyanite-staurolite schist from the South Brook Formation east of Corner Brook Lake, yielded high quality coarse-grained monazite and rutile, both as blocky crystals and needles. The monazite was very high in uranium content and the rutile very low. Duplicated concordant monazite fractions yielded a precise U-Pb age of 430 ± 2 Ma. Duplicated rutile yielded ^{206}Pb-^{238}U ages of 439 ± 10 Ma (fraction R1, Fig. 58B; Table 1) and 437 ± 6 Ma (fraction R2). The latter, better analysis is taken as the age of cooling, through the interpreted blocking temperature of rutile of ~400°C (Mezger et al., 1989). The rutile analyses are less well constrained than the monazite data because of their high common Pb content. Hornblende from amphibolite at three localities within the basement sequence yielded Ar-Ar plateau ages (59% to 78% of gas released) of 427 ± 3 Ma, 425 ± 3 Ma, and 424 ± 4 Ma (Fig. 58C-E). Muscovite from four samples of psammitic and pelitic schist from the cover sequence gave Ar-Ar plateau ages (61% to 70% of gas released) of 429 ± 3 Ma, 427 ± 3 Ma, 423 ± 4 Ma, and 413 ± 3 Ma (Fig. 58F-I).

K-Ar dating of muscovite from schists within the South Brook Formation near Steady Brook yielded ages of 420 ± 14 Ma and 437 ± 14 Ma (Wanless et al., 1973, recalculated to new constants; Cawood, 1993). Wanless et al. (1965) determined a K-Ar age for biotite from a granitoid dyke from the shore along the southwest arm of Grand Lake of 428 ± 20 Ma (recalculated to new constants). According to Kennedy (1981) emplacement of the dyke coincided with the last stages of peak metamorphic conditions. This dyke may be related to the unnamed pegmatite/granite recognized around Corner Brook Lake, but it is uncertain if the age determined by Wanless et al. (1965) is that of emplacement of the dyke and/or a post-metamorphic cooling age.

Discussion

The isotopic age data from the Corner Brook Lake region provide a remarkably uniform data set indicating that regional deformation, metamorphic mineral growth, and cooling within this segment of the eastern Humber Zone occurred in the early Silurian between 435-425 Ma. An older age limit on regional deformation in the area is provided by the principal foliation (S_1) which deforms the pegmatite, and hence can be no older than 436 Ma. The monazite age of 430 ± 2 Ma from the garnet-kyanite-staurolite schist records peak metamorphic conditions, and this assemblage postdates the principal foliation (Fig. 9D), but predates later penetrative fabrics (S_2 and S_3; Fig. 53, 54). A younger age limit on deformation is provided by the muscovite data. The muscovite recrystallized during the D_3 and D_4 events and, along with rutile and hornblende, gives a range of cooling ages, the bulk of which are around 430-420 Ma. One muscovite sample from a psammite in the Corner Brook Lake region gave a younger age of around 413 Ma. The reason for this younger age is uncertain, but relative to the other muscovite samples, it is located in an


A
Humber River
N
Mu 429 ± 3 Ma (F)
Mu 427 ± 3 Ma (C)
Kilometres
0
4
Mu 413 ± 3 Ma (I)
H 425 ± 3 Ma (D)
Corner Brook Lake
M 430 ± 2 Ma
R 437 ± 6 Ma
(B)
H 424 ± 4 Ma (E)
H 427 ± 3 Ma (G)
Z 434 +2/-3 Ma (Fig. 18)
Grand Lake
Mu 423 ± 4 Ma (H)
SILURIAN
Pegmatite and granite
COVER ROCKS
CAMBRIAN TO MIDDLE ORDOVICIAN
Carbonate sequence
BREECHES POND FORMATION
?LATE PROTEROZOIC TO EARLY CAMBRIAN
SOUTH BROOK FORMATION
BASEMENT ROCKS
MIDDLE AND LATE PROTEROZOIC
Granitoid gneiss
Stratigrahic contact
High-angle fault
Thrust fault
B
Garnet-Kyanite Schist
206Pb/238U
207Pb/235U
437±6 Ma
430±2 Ma
.0716
.0678
.512
.542
.572
420
430
440
450
460
R1
R2
M1
M2
C
427 ± 3 Ma
D
425 ± 3 Ma
E
424 ± 4 Ma
F
429 ± 3 Ma
G
427 ± 3 Ma
H
423 ± 4 Ma
I
413 ± 3 Ma
Age (Ma)
500
450
400
350
39Ar (%)
0
100

area of strong dynamic recrystallization during D_4 deformation. An alternative explanation is that this sample is located at the site of highest metamorphic grade, and its younger age may reflect the longer time to cool from peak metamorphic temperature to the muscovite blocking temperature.

The Corner Brook Lake region lies along strike from, and contains similar rock units to, the Baie Verte Peninsula (Fig. 2). Recent U-Pb data from this latter region give a similar age range to the Corner Brook area for peak metamorphism and deformation (Cawood and Dunning, 1993). Monazite from a syntectonic leucogranite melt within a psammite from the Fleur de Lys Supergroup yielded a precise concordant U-Pb age of 427 ± 2 Ma. Monazite from the muscovite-garnet bearing syntectonic S-type phase of the Wild Cove Igneous Suite gave an identical age (427 ± 2 Ma). Zircon and titanite from a K-feldspar megacrystic posttectonic I-type pluton of the suite, indicate an age of crystallization of 423 ± 3 Ma. Ar-Ar data from the Fleur de Lys Supergroup give ages as old as 429 ± 10 Ma for hornblende and 421 ± 10 Ma for muscovite (Dallmeyer, 1977; recalculated to new constants).

U-Pb and Ar-Ar age data covering an along strike distance of some 150 km between Corner Brook Lake and Baie Verte, constrain the time of peak amphibolite facies metamorphism and associated regional deformation to a relatively narrow time frame of around 435-425 Ma.

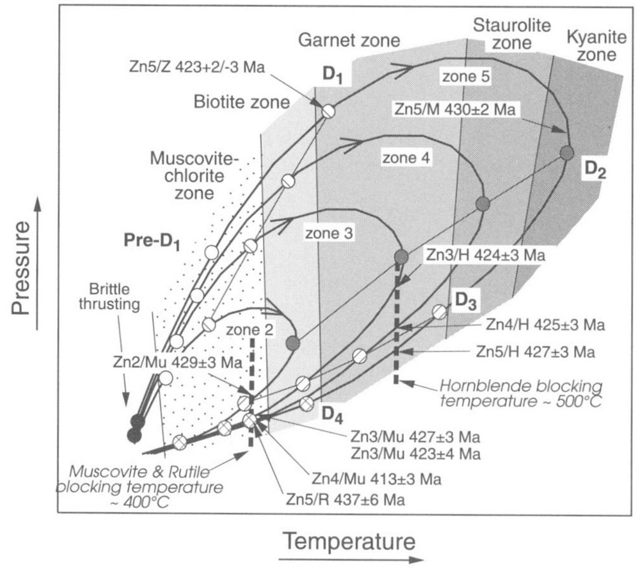


Figure 59. *Schematic P-T-t loops combining the geochronological data with the relative timing of deformation and metamorphism. Zn2 to Zn5 indicate the metamorphic zones from which the samples were collected. H - hornblende; M - monazite; Mu - muscovite; R -rutile; Z - zircon.*

Relationship to P-T loop

Figure 59 combines data on the relative and absolute timing of deformational events with metamorphic P-T paths. The integration of this variety of databases emphasizes the rapidity of the Silurian orogenic pulse. In metamorphic zone 5, the D_1 event had commenced by 434 +2/-3 Ma, on the basis of the dated pegmatite which is deformed by D_1. Metamorphic peak conditions, which reached ~9 kbar and ~650°C in zone 5 and which predated D_2, were reached by 430 ± 2 Ma, the age of monazite from the garnet-kyanite-staurolite schist. Hornblende from a basement amphibolite within the zone gave an age of 427 ± 3 Ma, indicating cooling of the zone through the hornblende blocking temperature of ~500°C by this time. Rutile from the garnet-kyanite-staurolite schist gave an age of 437 ± 6 Ma, which is interpreted as a cooling age through its metamorphic blocking temperature of ~400°C (Mezger et al., 1989). Integration of the pegmatite age with the metamorphic ages from the garnet-kyanite-staurolite schist and the basement amphibolite, indicate that the duration from D_1 to peak metamorphism is no more than a few million years, and may, within error, be essentially synchronous. Similarly, the time range within zone 5 from peak metamorphism down to the hornblende and rutile blocking temperatures and encompassing approximately the duration of D_2 to D_4 events is about 1 or 2 Ma. The database from the other metamorphic zones is limited to Ar-Ar data and only constrains the postmetamorphic peak decompression history of the P-T loop, but data are consistent with, and similar to, those from zone 5. A hornblende cooling age of 424 ± 3 Ma from zone 3, probably closely followed peak metamorphic temperatures within the zone which reached a maximum of less than ~580°C. Muscovite from this zone gave ages of 427 ± 3 Ma and 423 ± 4 Ma and overlaps, within error, the hornblende data, which requires cooling between the respective blocking temperatures of these minerals, along with associated decompression, to have taken no more than a few million years. Data for zones 2 and 4 are similar, with the exception of the relatively young muscovite age of 413 ± 3 Ma from zone 4, the reason for which is uncertain, but as noted earlier could be related to a high strain D_4 overprint, implying a relatively young age for this event.

The rapid cooling of the thrust pile following the metamorphic peak requires that uplift and exhumation were not driven by simple static cooling associated with erosion (cf. England and Richardson, 1977) but must have been tectonically driven. Ductile thrusting during deformational event D_2, and folding during D_3, are responsible for post-peak metamorphic uplift, and possibly caused tectonic or erosional unroofing which resulted in the rapid exhumation of the rocks.

←

Figure 58. *Geochronological data and sample locations in the area around Corner Brook Lake.* ***A)*** *Map showing the locations and ages of samples for U-Pb and Ar-Ar analyses. H - hornblende; M -monazite; Mu - muscovite; R - rutile; Z - zircon.* ***B)*** *U-Pb concordia plot for rutile and monazite from the garnet-kyanite-staurolite schist. R1, R2, M1, and M2 refer to rutile and monazite analyses, respectively, in Table 1.* ***C) to I)*** *Ar-Ar release spectra for hornblende from amphibolites (C-E) and muscovite from psammitic and pelitic schists (F-I).*

Dunnage Zone

Data from the Glover Island Belt provide some important new constrains on timing of regional deformation within the western Dunnage Zone. The main D_1 deformation postdates the Glover Island Granodiorite, the youngest deformed unit east of the Cabot Fault system, but regional penetrative deformation must predate deposition of the Carboniferous strata. The U-Pb age of 440 ± 2 Ma for the granodiorite suggests deformation must be Silurian-Devonian. There are insufficient data to determine if this deformation corresponds with the Silurian Salinian orogeny or with the Devonian Acadian orogeny, but the similarity of the main deformational event with deformation in the Corner Brook Lake Belt suggests that both may be of similar age, and hence Silurian.

The continuity of the tectonic fabric across the Kettle Pond shear zone and the shear zone separating the Corner Brook Lake and Glover Island belts on the west side of the island, suggests that formation of these structures, as well as the juxtaposition of main lithological assemblages within the area, including the Humber/Dunnage boundary, are also of mid-Paleozoic age. This contrasts with previous studies which have assumed that assembly of the lithotectonic units was related to the Ordovician Taconian orogeny (e.g. Williams and St-Julien, 1982; Knapp, 1982). However the intra-oceanic setting of the pre-Middle Ordovician rock units of the Glover Island Belt contrasts with the likely continental setting for emplacement of the Latest Ordovician to earliest Silurian Glover Island Granodiorite and related plutons, requiring obduction of this sequence, presumably during this time frame. The deformed xenoliths of Glover Formation within the Glover Island Granodiorite could be an expression of this. The apparent absence of regional deformational fabric within the Glover Formation predating granodiorite emplacement, may indicate that the main S_1 fabric, mapped within the formation and the granodiorite, was co-axial with an earlier fabric of Taconian age.

ECONOMIC GEOLOGY

Exploration activity in the Corner Brook region has focused on the potential for zinc mineralization in the carbonate rocks of the Humber Zone, and gold mineralization in igneous rocks of the Dunnage Zone. In addition, significant potential for industrial minerals exists in the internal domain of the Humber Zone with extensive outcrops of marble, garnet-kyanite schist, and quartzite.

Humber Zone

Zinc

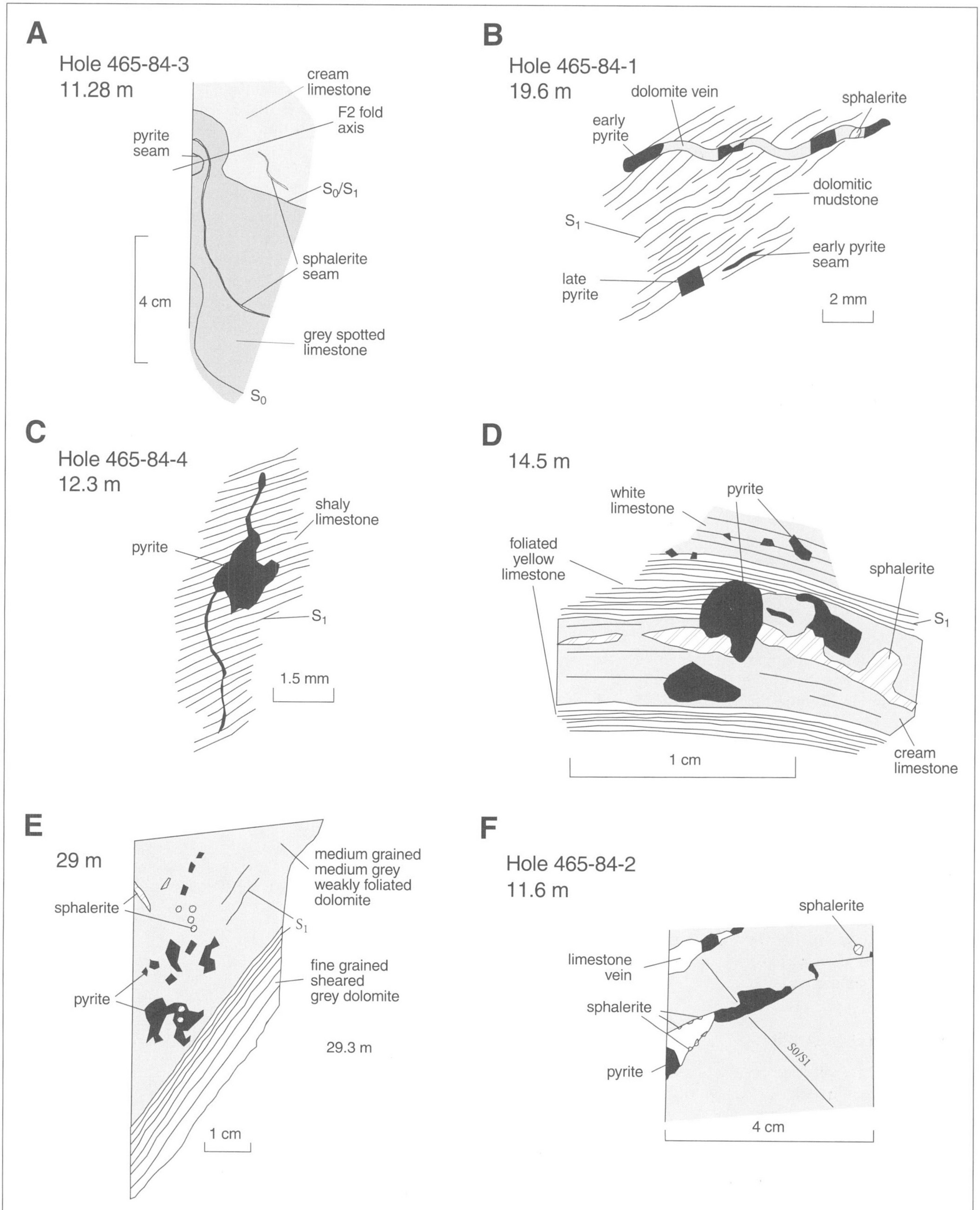
Zinc mineralization is relatively widespread within the carbonate rocks of the Appalachian Humber Zone, and significant sphalerite deposits were mined at Daniels Harbour on the Great Northern Peninsula. In the Corner Brook region, just north of Humber Gorge, a minor zinc showing consists of disseminated grains and stringers of sphalerite in association with pyrite±galena. It is located at the southern end of the Corner Brook Pulp and Paper Fee Simple Mining Grant (g.r. 373232), approximately 5 km due east of Corner Brook. It is herein termed the Zinc Pond showing, after the informal name given to the small pond adjacent to the site of mineralization.

Exploration at Zinc Pond was initiated by Westfield Minerals, following a regional lake sediment survey carried out by the Newfoundland Department of Mines and Energy in 1980. Additional lake sediment geochemistry, undertaken by Westfield Minerals, as well as stream sediment and soil geochemistry substantiated the anomalous zinc concentrations at Zinc Pond and highlighted a number of other potential exploration targets within the Fee Simple Mining Grant. Geological mapping and trenching have shown that mineralization at Zinc Pond occurs in a sequence of dolostone, limestone, and thin-bedded slaty carbonate. Five diamond-drill holes within the Zinc Pond showing showed subeconomic zinc mineralization.

Geological mapping at Zinc Pond enabled three lithological groupings, to be recognized all within the host Reluctant Head Formation: thin-bedded limestone and slate, massive dolostone, and thick-bedded to massive limestone and limestone conglomerate. All units have undergone multiple penetrative deformation and lower greenschist facies metamorphism. The limestone and dolostone sequences lie along a prominent ridge east of Zinc Pond. Mapping of trenches and re-evaluation of drill core from the area show that mineralization is concentrated within the massive, competent dolostone and limestone sequences at their western contact with the less competent slate and thin-bedded limestone. Sphalerite mineralization is associated with pyrite and locally galena, and occurs as disseminated grains, grain aggregates, and stringers.

Detailed relations on the timing of mineralization with respect to deformation, suggest mineralization both pre- and post-dated penetrative deformation, but with the bulk of the mineralization postdating deformation. Timing of mineralization is best determined in finer grained siliciclastic lithologies. Mineral growth prior to foliation development is shown by: flattening of pyrite seams parallel to the S_1 foliation; flattening of pyrite and sphalerite seams in the S_1 foliation with additional folding about F_2 fold axes (Fig. 60A); and folding of dolomite veins containing sphalerite and pyrite by the foliation (Fig. 60B). Mineral growth postdating foliation development

Figure 60. *Textural relations indicative of pre-foliation mineralization (A, B) and post-foliation mineralization (C-F) development in diamond-drill holes from Zinc Pond. Down core in each sketch is towards top of page.* ***A)*** *Pyrite and sphalerite seams flattened by bedding parallel S_1 cleavage, folded about F_2 fold axis at ~11.28 m in hole 465-84-3.* ***B)*** *Dolomite vein with pyrite and sphalerite cutting dolomitic mudstone deformed by foliation at ~19.6 m in hole 465-84-1.* ***C)*** *Pyrite aggregate with delicate extensions overgrowing S_1 foliation at ~12.3 m in hole 465-84-4.* ***D)*** *Sphalerite and pyrite aggregates overgrowing foliation in sequence of thin-bedded limestone at ~14.5 m in hole 465-84-4.* ***E)*** *Coarse pyrite and sphalerite grain aggregates postdating foliation development in foliated dolostone at ~29 m in hole 465-84-4.* ***F)*** *Sphalerite-pyrite mineral aggregates infilling veins in hole 465-84-2.*


A
Hole 465-84-3
11.28 m
cream limestone
F2 fold axis
pyrite seam
S_0/S_1
sphalerite seam
4 cm
grey spotted limestone
S_0
B
Hole 465-84-1
19.6 m
dolomite vein
sphalerite
early pyrite
dolomitic mudstone
S_1
early pyrite seam
late pyrite
2 mm
C
Hole 465-84-4
12.3 m
shaly limestone
pyrite
S_1
1.5 mm
D
14.5 m
white limestone
pyrite
foliated yellow limestone
sphalerite
S_1
1 cm
cream limestone
E
29 m
medium grained medium grey weakly foliated dolomite
sphalerite
S_1
fine grained sheared grey dolomite
pyrite
29.3 m
1 cm
F
Hole 465-84-2
11.6 m
sphalerite
limestone vein
sphalerite
S0/S1
pyrite
4 cm

is shown by: disseminated, equidimensional pyrite grains and aggregates, the latter often with fine delicate extensions, overgrowing and truncating the foliation (Fig. 60C); pyrite-sphalerite aggregates overgrowing the foliation (Fig. 60D). Some samples provide evidence for both pre- and post-foliation mineralization. These relations suggest early mineralization followed by either a second mineralizing event or remobilization of early formed mineral phases, the latter generating most of the mineralization.

In the competent carbonate-dominated lithologies, which form the main locus of sphalerite precipitation, establishing timing of mineralization with respect to foliation development is hindered by either the absence or poorly developed character of the foliation. The presence of undeformed coarse aggregates and veins of mineralization in samples with a weak foliation (Fig. 60E, F) suggests, however, that mineralization took place in late open fractures or voids, and was not subjected to later flattening. In addition, it may indicate that similar aggregates and veins in samples in which no foliation was observed, formed at a similar time.

Relationships within the drill core indicate multiple veining events, with both deformed and undeformed mineralized veins. A late veining event postdating mineralization is also recognized, with barren cream dolomite veins cutting an inferred late pyrite stringer. Further evidence for a late phase of barren dolomitization is shown by brecciation of coarse pyrite aggregates by later dolomite infiltration.

Evidence that deformation within the eastern Humber Zone is Silurian, combined with relations on the relative timing of mineralization, suggest that the mineralization is Silurian or younger. An upper age limit on the mineralization is potentially provided by its relationship with the widespread hematite staining event in the rocks, which postdates mineralization. In Humber Gorge and near Wild Cove Lake, hematite staining is associated with faulting and fracturing of inferred Carboniferous age. If it is of similar age at Zinc Pond, then this relationship suggests the mineralization is pre-Carboniferous.

Copper

Malachite occurs in trace amounts in carbonate of the Breeches Pond Formation at the thrust contact (pre-D_1 structure) with amphibolite of the Corner Brook Lake Complex (g.r. 365018). Mineralization is probably related to fluid movement along the contact.

Garnet-kyanite-staurolite

Garnet-kyanite-staurolite pelitic schist is well developed within the South Brook Formation in the vicinity of Corner Brook Lake, and has potentially commercial value in the abrasive and refractory industries. Garnet is used in the abrasives industry, kyanite after thermal conversion to mullite is an important refractory commodity, and staurolite can be used as a sandblasting agent. House (1993) carried out a preliminary investigation of the garnet-kyanite-staurolite schist discovered by Cawood and van Gool (1992).

Figure 10 shows the distribution of the garnet-kyanite-staurolite pelitic schist around Corner Brook Lake. The schist is up to tens of metres thick, can be traced along strike for several kilometres, and is repeated across strike by thrust faulting. Garnets are up to 5 cm diameter, but with most in the 1.5 to 3 cm range; kyanite crystals are up to 15 cm long, but are generally 5 cm or less in length; and staurolite, which can reach up to 7 cm long, is generally less than 3 cm in length. The proportion of garnet within the pelitic schist ranges from 1 to 25%, kyanite from 0 to 25%, and staurolite from 0 to 10%. Other components of the schist are muscovite, plagioclase, quartz, and rare biotite and graphite, along with the accessory phases Fe (Ti) oxides, epidote, chlorite, rutile, and tourmaline. Garnet, kyanite, and staurolite occur as porphyroblasts which overgrow the main foliation and contain numerous inclusions, notably quartz, graphite, and muscovite.

Silica

Quartzite is a minor but widespread phase within the South Brook Formation, and a rare phase within the Corner Brook Lake Complex. The best occurrences are in the region northeast of Marble Mountain, along the ridge south of the Trans Canada Highway. Minor occurrences are also found in association with the garnet-kyanite schist at Corner Brook Lake. The quartzite occurs in bands up to tens of metres thick, which can be traced for hundreds of metres along strike. Impurities are widespread, and consist of muscovite, feldspar, and magnetite.

Marble

Carbonate rocks are well developed in the northwest third of the Corner Brook map area and have formed an important commercial nonmetallic mineral deposit within the area. Limestone quarries in the St. George and Table Head groups immediately east of Corner Brook, as well a recently opened quarry in carbonate of the Pinchgut Formation to the south of town, have been used in cement and as a road aggregate. In addition, marble within the upper St. George Group and exposed along the Lady Slipper road system is currently being appraised for use in the dimension stone industry.

Dunnage Zone

Mafic and felsic igneous rocks of the Dunnage Zone host significant base metal deposits in the Appalachian Orogen (Swinden, 1991). Recent exploration with the Dunnage Zone has shown that it also forms an important potential source of precious metal deposits. The base metal potential of the map area is limited. Minor pyrite concentrations, sometimes with copper mineralization, were locally noted in mafic rocks of the Glover Formation, but all occurrences are spatially restricted and isolated. A major gold showing is currently under assessment within, and immediately south of, the map area, within the Kettle Pond shear zone.

Gold

The position of Glover Island along strike from the Baie Verte Peninsula, which was a locus of gold exploration in the 1980s (e.g. Tuach et al., 1988), led to staking of the island by Varna Gold Inc. in 1985. Subsequent exploration demonstrated widespread gold mineralization hosted by the Kettle Pond shear zone. Exploration within the area is currently being undertaken by Newfoundland Goldbar Resources Inc., and Barbour and French (1993) noted that thirteen important prospects have been identified along an 8 km strike length of the shear zone, between 'Tomahawk pond' (g.r. 462987) and 'Lunch pond' (g.r. 420946), the latter lying just south of the map sheet.

Mineralization is hosted by the Kettle Pond shear zone, consisting of mylonitic mafic to felsic igneous rocks and volcaniclastic sedimentary rocks. Barbour and French (1993) divided the mineralization into three types, based on the morphology of the host rock: quartz vein type, consisting of deformed auriferous bull quartz veins with individual veins up to 2 m thick, occurring in zones up to 25 m wide and 50 m long; felsite type, consisting of aphanitic siliceous felsite and/or fine-grained, iron carbonate-altered rock, cut by quartz vein stockwork; and silicification and disseminated pyrite, showing pervasive silicification of chlorite schist and ultramylonite, with pyrite stringers across zones 40-50 m in width.

Barbour and French (1993) reported gold values from grab samples at the main showings along the 7 km long mineralized zone, ranging from 0.5 to 14 g/t Au, with channel samples at the Discovery Vein at Kettle Pond giving values of up to 35g/t over 0.78 m. A number of Cu-Ni sulphide showings have also been recognized along the zone.

The concentration of mineralization within the Kettle Pond shear zone requires mineralization to be syn- to post-formation of the zone. Barbour and French (1993) suggested that some of the deformed felsite dykes within the shear zone may be derived from the Glover Island Granodiorite, indicating that both deformation and mineralization postdate the 440 Ma age of the granodiorite.

CORRELATION AND TECTONIC SIGNIFICANCE OF ROCK UNITS

Correlation

Lithological and isotopic age data for the rock units of the Corner Brook region allow correlation with other rock units in the Newfoundland Appalachians.

The gneissic character of the Corner Brook Lake Complex, a U-Pb age of around 1510 Ma for at least part of the complex, and its stratigraphic contact with the overlying sedimentary cover lithologies of the Humber Zone, suggest that it constitutes part of the Grenville basement to the Appalachian Orogen. The complex represents an along strike extension of the Disappointment Hill complex and an unnamed gneiss complex mapped around, and south of, the southern shore of Grand Lake by Currie (1987), Piasecki (1991), and Currie and van Berkel (1992a). Currie et al. (1992) determined a U-Pb zircon age of 1498+9/-8 Ma for a felsic granulitic gneiss of the Disappointment Hill complex. North of the Corner Brook Lake Complex, the quartzofeldspathic gneiss of the Long Range complex exposed within the Long Range Inlier (Fig. 2) has given a minimum U-Pb zircon age of 1550 Ma (H. Baadsgaard, pers. comm., in Owen, 1991). Undated Grenvillian inliers within the Humber Zone also occur at Indian Head (Williams and Cawood, 1989) and possibly on the Baie Verte Peninsula (de Witt, 1972, 1974, 1980; Hibbard, 1983a).

Tucker and Gower (1994) reported U-Pb ages of 1490 ± 5 Ma, 1479 ± 2 Ma, and 1472 ± 2 Ma for granitoids from the Pinware terrane in the Grenville Orogen of southeast Labrador. The Pinware terrane lies within the interior magmatic belt of the Grenville Orogen. It is situated immediately northwest of the Great Northern Peninsula of Newfoundland, and abuts the Appalachian Orogen at the Appalachian structural front. Tucker and Gower (1994) reviewed geochronological evidence from throughout eastern Laurentia and southwest Baltica, and suggested that the ages from the Pinware terrane are part of a regionally extensive pulse of granitoid magmatism, to which they proposed the name Pinwarian. The similarity of ages for the dated Grenvillian inliers in the Humber Zone to those recorded from the Pinware terrane, suggests all may be part of a single Grenvillian terrane extending from Labrador under the northwestern Appalachians.

The Late Precambrian Lady Slipper Pluton forms the oldest element of the Humber Zone exposed within the Corner Brook Lake region. Similar units include the Round Head Complex at the base of the Mount Musgrave Group in the Pasadina map sheet, the Hare Hill pluton intrusive into Grenville basement south of Grand Lake, and the Long Range dyke swarm which intrudes Grenville basement and feeds mafic flows on basement around the Long Range Inlier on the Great Northern Peninsula and in southeast Labrador. This igneous activity and associated siliciclastic sedimentation (e.g. Mount Musgrave Group) represent the earliest phase of the Appalachian orogenic cycle within the Humber Zone, and are interpreted to be related to rifting and stretching of continental lithosphere of the Grenville orogen associated with establishment of the eastern margin of Laurentia (e.g. Williams and Hiscott, 1987). The U-Pb zircon age determined for the Lady Slipper pluton of 555+3/-2 Ma is significantly younger than the 620-600 Ma age range determined for the other rift-related igneous events in western Newfoundland and southeast Labrador (Stukas and Reynolds, 1974; Currie et al., 1992; Williams et al., 1985; Kamo et al., 1989). The age of the pluton is similar to the 554+4/-2 Ma age determined for the Tibbit Hill volcanic unit in Quebec (Kumarapeli et al., 1989). Rift-related igneous activity as young as 555 to 550 Ma raises the possibility that the timing of commencement of sedimentation over the igneous rocks elsewhere, may be considerably younger than previously thought. Knight and Cawood (1991) have previously pointed out that the volcanic flows of the Lighthouse Cove Formation on the Great Northern Peninsula have a ferrugenous cap with breccias and spheroids, that is locally reworked into a ferrugenous granular sandstone, suggesting that the flows were subaerially weathered during a significant hiatus before deposition of the Bradore Formation of the Labrador Group.

Overlying basement of the Corner Brook Lake Complex and the Lady Slipper Pluton, is the sedimentary cover sequence of the Humber Zone. This is divisible into a lower siliciclastic-dominated sequence, and an upper carbonate-dominated sequence. In the map area, the basement-cover contact is preserved only within the internal domain of the zone, represented by the Corner Brook Lake Belt. In this belt the siliciclastic sequence is represented by the South Brook Formation, and the carbonate sequence by the Breeches Pond Formation. In the Humber Arm Allochthon, this bipartite division is represented by the Summerside Formation and the Irishtown and Pinchgut formations, respectively. In the Carbonate Belt, only the upper carbonate-dominated part of the cover sequence is preserved, although to the north and south of the map area, the lower siliciclastic sequence and the basement-cover contact are recognized (e.g. Williams and Cawood, 1989).

The lithological character and ordering of rock units within the Grand Lake Complex suggest that it represents a disrupted ophiolite. It consists of basal greenschist, interpreted by Knapp (1982) to represent a retrogressed metamorphic sole, overlain by ultramafic cumulates of websterite and clinopyroxenite, passing up into gabbro with minor trondhjemite at the top. The ultramafic cumulate and gabbro section is similar to the transition zone preserved in other more complete ophiolite sequences, both in the Appalachian Orogen and other orogenic belts worldwide (e.g. Coleman, 1977). The upper segments of a normal ophiolite sequence, including the sheeted dyke complex and overlying pillow basalts, are absent, presumably due to removal during subsequent faulting at the top of the complex along the Kettle Pond shear zone. The 490 Ma age of the Grand Lake Complex is similar to other ophiolites within the Appalachian-Caledonian Orogen (Dunning and Krogh, 1985; Dunning and Pederson, 1988). Ophiolites within the orogen occupy a limited time range from 495-480 Ma, in contrast to island arc-type igneous activity, which may range from around 515 Ma to 460 Ma (Dunning et al., 1987, 1991).

The range in composition of the igneous rocks of the Glover Formation, from mafic to felsic, the low-K tholeiite composition of these rocks (Knapp, 1982), and the evidence for at least some explosive igneous activity indicated by the presence of tuffs and associated volcaniclastic sedimentary rocks, suggest that the formation accumulated in an island arc setting. Although Knapp (1982) thought that the formation accumulated on the ophiolitic basement of the Grand Lake Complex, the recognition of a fault at the basal contact of the formation (see map) means its original paleogeographic relationship with the complex is unknown. However, the low-K character of the rock units is consistent with an intra-oceanic setting.

Williams and St-Julien (1982) and Knapp (1982) have compared the Dunnage Zone rock units on Glover Island (Grand Lake Complex and Glover Formation) with similar sequences on Baie Verte Peninsula. In particular, the Grand Lake Complex was correlated with the Advocate Complex, on the basis of its deformed and disrupted ophiolitic character, and the inferred similar structural position immediately east of the Baie Verte Line. The interpreted unconformable contact between the complex and the overlying Glover Formation (group of Knapp, 1982), and the recognition of conglomerate along the contact, led to the correlation of this unit (Knapp's Kettle Pond formation) with conglomerate of the Flatwater Pond Group on Baie Verte Peninsula. The recognition of a structural, rather than a stratigraphic contact between these units, largely invalidates the basis for these correlations.

The Glover Island Granodiorite has been included within the Hungry Mountain Complex, which outcrops largely to the east and north of Grand Lake (Whalen, 1989). However, a U-Pb zircon date of 440 Ma for the granodiorite indicates that it is significantly younger than the early to middle Ordovician age of the complex (e.g. Dunning et al., 1987; Whalen et al., 1987). The age of the Glover Island Granodiorite is similar to the eastern portion of the Burlington Granodiorite. Dunning and Cawood (1993) determined U-Pb zircon ages of 440 ± 2 Ma for the eastern portion of the Burlington body and 432 ± 2 Ma for its western portion.

Tectonic setting

The Humber Zone is considered to represent part of the eastern continental margin of Laurentia (e.g. Williams and Stevens, 1974; Williams, 1979). Figure 61 is a schematic palinspastic reconstruction of the Laurentian margin, incorporating results from the Corner Brook Lake-Humber Arm region. The transition within the Humber Zone from a lower siliciclastic-dominated to an upper carbonate-dominated sequence is interpreted to correspond with the rift-drift transition of modern continental margins. As pointed out by Williams and Hiscott (1987), the lower siliciclastic sequence accumulated in both nonmarine and marine settings, and shows both rapid along- and across-strike variations in facies type and thickness of stratigraphic units. This, combined with the unconformable relationship between basement and cover and the spatial and temporal association with igneous activity (Lady Slipper Pluton and amphibolite dykes), suggests an unstable tectonic setting associated with rifting of continental lithosphere and upwelling of underlying asthenosphere. In contrast, the carbonate-dominated sequence is characterized by stratigraphic continuity of facies types and uniform lateral thickness of stratigraphic units, open marine sedimentation with an overall deepening to the east, and an absence of any associated igneous activity. This suggested to Williams and Hiscott (1987) accumulation is a stable tectonic environment associated with passive thermal subsidence of an open continental margin.

The analysis of the spatial and temporal variation in the character of the Humber Zone sequence was based largely on the excellent exposures preserved within the external domain of the zone on the Great Northern Peninsula and the Port au Port Peninsula. The overall metamorphic and deformational state of the rock units within the Corner Brook map area prevents a detailed analysis of facies thickness variations within and between rock units. However, broad relationships within the area are consistent with the change from siliciclastic- to carbonate-dominated sedimentation, corresponding with the rift-drift transition of continental margin development. In particular, the siliciclastic sequence within the Corner Brook Lake Belt shows rapid across strike variations in thickness typical of rift-facies rocks. Igneous activity in the form of amphibolite dykes and the Lady Slipper Pluton is associated with this sequence, but absent from the carbonate sequence.

Evidence of facies and thickness variations within the South Brook Formation is displayed around the basement inliers by Corner Brook Lake and by the southwestern arm of Grand Lake (see map; Fig. 61). These regions establish a direct correspondence between the distribution of basement outcrops and the thickness of the siliciclastic sequence. Around the large basement inlier at the west end of the southwestern arm of Grand Lake (Fig. 62), the thickness of the South Brook Formation varies from less than 15 m on its western margin up to a few hundred metres on the eastern side of the block. Corresponding with this change in thickness, the facies of the South Brook Formation (g.r. 218944) also varies dramatically from west to east. In the west the formation consists of 5 m of quartz-rich arkose directly overlying basement, passing up into 5 m of micaceous marble which is in turn overlain by 4-5 m of magnetite-rich arkose, overlain by calc-schist and pelite of the Breeches Pond Formation. Outcrop on the eastern margin of the block is poor, but consists of a lower 10 m sequence of quartz pebble conglomerate passing up into a thick, poorly exposed sequence of arkosic psammite. The change in facies types and unit thickness approximately corresponds with the northern end of the basement block and, based on available outcrop data, appears to be abrupt. We consider this abrupt change to reflect the presence of an original horst and graben structure within the block, with the eastern margin of the horst block approximately corresponding with the hinge of the antiform (Fig. 62).

A further example of a relationship between basement outcrops and thickness of the siliciclastic sequence occurs at Corner Brook Lake. Here, the thrust sheet containing the main basement block has only a thin veneer of siliciclastic strata of the South Brook Formation (5-10 m) before passing into carbonate lithologies of the Breeches Pond Formation (Fig. 10). In overlying thrust sheets, basement occurs in thin and commonly discontinuous slices at the base of a sheet, and passes up into a South Brook Formation sequence in the order of tens of metres to a maximum of 100 to 200 m thick (Fig. 10). East of this basement-cored thrust stack, is a large overthrust block of South Brook Formation, which extends east to Grand Lake, and which lacks any associated basement. Although the thickness of the formation within this block has not been accurately established, it is at least many hundreds of metres, and is probably many thousands of metres. These relationships are consistent with a model in which areas of basement involvement correspond with horsts during sedimentation, allowing accumulation of only a thin sequence of rift-related siliciclastic sediments, whereas thrust sheets where basement is absent preserve depositional grabens where thick rift-facies sediments accumulated. In such a

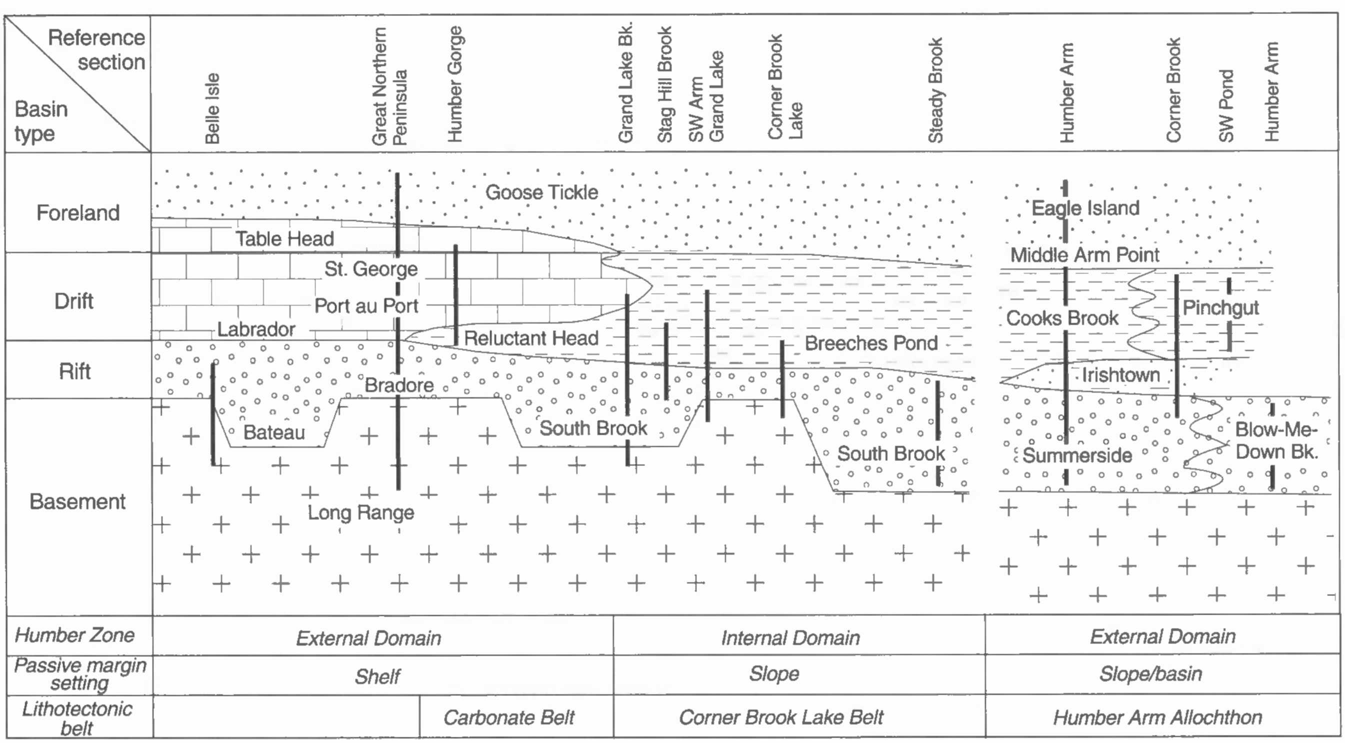


__Figure 61.__ Paleogeographic setting and correlation of principal rock units in the Humber Zone. Vertical lines indicate the location of reference sections within the Corner Brook-Glover Island region and comparison with reference sections from Great Northern Peninsula and Humber Arm.

model, the main thrust fault between the large overthrust block of South Brook Formation and the basement cored thrust stack at Corner Brook Lake may correspond with a reactivated normal fault, which originally separated an area of horsts to the west from a large graben structure to the east.

The presence of basement incorporation in the thrust stack at Corner Brook Lake and its absence from the area immediately to the west by Stag Hill Brook, as well as to the east, suggest that rifting of the continental margin did not involve a simple downstepping of the margin in a series of eastward-facing half-graben. Rather, the Humber Zone margin, like many modern margins (Falvey and Mutter, 1981), included outlying horsts or marginal plateaus which were subsequently plucked and incorporated into the thrust stack during subsequent collisional orogenesis.

Reassessment of the Baie Verte Line and Taconian Orogeny

The Baie Verte Line is classically interpreted as the root zone where the continent-arc collision occurred, and over which allochthons were obducted during the Ordovician Taconian Orogeny. Although the U-Pb and Ar-Ar data we have presented show no evidence of an earlier Taconian orogeny in the Corner Brook-Glover Island region, the possibility that such an event did affect the area cannot be ruled out. The evidence for westward emplacement of the west Newfoundland allochthons during the Taconian (Williams and Stevens, 1974) requires that Dunnage Zone rock units must have overridden the eastern Humber Zone. In southwest Newfoundland, foliated intrusive tonalitic orthogneiss and its host assemblage of psammite equivalent to the Fleur de Lys Supergroup and ophiolitic assemblages, are cut by undeformed high level gabbro of late Ordovician and early Silurian age (Dunning et al., 1987, 1990; Dunning, unpublished data). These relations require a penetrative deformational event of Taconian age. However, our data combined with that of Cawood and Dunning (1993) suggest that the Taconian orogeny, if present in the eastern Humber Zone, did not involve large-scale subduction, deformation, and metamorphism of the margin, as traditionally envisaged. Studies of metamorphism by Jamieson (1990) show that the preservation of low-temperature eclogite on the Baie Verte Peninsula requires metamorphism of cold continental crust, in agreement with our isotopic data, requiring no significant pre-Silurian thermal event. The character of any Taconian orogenic event in the Corner Brook-Baie Verte region could have been similar to that described from west Newfoundland, where deformation is restricted to the thin superficial thrust sheets of the allochthon, and is characterized by formation of scaly shale mélanges (e.g. Cawood and Williams, 1988). The mélanges recorded by Williams (1977) in the Birchy Complex at Coachmans Cove on the Baie Verte Peninsula could be an

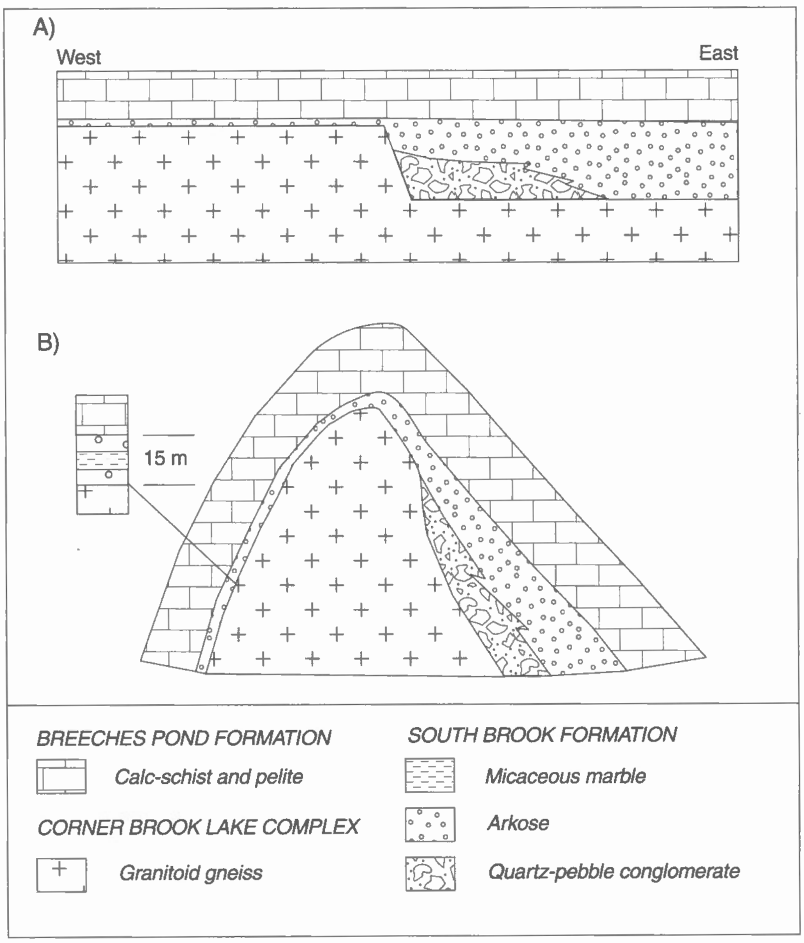


Figure 62.

*Schematic diagram showing: **A**) Reconstruction of inferred depositional setting for South Brook Formation across the basement block in the southwest corner of the Corner Brook-Glover Island region; and **B**) Current structural profile across the block (based on cross-section G).*

expression of this event. The juxtaposition of ophiolitic serpentinite sheets with continental margin sedimentary rocks such as at Matthews Brook (see map) and at Fleur de Lys on the Baie Verte peninsula (Hibbard, 1983a, b), provides a similar relationship to the emplacement of Bay of Islands and St. Anthony ophiolites over rift-facies sedimentary rocks in west Newfoundland (Lindholm and Casey, 1989; Williams and Cawood, 1989; Cawood, 1989). Thus, the Matthews Brook Serpentinite could be a relict of high-level Taconian ophiolite emplacement over the eastern Humber Zone, which was subsequently overprinted by pervasive Salinian deformation.

In the Dunnage Zone rocks of the Corner Brook-Glover Island region, the inclusion of deformed xenoliths of the Glover Formation within the Glover Island Granodiorite, indicates an Ordovician or older deformation of this unit, and can broadly be correlated with the Taconian Orogeny. Knapp (1982) considered the basal greenschist of the Grand Lake Complex to be a sliver of a retrogressed ophiolitic metamorphic sole, formed during Ordovician obduction of the complex.

Although some evidence for Taconian features may be preserved within the Corner Brook-Glover Island region, our data indicate that the principal orogenic event is mid-Silurian. In the eastern Humber Zone, this event involved multiple penetrative deformation and metamorphism up to amphibolite grade. In the Dunnage Zone, Silurian orogenesis is expressed through magmatic activity, greenschist facies metamorphism, and one main penetrative deformation. The contrasting expression of the Silurian orogeny within the rocks of the Humber and Dunnage zones indicates that they were not juxtaposed to their present relative positions until during, or after, Silurian orogenesis. Hence, the Humber-Dunnage contact, as currently exposed, is a Silurian or younger, rather than an Ordovician, feature.

The Dunnage Zone rocks on Glover Island (as well as the Matthews Brook serpentinite) are in thrust contact with Corner Brook Lake Belt rocks, rather than those of the Humber Arm Allochthon which, at the onset of thrusting, are located farther outboard (Fig. 5). This means that the docking of lithotectonic units did not occur in a simple piggy-back fashion. Either out-of-sequence thrusting or extensional faulting must be responsible for the removal of the eastern equivalents of Summerside, Irishtown, and Pinchgut formations, on which the Dunnage Zone rocks were most likely originally emplaced.

REFERENCES

Baird, D.M.
1959: Geology of Sandy Lake (west half), Newfoundland; Geological Survey of Canada, Map 47-1959.

Barbour, D.M. and French, V.A.
1993: Geology and gold mineralization of the Glover Island property, western Newfoundland; in Ore Horizons, Volume 2, A. Hogan and H.S. Swinden (ed.); Newfoundland Department of Mines and Energy, Geological Survey Branch, p. 101-117.

Barss, S.M.
1974: Palynological zonation of the Carboniferous and Permian rocks of the Atlantic Provinces; in Report of Activities, Part B; Geological Survey of Canada, Paper 74-1B, p. 135-136.

Batterson, M.J. and McGrath, B.
1993: Quaternary geology of the Deer Lake and Pasadina map areas (NTS 12 H/3 and 12H/4); in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 93-1, p. 103-112.

Batterson, M.J. and Taylor, D.M.
1990: Glacial history of the Humber River basin; in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 90-1, p. 1-6.

Batterson, M.J. and Vatcher, S.V.
1992: Quaternary geology of the Corner Brook-Pasadina area (NTS 12 A/13 and 12H/4); in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 92-1, p. 1-12.

Berman, R.G.
1988: Internally-consistent thermodynamic data for stoichiometric minerals in the system Na_2O-K_2O-CaO-MgO-FeO-Fe_2O_3-Al_2O_3-SiO_2-TiO_2-H_2O-CO_2; Journal of Petrology, v. 29, p. 445-522.
1991: Thermobarometry using multiequilibrium calculations: a new technique with petrologic applications; Canadian Mineralogist, v. 29, p. 833-855.

Belt, E.S.
1969: Newfoundland Carboniferous stratigraphy and its relation to the Maritimes and Ireland; American Association of Petroleum Geologists, Memoir 12, p. 734-753.

Billings, E.
1865: Palaeozoic fossils. Volume 1; Geological Survey of Canada, 426 p.

Botsford, J.W.
1988: Depositional history of Middle Cambrian to Lower Ordovician deep water sediments, Bay of Islands, western Newfoundland; Ph.D. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 534 p.

Brookes, I.A.
1970: New evidence for an independent Wisconsin-age ice cap over Newfoundland; Canadian Journal of Earth Sciences, v. 7, p. 1374-1382.
1973: Late Wisconsin glaciation of southwestern Newfoundland (with special reference to the Stephenville map-area); Geological Survey of Canada, Paper 73-40, 31 p.

Brückner, W.D.
1966: Stratigraphy and structure of west-central Newfoundland; in Geology of Part of Atlantic Provinces, W.H. Poole (ed.); Geological Association of Canada/Mineralogical Association of Canada, Guidebook, p. 137-151.

Cawood, P.A.
1993: Acadian orogeny in west Newfoundland: definition, character and significance; in The Acadian Orogeny: Recent Studies in New England, Maritime Canada, and the Autochthonous Foreland, D.C. Roy and J.W. Skehan (ed); Geological Society of America, Special Paper 275, p. 135-153.

Cawood, P.A. and Dunning, G.R.
1993: Silurian age for movement on the Baie Verte Line: implications for accretionary tectonics in the Northern Appalachians; Geological Society of America, Abstracts with Programs, v. 25, p. A422.

Cawood, P.A. and van Gool, J.A.M.
1992: Stratigraphic, structural, and metamorphic relations along the eastern margin of the Humber Zone, Corner Brook map area, western Newfoundland; in Current Research, Part E; Geological Survey of Canada, Paper 92-1E, p. 239-247.
1993: Stratigraphic and structural relations within the western Dunnage Zone, Glover Island region, western Newfoundland; in Current Research, Part E; Geological Survey of Canada, Paper 93-1D, p. 29-37.

Cawood, P.A. and Williams, H.
1988: Acadian basement thrusting, crustal delamination, and structural styles in and around the Humber Arm Allochthon, western Newfoundland; Geology, v. 16, p. 370-373.

Cawood, P.A., Williams, H., O'Brien, S.J., and O'Neill, P.P.
1988: A geologic cross-section of the Appalachian Orogen; Geological Association of Canada, Mineralogical Association of Canada, and Canadian Society of Petroleum Geologists, Joint Annual Meeting, Field Excursion Guidebook, 160 p.

Coleman, R.G.
1977: Ophiolites – Ancient Oceanic Lithosphere? Minerals and Rocks, v. 12; Springer-Verlag, New York, 229 p.

Colman-Sadd, S.P., Hayes, J.P., and Knight, I.
1990: Geology of the island of Newfoundland; Geological Survey Branch, Department of Mines and Energy, Government of Newfoundland and Labrador, Map 90-01, scale 1:1 000 000.

Cooper, J.R.
1937: Geology and mineral deposits of the Hare Bay area; Newfoundland Department of Natural Resources, Geological Section, Bulletin No. 9, 36 p.

Cumming, L.M.
1983: Lower Paleozoic autochthonous strata of the Strait of Belle Isle area; in Geology of the Strait of Belle Isle area, northwestern insular Newfoundland, southern Labrador and adjacent Quebec; Geological Survey of Canada, Memoir 400, Part 2, p. 75-108.

Currie, K.L.
1987: A preliminary account of the geology of Harrys River map area, southern Long Range of Newfoundland; in Current Research, Part A; Geological Survey of Canada, Paper 87-1A, p. 653-662.

Currie, K.L. and van Berkel, J.T.
1992a: Geology, southern Long Range Mountains, Newfoundland; Geological Survey of Canada, Map 1815A, scale 1:100 000.
1992b: Notes to accompany a geological map of the southern Long Range, southwestern Newfoundland; Geological Survey of Canada, Paper 91-10, 10 p.

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

Dallmeyer, R.D.
1977: $^{40}Ar/^{39}Ar$ age spectra of minerals from the Fleur de Lys terrane in northwest Newfoundland: their bearing on chronology of metamorphism within the Appalachian orthotectonic zone; Journal of Geology, v. 85, p. 89-103.

de Wit, M.J.
1972: The geology around Bear Cove, eastern White Bay, Newfoundland; Ph.D. thesis, Cambridge University, England, 232 p.
1974: On the origin and deformation of the Fleur de Lys metaconglomerate, Appalachian fold belt, northwest Newfoundland; Canadian Journal of Earth Sciences, v. 11, p. 1168-1180.
1980: Structural and metamorphic relationships of pre-Fleur de Lys and Fleur de Lys rocks of the Baie Verte Peninsula, Newfoundland; Canadian Journal of Earth Sciences, v. 17, p. 1559-1575.

Dean, P.L.
1978: The volcanic stratigraphy and metallogeny of Notre Dame Bay, Newfoundland; Memorial University of Newfoundland, Report 7, 205 p.

Dean, P.L. and Meyer, J.R.
1985: West coast shale projects; in Current Research; Newfoundland Department of Mines and Energy, Mineral Development Division, Report 85-1, p. 74-78.

Dunning, G.R. and Krogh, T.E.
1985: Geochronology of ophiolites of the Newfoundland Appalachians; Canadian Journal of Earth Sciences, v. 22, p. 1659-1670.

Dunning, G.R. and Pedersen, R.B.
1988: U/Pb ages of ophiolites and arc-related plutons of the Norwegian Caledonides: implications for the development of Iapetus; Contributions to Mineralogy and Petrology, v. 98, p. 13-23.

Dunning, G.R., Kean, B.F., Thurlow, J.G., and Swinden, H.S.
1987: Geochronology of the Buchans, Roberts Arm and Victoria Lake Groups and the Mansfield Cove Complex, Newfoundland; Canadian Journal of Earth Sciences, v. 24, p. 1175-1184.

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

Dunning, G.R., Swinden, H.S., Kean, B.F., Evans, D.T.W., and Jenner, G.A.
1991: A Cambrian island arc in Iapetus: geochronology and geochemistry of the Lake Ambrose Volcanic Belt, Newfoundland Appalachians; Geological Magazine, v. 128, p. 1-17.

England, P.C. and Richardson, S.W.
1977: The influence of erosion upon the mineral facies of rocks from different metamorphic environments; Geological Society of London, Journal, v. 134, p. 201-213.

Falvey, D.A. and Mutter, J.D.
1981: Regional plate tectonics and the evolution of Australia's passive continental margin; Australian Bureau of Mineral Resources, Journal of Australian Geology and Geophysics, v. 6, p. 1-29.

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

Gillespie, R.T.
1983: Stratigraphic and structural relationships among rock groups at Old Man Pond, western Newfoundland; M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 198 p.

Greenwood, H.J.
1975: The buffering of pore fluids by metamorphic reactions; American Journal of Science, v. 275, p. 573-593.

Hayes, A.O. and Johnson, H.
1938: Geology of the Bay St. George Carboniferous area; Geological Survey of Newfoundland, Bulletin 12, 62 p.

Heyl, G.R.
1937: The geology of the Sops Arm area, White Bay, Newfoundland; Geological Survey of Newfoundland, Bulletin 8, 42 p.

Hibbard, J.P.
1983a: Geology of the Baie Verte Peninsula, Newfoundland; Newfoundland Department of Mines and Energy, Mineral Development Division, Memoir 2, 279 p.
1983b: Notes on the metamorphic rocks in the Corner Brook area (12A/13) and regional correlation of the Fleur de Lys belt, Newfoundland; in Current Research, J. Murray, P. Saunders, D. Boyce, and R. Gibbons (ed.); Newfoundland Department of Mines and Energy, Report 83-1, p. 41-50.

Holdaway, M.J.
1971: Stability of andalusite and the aluminium silicate phase diagram; American Journal of Science, v. 271, p. 97-131.

House, A.F.
1993: Kyanite and related minerals in western Newfoundland; in Current Research; Newfoundland Department of Mines and Energy, Report 93-1, p. 351-356.

Hyde, R.S.
1978: Stratigraphic subdivision and mapping of the Lower Mississippian Anguille Group, Deer Lake-White Bay area, Newfoundland; in Report of Activities for 1977, J. Murray, P. Saunders, D. Boyce, and R. Gibbons (ed.); Newfoundland Department of Mines and Energy, Mineral Development Division, Report 78-1, p. 116-123.
1979a: Geology of Carboniferous strata in portions of the Deer Lake Basin, western Newfoundland; Newfoundland Department of Mines and Energy, Mineral Development Division, Report 79-6, 43 p.
1979b: Geology of Carboniferous strata in portions of the Deer Lake Basin, western Newfoundland; Newfoundland Department of Mines and Energy, Mineral Development Division, Open File Report No. 1012, with five 1:50 000 geological maps.
1982: Geology of the Carboniferous Deer Lake Basin; Mineral Development Division, Newfoundland Department of Mines and Energy, Map 82-7, scale 1:100 000.
1984: Geologic history of the Carboniferous Deer Lake Basin, west-central Newfoundland, Canada; in Atlantic Coast Basins, Paleogeography and Paleotectonics, Sedimentology and Geochemistry, Vol. 3, H.H.J. Geldsetzer, W.W. Nassichuk, E.S. Belt, and R.W. Macqueen, (ed.); Ninth International Congress of Carboniferous Stratigraphy and Geology, Southern Illinois University Press, Carbondale, Illinois, p. 85-104.

Hyde, R.S., Miller, H.G., Hiscott, R.N., and Wright, J.A.
1988: Basin architecture and thermal maturation in the strike-slip Deer Lake Basin, Carboniferous of Newfoundland; Basin Research, v. 1, p, 85-105.

James, N.P. and Stevens, R.K.
1982: Anatomy and evolution of a lower Paleozoic continental margin, western Newfoundland; International Association of Sedimentologists, Field Excursion Guide Book, Excursion 2B.

Jamieson, R.A.
1990: Metamorphism of an Early Paleozoic continental margin, western Baie Verte Peninsula, Newfoundland; Journal of Metamorphic Geology, v. 8, p. 269-288.

Jamieson, R.A. and O'Beirne-Ryan, A.M.
1991: Decompression-induced growth of albite porphyroblasts, Fleur de Lys Supergroup, western Newfoundland; Journal of Metamorphic Geology, v. 9, p. 433-439.

Kamo, S.L., Gower, C.F., and Krogh, T.E.
1989: Birthdate for the Iapetus Ocean? A precise U-Pb zircon and baddeleyite age for the Long Range dikes, southeast Labrador; Geology, v. 17, p. 602-605.

Kennedy, D.P.
1980: Geology of the Corner Brook Lake area, western Newfoundland; in Current Research, Part A; Geological Survey of Canada, Paper 80-1A, p. 235-240.
1981: Geology of the Corner Brook Lake area, western Newfoundland; M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 370 p.

Knapp, D.A.
1982: Ophiolite emplacement along the Baie Verte-Brompton Line at Glover Island, western Newfoundland; Ph.D. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 338 p.

Knapp, D., Kennedy, D.P., and Martineau, Y.
1979: Stratigraphy, structure, and regional correlation of rocks at Grand Lake, western Newfoundland; in Current Research, Part A; Geological Survey of Canada, Paper 79-1A, p. 317-325.

Knight, I.
1992: Geology of Marmorized, lower Paleozoic, platformal carbonate rocks, 'Pye's Ridge', Deer Lake; in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 92-1, p. 141-157.
1994: Geology of Cambrian-Ordovician platformal rocks of the Pasadena map area (12H/4); in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 94-1, p. 175-186.

Knight, I. and Boyce, W.D.
1991: Deformed lower Paleozoic platform carbonates, Goose Arm – Old Man's Pond; in Current Research; Newfoundland Department of Mines and Energy, Geological Survey Branch, Report 91-1, p. 141-153.

Knight, I. and Cawood, P.A.
1991: Paleozoic geology of western Newfoundland: an exploration of a deformed Cambro-Ordovician passive margin and foreland basin, and Carboniferous successor basin; Centre for Earth Resources Research, Memorial University of Newfoundland short course notes, 403 p.

Knight, I. and James, N.P.
1987: The stratigraphy of the Lower Ordovician St. George Group, western Newfoundland: the interaction between eustasy and tectonics; Canadian Journal of Earth Sciences, v. 24, p. 1927-1951.

Kretz, R.
1983: Symbols for rock-forming minerals; American Mineralogist, v. 68, p. 277-279.

Kumarapeli, P.S., Dunning, G.R., Pintson, H., and Shaver, J.
1989: Geochemistry and U-Pb zircon age of comenditic metafelsites of the Tibbit Hill Formation, Quebec Appalachians; Canadian Journal of Earth Sciences, v. 26, p. 1374-1383.

Lilly, H.D.
1963: Geology of Hughes Brook-Goose Arm area, west Newfoundland; Memorial University of Newfoundland, Geology Report, no. 2, 123 p.

Lindholm, R.M. and Casey, J.F.
1988: The significance of Oldhamia in the Blow Me Down Brook Formation, western Newfoundland (abstract); Geological Association of Canada, Program with Abstracts, v. 13, p. A74.
1989: Regional significance of the Blow Me Down Brook Formation, western Newfoundland: new fossil evidence for an Early Cambrian age; Geological Society of America Bulletin, v. 101, p. 1-13.

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

McKillop, J.H.
1963: Geology of the Corner Brook area, Newfoundland, with emphasis on the carbonate deposits; Memorial University of Newfoundland, Geology Report No. 1, 102 p.

Mezger, K., Hanson, G.N., and Bohlen, S.R.
1989: High-precision U-Pb ages of metamorphic rutile: application to the cooling of high-grade terranes; Earth and Planetary Science Letters, v. 96, p.106-118.

Miyashiro, A.
1973: Metamorphism and metamorphic belts; George Allen and Unwin, London, 492 p.

Owen, J.V.
1991: Geology of the Long Range Inlier, Newfoundland; Geological Survey of Canada, Bulletin 395, 89 p.

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

Popper, G.H.P.
1970: Paleobasin analysis and structure of the Anguille Group, west-central Newfoundland; Ph.D. thesis, Lehigh University, Bethlehem, Pennsylvania, U.S.A. 226 p.

Reading, H.G.
1980: Characteristics and recognition of strike-slip fault systems; in Sedimentation in Oblique-slip Mobile Zones, P.F. Ballance and H.G. Reading (ed.); International Association of Sedimentologists, Special Publication no. 4, p. 7-26.

Riley, G.C.
1957: Red Indian Lake (west half), Newfoundland; Geological Survey of Canada, Map 8-1957, scale 1 in. to 4 miles.

Rivers, T.
1983: Progressive metamorphism of pelitic and quartz-feldspathic rocks in the Grenville Province of western Labrador -tectonic implications of bathozone 6 assemblages; Canadian Journal of Earth Sciences, v. 20, p. 1791-1804.

Ryan, J.J.
1992: Stratigraphic, structural and metamorphic relations within the Corner Brook Lake thrust belt, western Newfoundland; B.Sc. (Honours) thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 111 p.

Spear, F.S. and Cheney, J.T.
1989: A petrogenetic grid for pelitic schists in the system SiO_2-Al_2O_3-FeO--MgO-K_2O-H_2O; Contributions to Mineralogy and Petrology, v. 101, p. 149-164.

Stacey, J.S. and Kramers, J.D.
1975: Approximation of terrestrial lead isotope evolution by a two-stage model; Earth and Planetary Science Letters, v. 26, p. 207-221.

Stenzel, S.R., Knight, I., and James, N.P.
1990: Carbonate platform to foreland basin: revised stratigraphy of the Table Head Group (Middle Ordovician), western Newfoundland; Canadian Journal of Earth Sciences, v. 27, p. 14-26.

Stevens, R.K.
1965: Geology of the Humber Arm, West Newfoundland; M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland.
1970: Cambro-Ordovician flysch sedimentation and tectonics in west Newfoundland and their possible bearing on a Proto-Atlantic Ocean; in Flysch sedimentology in North America, J. Lajoie (ed.); Geological Association of Canada, Special Paper 7, p. 165-178.

Stukas, V. and Reynolds, P.H.
1974: $^{40}Ar/^{39}Ar$ dating of the Long Range dikes, Newfoundland; Earth and Planetary Science Letters, v. 22, p. 256-266.

Swinden, H.S.
1991: Paleotectonic settings of volcanogenic massive sulphide deposits in the Dunnage Zone, Newfoundland Appalachians; Canadian Institute of Mining and Metallurgy Bulletin, v. 84, no. 946, p. 59-69.

Swinden, H.S., Jenner, G.A., Fryer, B.F., Hertogen, J., and Roddick, J.C.
1990: Petrogenesis and paleotectonic history of the Wild Bight Group: an Ordovician rifted island arc in central Newfoundland; Contributions to Mineralogy and Petrology, v. 105, p. 219-241.

Tuach, J., Dean, P.L., Swinden, H.S., O'Driscoll, C.F., Kean, B.F., and Evans, D.T.W.
1988: Gold mineralization in Newfoundland: a 1988 review; in Current Research; Newfoundland Department of Mines, Mineral Development Division, Report 88-1, p. 279-306.

Tucker, R.D. and Gower, C.F.
1994: A U-Pb geochronological framework for the Pinware Terrane, Grenville Province, Labrador; Journal of Geology, v. 102, p. 67-78.

Twenhofel, W.H. and MacClintock, P.
1940: Surface of Newfoundland; Geological Society of America Bulletin, v. 51, p. 1665-1728.

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

van Gool, J.A.M. and Cawood, P.A.
1994: Frontal vs. basal accretion and contrasting particle paths in metamorphic thrust belts; Geology, v. 22, p. 51-54.

Waldron, J.W.F. and Milne, J.V.
1991: Tectonic history of the central Humber Zone, western Newfoundland Appalachians: post-Taconian deformation in the Old Man's Pond area; Canadian Journal of Earth Sciences, v. 28. p. 398-410.

Walthier, T.N.
1949: Geology and mineral deposits of the area between Corner Brook and Stephenville, western Newfoundland (Part I); Geology and mineral deposits of the area between Lewis Hills and Bay St. George, western Newfoundland (Part II); Newfoundland Geological Survey, Bulletin No. 35, 87 p.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and DiLabio, R.N.
1973: Age determinations and geological studies, K-Ar isotopic ages, Report II; Geological Survey of Canada, Paper 73-2A, 139 p.

Wanless, R.K., Stevens, R.K., Lachance, G.R., and Rimsaite, J.Y.H.
1965: Age determinations and geological studies: K-Ar isotopic ages, Report 5; Geological Survey of Canada, Paper 64-17, Part 1, 136 p.

Weitz, J.L.
1953: Geology of the Bay of Islands Area, Western Newfoundland; Ph.D. thesis, Yale University, U.S.A.

Werner, H.J.
1956: The Geology of the Humber Valley, Newfoundland; Ph.D. thesis, Syracuse University, Syracuse, New York, U.S.A., 111 p.

Whalen, J.B.
1989: The Topsails igneous suite, western Newfoundland: an Early Silurian subduction-related magmatic suite?; Canadian Journal of Earth Sciences, v. 26, p. 2421-2434.

Whalen, J.B., Currie, K.L., and Piasecki, M.A.J.
1993: A re-examination of relations between Dunnage subzones in southwest Newfoundland; in Current Research, Part D; Geological Survey of Canada, Paper 93-1D, p 65-72.

Whalen, J.B., Currie, K.L., and van Breemen, O.
1987: Episodic Ordovician-Silurian plutonism in the Topsails igneous terrane, western Newfoundland; Transactions of the Royal Society of Edinburgh: Earth Sciences, v. 78, p. 17-28.

Williams, H.
1977: Ophiolitic melange and its significance in the Fleur de Lys Supergroup, northern Appalachians; Canadian Journal of Earth Sciences, v. 12, p. 1874-1894.
1978: Tectonic-lithofacies map of the Appalachian orogen (1:1 000 000); Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, Map 1.
1979: Appalachian Orogen in Canada; Canadian Journal of Earth Sciences, v. 16, p. 792-807.

Williams, H. (ed.)
in press: Geology of the Appalachian-Caledonian Orogen in Canada and Greenland; Geological Survey of Canada, Geology of Canada, no. 6.

Williams, H. and Cawood, P.A.
1986: Relationships along the eastern margin of the Humber Arm Allochthon between Georges Lake and Corner Brook, western Newfoundland; in Current Research, Part A; Geological Survey of Canada, Paper 86-1A, p. 759-765.
1989: Geology, Humber Arm Allochthon, western Newfoundland; Geological Survey of Canada, Map 1678A, scale 1:250 000.

Williams, H. and Hiscott, R.N.
1987: Definition of the Iapetus rift-drift transition in western Newfoundland; Geology, v. 15, p. 1044-1047.

Williams, H. and St-Julien, P.
1978: The Baie Verte-Brompton Line in Newfoundland and regional correlations in the Canadian Appalachians; in Current Research, Part A; Geological Survey of Canada, Paper 78-1A, p. 225-229.
1982: The Baie Verte-Brompton Line: Continent-Ocean interface in the Northern Appalachians; in Major Structural Zones and Faults of the Northern Appalachians, P. St-Julien and J. Beland (ed.); Geological Association of Canada, Special Paper 24, p. 177-207.

Williams, H. and Stevens, R.K.
1974: The Ancient Continental Margin of Eastern North American; in The Geology of Continental Margins, C.A. Burke and C.L Drake (ed.); Springer-Verlag Publisher, p. 781-796.

Williams, H., Colman-Sadd, S.P., and Swinden, H.S.
1988: Tectonic-stratigraphic subdivisions of central Newfoundland; in Current Research, Part B; Geological Survey of Canada, Paper 88-1B, p. 91-98.

Williams, H., Gillespie, R.T., and Knapp, D.A.
1982: Geology of Pasadena map-area, Newfoundland; in Current Research, Part A; Geological Survey of Canada, Paper 82-1A, p. 281-288.
1983: Geology of Pasadena map area, 12H/4, western Newfoundland; Geological Survey of Canada, Open File 928, scale 1:50 000.

Williams, H., Gillespie, R.T., and van Breemen, O.
1985: A late Precambrian rift-related igneous suite in western Newfoundland; Canadian Journal of Earth Sciences, v. 22, p. 1727-1735.

Williams, H., Dickson, W.L., Currie, K.L., Hayes, J.P., and Tuach, J.
1989a: Preliminary report on a classification of Newfoundland granitic rocks and their relation to tectonostratigraphic zones and lower crustal blocks; in Current Research, Part B; Geological Survey of Canada, Paper 89-1B, p. 47-53.

Williams, H., Piasecki, M.A.J., and Colman-Sadd, S.P.
1989b: Tectonic relationships along the proposed central Newfoundland Lithoprobe transect and regional correlations; in Current Research, Part B; Geological Survey of Canada, Paper 89-1B, p. 55-66.

Williams, S.H.
1989: New graptolite discoveries from the Ordovician of central Newfoundland; in Current Research; Newfoundland Department of Mines, Geological Survey of Newfoundland, Report 89-1, p. 149-157.

Yardley, B.W.D.
1989: An introduction to metamorphic petrology; Longman Earth Sciences Series, Longman, Essex, 248 p.