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STRUCTURAL DENUDATION OF SILURIAN–DEVONIAN HIGH-GRADE METAMORPHIC ROCKS AND POSTOROGENIC DETACHMENT FAULTING IN THE MARITIMES BASIN, NORTHERN NOVA SCOTIA

Gregory Lynch

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Cover illustration

Outcrop of Namurian Hastings Formation from the Mabou Group, comprising interbedded shale, siltstone, and stromatolitic dolostone exposed along the banks of the Margaree River in central Cape Breton Island. View is looking northeast and strata are gently dipping to the southeast. Streams and river banks provide main access to outcrops inland from the coastal exposures on the island. Much of the Mabou Group in western Cape Breton Island is allochthonous above the Ainslie detachment triggered during postorogenic salt-controlled gravity slide. Photograph by G. Lynch. GSC 2000-036

Author's address

Shell Canada Limited
400-4th Avenue S.W.
P.O. Box 100, Station M
Calgary, Alberta
T2P 2H5

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Abstract

In northern Nova Scotia, the Canadian Appalachian belt underwent dynamic orogenic build-up and post-orogenic collapse, from Silurian to Carboniferous time. Early shortening and high grade metamorphism occurred during the Acadian event, followed by crustal extension and sedimentation within the Maritimes Basin. Three regional-scale fault structures which, in part, accomodated this orogenic cycle are described here, namely (1) the Highlands Shear Zone, (2) the Margaree Shear Zone, and (3) the Ainslie Detachment.

Neoproterozoic basement rocks were imbricated with Ordovician-Silurian cover sequences during compression. High grade gneissic units of late Silurian age in the region indicate that significant tectonic burial and crustal thickening occurred as a result of the thrusting. Partial denudation of the high grade assemblages occurred during Early Devonian thrust emplacement of the Cabot nappe toward the northwest, along the Highlands Shear Zone. The nappe is characterized by an amphibolitic gneiss and high-grade schist complex defining a large folded klippe. Inverted metamorphic isograds occur along the margins of the nappe. Further unroofing of the metamorphic rocks took place during Late Devonian extension from beneath the low-angle Margaree Shear Zone, which was eventually incised by Tournaisian coarse clastic rocks of the Horton Group. Tectonic sag of the extensional complex is marked by a cover of marine carbonates and evaporites from the Visean Windsor Group. Late Carboniferous block faulting in the basement and uplift of Cape Breton Island as a horst, triggered a regional scale evaporite-controlled bedding-parallel gravity slide near the base of the Windsor Group, defining the Ainslie Detachment.

Résumé

Dans le nord de la Nouvelle-Écosse, la ceinture orogénique des Appalaches canadiennes a été le siège d'une édification orogénique dynamique et d'un effondrement postorogénique, qui se sont échelonnés du Silurien au Carbonifère. L'événement acadien s'est traduit initialement par un raccourcissement et un métamorphisme de degré élevé et, par la suite, par une distension crustale et une sédimentation dans le Bassin des Maritimes. Trois systèmes de failles d'amplitude régionale, qui ont en partie canalisé la déformation associée à ce cycle orogénique, sont décrits dans le présent document, à savoir (1) la zone de cisaillement de Highlands, (2) la zone de cisaillement de Margaree et (3) le détachement d'Ainslie.

Au cours de la compression, les roches du socle néoprotérozoïque ont été imbriquées avec les successions de couverture de l'Ordovicien et du Silurien. La présence, dans la région, d'unités gneissiques de haut degré de métamorphisme du Silurien tardif indique qu'un enfouissement tectonique d'envergure et un épaississement crustal se sont produits sous l'effet du charriage. Les assemblages de haut degré de métamorphisme ont été en partie dénudés au cours du Dévonien précoce lors de la mise en place de la nappe de Cabot, qui s'est déroulée par un charriage en direction du nord-ouest le long de la zone de cisaillement de Highlands. Cette nappe se distingue par la présence de gneiss à amphiboles et d'un complexe de schistes de haut degré de métamorphisme don't l'arrangement défini une grande klippe plissée. Des isogrades inverses de métamorphisme peuvent être définies le long des bordures de la nappe. Au cours de la distension du Dévonien tardif, la dénudation des roches métamorphiques s'est poursuivie sous la zone de cisaillement faiblement inclinée de Margaree. Par la suite, cette zone de cisaillement a été incisée et les roches détritiques à granulométrie grossière du Groupe de Horton du Tournaisien se sont déposées sur la surface d'érosion. L'affaissement tectonique du complexe de distension est révélé par la présence d'une couverture de roches carbonatées de milieu marin et de roches évaporitiques appartement au Groupe de Windsor du Viséen. Un morcellements par failles du socle et un soulèvement de l'île du Cap-Breton sous forme d'un horst au Carbonifère tardif, ont provoqué un glissement gravitaire d'importance régionale le long d'une surface parallèle à la foliation située à proximité de la base du Groupe de Windsor et don't la position a été régie par la présence d'évaporites. Cette surface de glissement délimite le détachement d'Ainslie.

SUMMARY

The geology of Cape Breton island is highly complex, comprising a great diversity of lithologies that formed at different periods, spanning the interval from Neoproterozoic to late Paleozoic. Particularly striking in the Highlands are the range of crustal levels exposed at the present-day surface, and the abundance of mylonite which accomodated transport of these rocks. Much of this allochthonous transport occurred intermittently during Silurian to Carboniferous time when the Canadian Appalachian belt underwent an intense tectonic transformation, following Ordovician closure of the Iapetus ocean. This report attemps to classify, correlate, and characterize some of the major shear zones which were active during this phase of orogenesis. In northern Nova Scotia deformation is characterized by early shortening and high grade metamorphism, followed by later crustal extension and sedimentation within the Magdalen Basin. Three regional-scale fault structures which accomodated and influenced this orogenic cycle are now recognized and described here, namely (1) the Highlands Shear Zone and Cabot nappe, (2) the Margaree Shear Zone, and (3) the Ainslie Detachment.

Early Silurian pyroclastic units, calk-alkaline volcanics, and associated coarse siliciclastic rocks in northern Nova Scotia occur as part of a vast Appalachian overlap assemblage deposited unconformably on accreted terranes. The overlap assemblages suggest that amalgamation of terranes was largely complete by Silurian time. Compressional deformation with south directed transport followed Early Silurian volcanism and resulted in imbrication of Neoproterozoic basement rocks with Ordovician-Silurian cover sequences across thick zones of mylonite. High grade metamorphism and gneissic units of late Silurian age in the region indicate that significant tectonic burial and crustal thickening occurred as a result of the thrusting. Partial denudation of the high grade assemblages occurred during Early Devonian thrust emplacement of the Cabot nappe toward the northwest, along the Highlands Shear Zone. The nappe is characterized by an amphibolitic gneiss and high-grade schist complex defining a large folded klippe blanketing much of the Cape Breton Highlands. A distinctive feature of the thrust sheet is associated inverted metamorphism; cooler greenschist-grade rocks occur beneath the nappe, and staurolite is regionally distributed in pelitic units in the immediate footwall of the Highlands Shear Zone forming a discontinuous halo around the klippe. Greenschist-grade footwall rocks are exposed in structural windows as a result of folding and faulting, and the hinge zone of the synclinorium is exposed in the Middle River area of central Cape Breton Island. Pressure- temperature determinations on garnet amphibolites indicate that peak metamorphism in the Cabot nappe reached upper amphibolite conditions, of >700°C and > 8 kilobars, in Late Silurian time. In contrast metamorphic conditions in the immediate footwall to the Cabot nappe adjacent to the Highlands Shear Zone did not exceed approximately 650°C and 6.5 kilobars, and cooling continued into Early Devonian time. Mineralogical

SOMMAIRE

La géologie de l'île du Cap-Breton est très complexe et s'exprime par une grande variété de lithologies qui se sont formées à différentes périodes dans l'intervalle du Néoprotérozoïque au Paléozoïque tardif. L'un des aspects les plus frappants des hautes terres du Cap-Breton repose dans la grande variété de niveaux crustaux représentés à la surface et l'abondance de mylonites qui marquent les lieux de transport de ces roches. Le transport de ces allochtones a eu lieu par intermittence du Silurien au Carbonifère alors que la ceinture orogénique des Appalaches canadiennes subissait une intense transformation tectonique après la fermeture de l'Océan Iapetus, à l'Ordovicien. Le présent rapport tente de classifier, de mettre en corrélation et de caractériser certaines des principales zones de cisaillement qui étaient actives durant cette phase de l'orogenèse. Dans le nord de la Nouvelle-Écosse, la déformation s'est traduite initialement par un raccourcissement et un métamorphisme de degré élevé puis, par la suite, par une distension crustale qui a été accompagnée par une sédimentation dans le bassin de la Madeleine. Trois systèmes de failles d'importance régionale ayant favorisé la réalisation de cycle orogénique et influencé son déroulement ont été mises en évidence et sont décrites dans le présent ouvrage, soit (1) la zone de cisaillement de Highlands, (2) la zone de cisaillement de Margaree et (3) le détachement d'Ainslie.

Dans le nord de la Nouvelle-Écosse, des unités pyroclastiques, des roches volcaniques calco-alcalines et des roches silicoclastiques à granulométrie grossière associées font partie d'un vaste assemblage de recouvrement appalachien du Silurien précoce, qui s'est déposé en discordance sur des terranes accrétés. Les assemblages de recouvrement laissent supposer que la réunion des terranes était en grande partie terminée au Silurien. Une déformation par compression s'étant traduite par un transport en direction du sud s'est produite après les manifestations volcaniques du Silurien précoce et a entraîné l'imbrication d'unités du socle néoprotérozïque avec des successions de couverture de l'Ordovicien et du Silurien le long d'épaisses zones de mylonites. La présence, dans la région, d'indices d'un métamorphisme de degré élevé et d'unités gneissiques du Silurien tardif indique que le charriage a provoqué un enfouissement tectonique d'envergure et un épaississement crustal. La dénudation partielle des assemblages de haut degré de métamorphisme a eu lieu au cours du Dévonien précoce pendant la mise en place de la nappe de Cabot, qui s'est déroulée par charriage à vergence nord-ouest le long de la zone de cisaillement de Highlands. La nappe se distingue par la présence de gneiss à amphiboles et d'un complexe de schistes de degré élevé de métamorphisme don't l'arrangement défini une grande klippe plissée qui recouvre en grande partie les hautes terres du Cap-Breton. Une des particularités de la nappe de charriage est liée à l'existence d'isogrades inverses de métamorphisme. Des roches du faciès des schistes verts témoignant d'un métamorphisme de plus faible température se trouvent sous la nappe; à l'échelle régionale, de la staurotide est répartie dans les unités pélitiques qui forment le mur immédiat de la zone de cisaillement de Highlands, formant une auréole discontinue autour de la klippe. Les roches du faciès des schistes verts du mur affleurent dans des fenêtres structurales formées à la faveur de plissements et du jeu de failles. La zone charnière du synclinorium affleure dans la région de Middle River, au centre de l'île du Cap-Breton. La détermination des conditions de pression et de température à partir d'amphibolites à grenats révèle que, dans la nappe de Cabot, le métamorphisme maximal s'est déroulé dans des conditions du faciès des amphibolites supérieur, celles-ci s'élevant à plus de 700 °C et à plus de 8 kbar au cours du

features of massive sulfide lenses from Silurian volcanic rocks are also investigated in this report. Furthermore, mesothermal gold mineralisation may be related to hydrothermal fluid circulation during emplacement of the Cabot nappe.

Silurien tardif. À l'opposé, dans le mur immédiat de la nappe de Cabot, tout à côté de la zone de cisaillement de Highlands, les conditions du métamorphisme n'ont pas dépassé environ 650 °C et 6,5 kbar et le refroidissement s'est poursuivi au cours du Dévonien précoce. Les caractéristiques minéralogiques des lentilles de sulfures massifs contenues dans les roches volcaniques du Silurien sont également étudiées dans ce rapport. D'autre part, une minéralisation d'or mésothermale est sans doute associée à la circulation de fluides hydrothermaux au cours de la mise en place de la nappe de Cabot.

Exhumation of the Cabot nappe and its footwall assemblages occurred during Late Devonian extension along the low-angle Margaree Shear Zone. The shear zone consists of thick shallow-dipping retrogressive mylonite and ultramylonite overprinted by cataclastic horizons, brittle detachment faults, and chloritic breccia. The shear zone crosscuts and transports low-grade Late Devonian bimodal volcanic rocks and conglomerate of the Fisset Brook Formation in its hanging wall towards the west-southwest, and juxtaposes them against exhumed medium to high grade metamorphic rocks and basement in its footwall. Major and trace element geochemistry from mylonitic basalt near the top of the shear zone demonstrate that the Fisset Brook Formation has been affected by the shearing. Locally the mylonite is 200-1000 m thick and outcrops discontinuously along the southern and southwestern margins of the Cape Breton Highlands. Tournaisian coarse clastic units including debris flow deposits of the Horton Group unconformably overlie ductile mylonite of the Margaree Shear Zone, placing an upper limit on shearing. The Margaree Shear Zone is interpreted to be a thick brittle-ductile low-angle extensional fault which was active in mid- to Late Devonian time, at the initiation of the Maritimes Basin. Crustal-scale thinning, a pronounced gravity high in the Gulf of Saint Lawrence, and the exhumation of high grade metamorphic rocks on Cape Breton Island relate to the extension that was partly accomodated by the Margaree Shear Zone.

L'exhumation de la nappe de Cabot et des assemblages qui composent son mur s'est effectuée au cours d'un épisode de distension au Dévonien tardif, par un déplacement le long de la zone de cisaillement faiblement inclinée de Margaree. Cette zone de cisaillement est formée d'épaisses unités à faible pendage de mylonites et d'ultramylonites rétromorphosées, auxquelles se superposent des horizons de roches cataclastiques, des failles de détachement à comportement fragile et des brèches chloriteuses. La zone de cisaillement de Margaree traverse la succession de roches volcaniques bimodales et de conglomérats faiblement métamorphisés du Dévonien tardif de la Formation de Fisset Brook et a également servi au transport de cette unité. Cette succession, qui forme le toit de la zone de cisaillement, a été transportée vers l'ouest- sud-ouest et repose sur des roches exhumées de degré de métamorphisme moyen à élevé et le socle qui forment le mur. La géochimie des éléments majeurs et en traces des basaltes mylonitisés présents près du sommet de la zone de cisaillement révèle que la Formation de Fisset Brook a été l'objet d'un cisaillement. Les mylonites montrent, par endroits, une épaisseur de 200 à 1 000 m et elles affleurent par intermittence le long des bordures sud et sud-ouest des hautes terres du Cap-Breton. Des unités détritiques à granulométrie grossière du Tournaisien renfermant des dépôts de coulées de débris du Groupe de Horton reposent en discordance sur les mylonites ductiles de la zone de cisaillement de Margaree, établissant ainsi une limite d'âge supérieure pour l'époque du cisaillement. Cette zone de cisaillement serait une faille de distension faiblement inclinée à comportement fragile-ductile de forte épaisseur qui était active au Dévonien moyen et tardif, au moment où le Bassin des Maritimes prenait forme. L'amincissement crustal, une crête gravimétrique prononcée dans le golfe du Saint-Laurent et l'exhumation de roches de haut degré de métamorphisme dans l'île du Cap-Breton sont liés à la distension don't les effets ont été en partie canalisés par la zone de cisaillement de Margaree.

Regional transgression, and marine flooding of the extensional complex is marked by the thick Visean carbonate and evaporite deposits of the Windsor Group. The carbonates reflect a relatively quiescent tectonic regime, and broad subsidence was likely due to the cessation of extension-related thermally induced buoyancy in a tectonically thinned crust. Late Carboniferous block faulting in the basement and uplift of Cape Breton Island as a horst, triggered a regional scale evaporite-controlled bedding-parallel gravity slide near the base of the Windsor Group, which defines the Ainslie Detachment. The front end of the gravity slide is characterized by compressional structures and a broad diapir field to the west of Cape Breton Island in the Gulf of St. Lawrence, whereas the back end is characterized by extensional structures, listric normal faults and stratigraphic gaps in the succession directly above the Ainslie Detachment onshore. Hydrothermal

Une transgression régionale et une submersion marine du complexe de distension sont révélées par l'épais dépôt de roches carbonatées et de roches évaporitiques d'âge viséen du Groupe de Windsor. Les roches carbonatées témoignent d'un régime tectonique relativement calme et la subsidence de vaste étendue est sans aucun doute liée à l'interruption de la poussée vers le haut d'origine thermique, qui était associée à la distension dans une croûte amincie par la tectonique. Le morcellement par failles du socle et le soulèvement de l'île du Cap-Breton sous forme d'un horst au cours du Carbonifère tardif ont provoqué un glissement gravitaire d'importance régionale le long d'une surface parallèle à la foliation située à proximité de la base du Groupe de Windsor et don't la position a été régie par la présence d'évaporites. Cette surface de glissement délimite le détachement d'Ainslie. Les terrains situés directement à l'avant de ce glissement sont caractérisés par la présence de structures de compression et celle d'un vaste champ de diapirs situé à l'ouest de l'île du Cap-Breton, dans le golfe du Saint-Laurent, alors que les terrains situés alteration and fluid flow along the detachment resulted in Pb-Zn mineralization in carbonate units of the lower Windsor Group.

tout juste à l'arrière se distinguent par l'existence de structures de distension, de failles normales listriques et de lacunes stratigraphiques dans la succession présente sur la terre ferme, directement au-dessus du détachement d'Ainslie. L'altération hydrothermal et la circulation de fluides le long du détachement ont donné naissance à une minéralisation de Pb-Zn dans les unités carbonatées de la partie inférieure du Groupe de Windsor.

INTRODUCTION

Cape Breton Island and northern Nova Scotia are situated in the central Canadian Appalachian Mountains, along the axis of the St. Lawrence basement promontory, between Newfoundland to the northeast and New Brunswick to the west, at a juncture where the main lithotectonic zones of the Appalachian orogenic belt have been structurally telescoped (e.g. Williams, 1979). Unconformably overlying these lithotectonic zones and the deformation events which have affected them, is the Maritimes Basin (e.g. Howie and Barss, 1975) which comprises a large postorogenic intercontinental Late Devonian to Permian stratigraphic succession with a well defined internal stratigraphy (Fig. 1). Both the Maritimes Basin and its basement units have been the subject of intense geological study over the past thirty years, due to their accessibility in southern Canada and because of the significant resource endowment of metallic and industrial minerals, as well as coal and significant hydrocarbon accumulations. Although numerous mapping campaigns have covered both basement and Devonian–Carboniferous cover, typically the younger Maritimes Basin and older underlying units of the Appalachian orogenic belt have been studied as separate, largely unrelated entities. On the one hand, investigations of Lower Paleozoic rocks have emphasized terrane definition, amalgamation, and associated accretionary processes; whereas sedimentology and detailed stratigraphic analysis have been emphasized for the Upper Paleozoic basin which is thought to have evolved largely in a postorogenic strike-slip setting (e.g. Bradley, 1982). Nonetheless an ever increasing geochronological database indicates that significant temporal overlap exists between early stratigraphic units at the base of the Maritimes Basin succession and metamorphism and/or plutonism in the underlying basement, allowing for a genetic link to be inferred between basement processes and basin formation. This temporal overlap is particularly evident from age determinations made in basement units of the Cape Breton Highlands in northern Nova Scotia (e.g. Jamieson et al., 1986; Reynolds et al., 1989; Dunning et al., 1990a; Keppie et al., 1992; Dallmeyer and Keppie, 1993; Barr et al., 1995), of the Liscomb Complex of central Nova Scotia (Clarke et al., 1993), and of the Meguma terrane and South Mountain Batholith of southern Nova Scotia (Keppie and Dallmeyer, 1995). Cape Breton Island in particular provides a large window through the centre of the Maritimes Basin to high-grade basement units metamorphosed from the Silurian into Late Devonian. Rapid structural exhumation of the high-grade rocks to surface at the inception of Maritimes Basin subsidence is one of the main problems addressed in this bulletin. Furthermore, debate continues regarding terrane definition and the nature of the boundaries separating terranes. Consequently a second objective of this research has been to study some of the fault zones along these major basement boundaries. This has been accomplished through new regional mapping with emphasis on well known major faults, as well as newly recognized faults, establishing a structural framework that accommodates juxtaposition of high-grade metamorphic rocks with nonmetamorphosed rock units of the Maritimes Basin. Significant new advances have also been made with regards to upper Carboniferous salt tectonics, and a newly defined regional-scale gravity slide has been recognized, providing insight into the establishment of the offshore diapir field in the Gulf of St. Lawrence.

Mapping in Cape Breton Island was conducted at the 1:50 000 scale during the summers of 1991–1995, with the new data resulting in the production of a number of Open File maps covering the central portion of the island (Lynch and Tremblay, 1992a; Lynch et al., 1993b, c; Lynch and Brisson, 1994; Lynch et al., 1995a, b, c). Previous geological maps covering segments of these areas emphasize either the basement geology (e.g. Raeside and Barr, 1992) or the Devonian–Carboniferous cover (e.g. Kelley, 1967); the objective here was to integrate these maps to provide complete coverage, to fill in areas where added information was required, and to establish the structural framework for the region. Work was supported by the Canada–Nova Scotia Cooperation Agreement on Mineral Development, by the Magdalen Basin NATMAP project, and by the Geological Survey of Canada. A primary goal of the NATMAP project was to generate the first 1:250 000 compilation of northern Nova Scotia and integrated digital database. A digital version of this map is available as GSC Open File D3564 (Lynch et al., 1998) the text of this bulletin provides descriptions and interpretations of some of the new major structural features which are outlined on the Open File maps. These include the Highlands shear zone and Cabot nappe (Lynch, 1996a), the Margaree shear zone (Lynch and Tremblay, 1994; Lynch, 1996b), and the Ainslie detachment (Lynch and Giles, 1996). Two of these structures, the Margaree shear zone and the Ainslie detachment, are shallow-dipping to flat-lying or bedding-parallel faults, whereas the Highlands shear zone has been folded but is interpreted to also reflect a large component of lateral transport. As such these are the first large fault structures to be recognized in this region containing important flat or shallow-dipping segments.

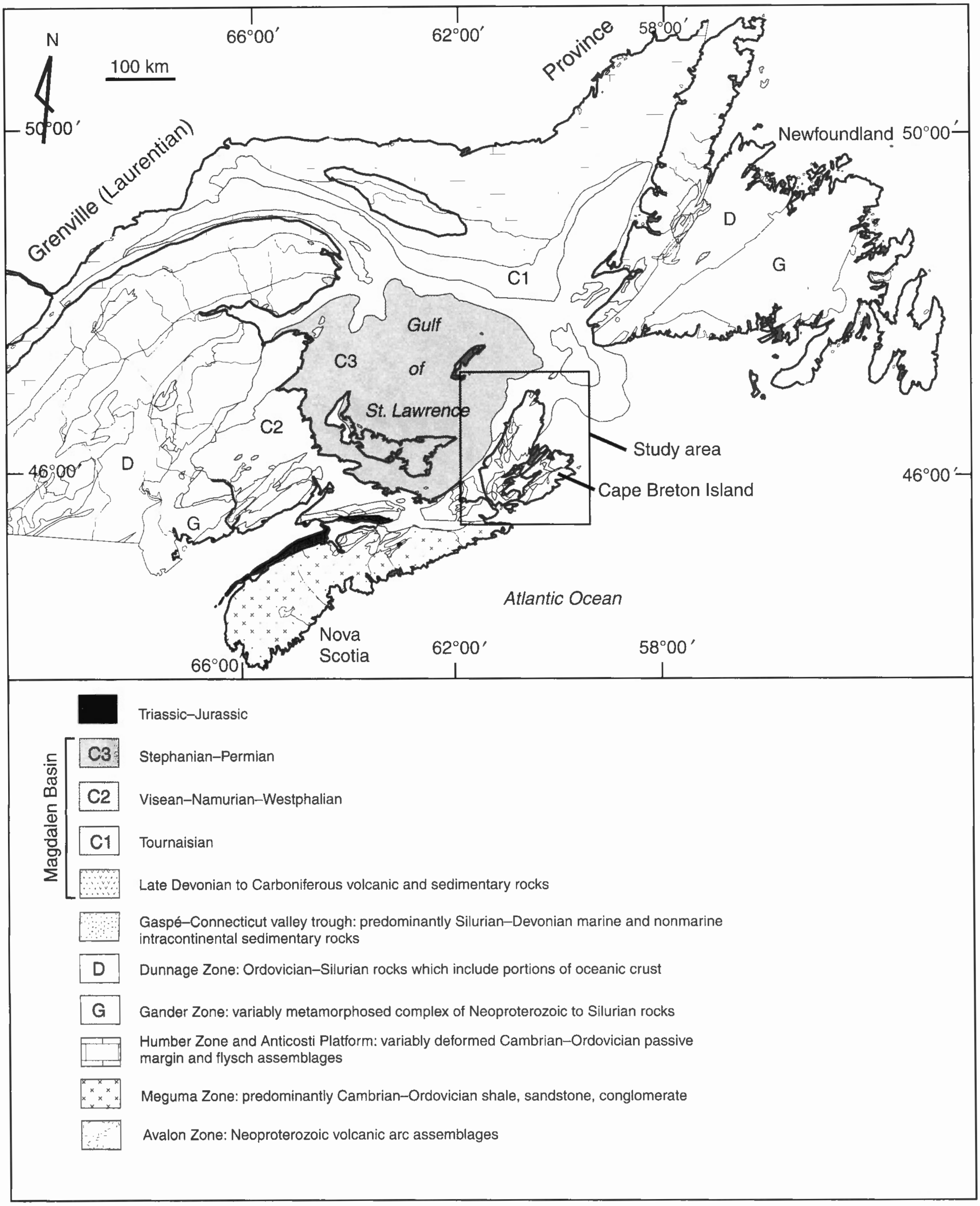


Figure 1. *Location map of study area (box), highlighting position of Magdalen Basin (grey shades) centred in the Gulf of St. Lawrence. Map displays major basins, tectonic zones, and terranes of the Canadian Appalachian orogen.*

SILURIAN–DEVONIAN THRUSTING AND LATE- TO POSTOROGENIC EXTENSION IN CAPE BRETON ISLAND

Geological setting

Cape Breton Island is contained within the hinterland of the Appalachian orogen where rocks have been affected by intense shortening and crustal thickening in relation to indentation with the St. Lawrence basement promontory of Laurentia (Williams, 1979; Williams and Hatcher, 1983; Keppie et al., 1992; Lynch, 1992; Dallmeyer and Keppie, 1993; Lin et al., 1994). The style of deformation is typical of that for the internal zones of orogens, featuring thick-skinned basement-involved thrusting, high-grade metamorphism and thermal inversions along major thrust faults, and the development of an advanced stage of late- to postorogenic extension (Lynch and Tremblay, 1994). The Cape Breton segment of the Appalachian orogen has been interpreted either as a collage of exotic terranes (Barr and Raeside, 1989; Barr, 1990; Dunning et al., 1990a) or as a partly contiguous stratigraphic succession of rocks exposed at different crustal levels, across a number of important unconformities (Keppie, 1989; 1990; Keppie et al., 1990a, b, 1992; Lynch and Tremblay, 1992a, b; Lynch et al., 1993a; Dallmeyer and Keppie, 1993). Implicit in both of these interpretations are large bounding faults either occurring at the margins of terranes or juxtaposing relatively deep and shallow crustal levels.

The oldest dated rocks in Cape Breton Island, which occur in the northern part of the island within the Blair River complex (Fig. 2), have been assigned to the Grenville Province (Laurentia) (Barr and Raeside, 1989; Miller et al., 1996); here U-Pb zircon dating indicates that the Sailor Brook gneiss has a minimum protolith age of 1217 Ma and was metamorphosed at 1035 +12/-10 Ma, whereas the Lowland Brook syenite has an igneous crystallization age of 1080 +5/-3 Ma (Miller et al., 1996). Other units contained within the Blair River complex include anorthosite and gabbro of Proterozoic age, amphibolitic and granulitic gneiss units of uncertain age, Silurian granite, as well as gneiss, metasedimentary rocks, paragneiss, and marble of the Mesoproterozoic Polletts Cove Group (Raeside and Barr, 1992). It should be mentioned that rocks of the Polletts Cove Group have been correlated regionally across Cape Breton Island with rocks of the Proterozoic George River Group based principally on the distribution and nature of marble and carbonate units (Milligan, 1970; Keppie, 1990; Keppie et al., 1990a, b; Hill, 1990), apparently linking basement segments (Keppie et al., 1992; Dallmeyer and Keppie, 1993). Although this correlation is debated (e.g. Barr and Raeside, 1989, 1990; Jamieson et al., 1991), Pb isotope values from sulphide occurrences within the marble units of the George River Group suggest that correlation across the island is a possibility (Sangster et al., 1990). The George River Group is interpreted as comprising a Proterozoic carbonate-detrital clastic shelf sequence typical of Gondwana basement units in the Avalon Zone (Keppie, 1989).

Avalonia rocks, which are characterized by 630–570 Ma arc-related volcanic and sedimentary successions and cogenetic plutons, are widespread across much of north-central Nova Scotia including southwestern and southeastern Cape Breton Island, and are part of a much broader terrane within the Appalachian–Caledonian orogen (Murphy et al., 1996). Eight major groups are defined within Proterozoic units of Avalonia in Nova Scotia. In south-central Cape Breton Island and the Antigonish Highlands, Cambrian–Ordovician shelf- to rift-related volcanic, clastic, and carbonate successions containing Acado–Baltica fauna are contained within Avalonia (Hutchinson, 1952; Weeks, 1954; Pickerill and Hurst, 1983; Landing, 1991; Landing and Murphy, 1991; White et al., 1995; Murphy et al., 1996) and unconformably overly the Late Proterozoic volcanic units. The Avalon Zone of the Canadian Appalachian Mountains (Fig. 1) is characterized by a vast Neoproterozoic volcano-plutonic arc system (Williams, 1984; Murphy et al., 1996). On Cape Breton Island, the Avalon Zone contains five volcanic belts (Barr et al., 1988), ranging from 680 to 575 Ma (Bevier et al., 1993). Typically, metamorphism is weak within the Proterozoic units, being of lower greenschist to subgreenschist grade. Rocks consist of volcanic flows and pyroclastic deposits ranging in composition from basalt, andesite, dacite, to rhyolite. Geochemical studies have identified a calc-alkalic trend (Dostal and McCutcheon, 1990; Barr, 1993) with trace element patterns indicating a continental margin arc setting (Dostal and McCutcheon, 1990). The volcanic rocks are crosscut by large composite plutons comprising granite, granodiorite, monzonite, and diorite (Chatterjee and Oldale, 1980; Barr, 1993).

Early Silurian calc-alkalic pyroclastic deposits and associated clastic rocks are regionally distributed across a wide portion of the Canadian Appalachian Mountains (Williams, 1979; Barr and Jamieson, 1991), occurring as a tectonic overlap assemblage (Williams, 1979; Chandler et al., 1987; O'Brien et al., 1991; Murphy et al., 1996) deposited unconformably upon various Ordovician or older terranes and basement units providing an apparent upper limit on the age of accretion. Characteristics and contact relations of the overlap assemblage are well exposed and documented in parts of northern Nova Scotia and southwestern Newfoundland where bimodal volcanic rocks or calc-alkalic assemblages were deposited under marine and nonmarine conditions above a major regional unconformity (O'Brien et al., 1991; Lynch and Tremblay, 1992a; Murphy et al., 1996). Geochemical aspects of the rocks suggest that volcanic activity was subduction related (Jamieson et al., 1990; Barr and Jamieson, 1991), although subduction-diagnostic rock types such as blueschist, ultramafic rocks, or trench mélange of similar age have not been found.

Stratified rocks of Late Silurian to Early Devonian age are lacking from north-central Cape Breton Island, which is a period characterized by intense deformation, metamorphism, uplift, and erosion. Metamorphic conditions for high-grade gneiss are reported to have peaked at approximately 424–411 Ma (Jamieson et al., 1986; Dunning et al., 1990a; Barr and

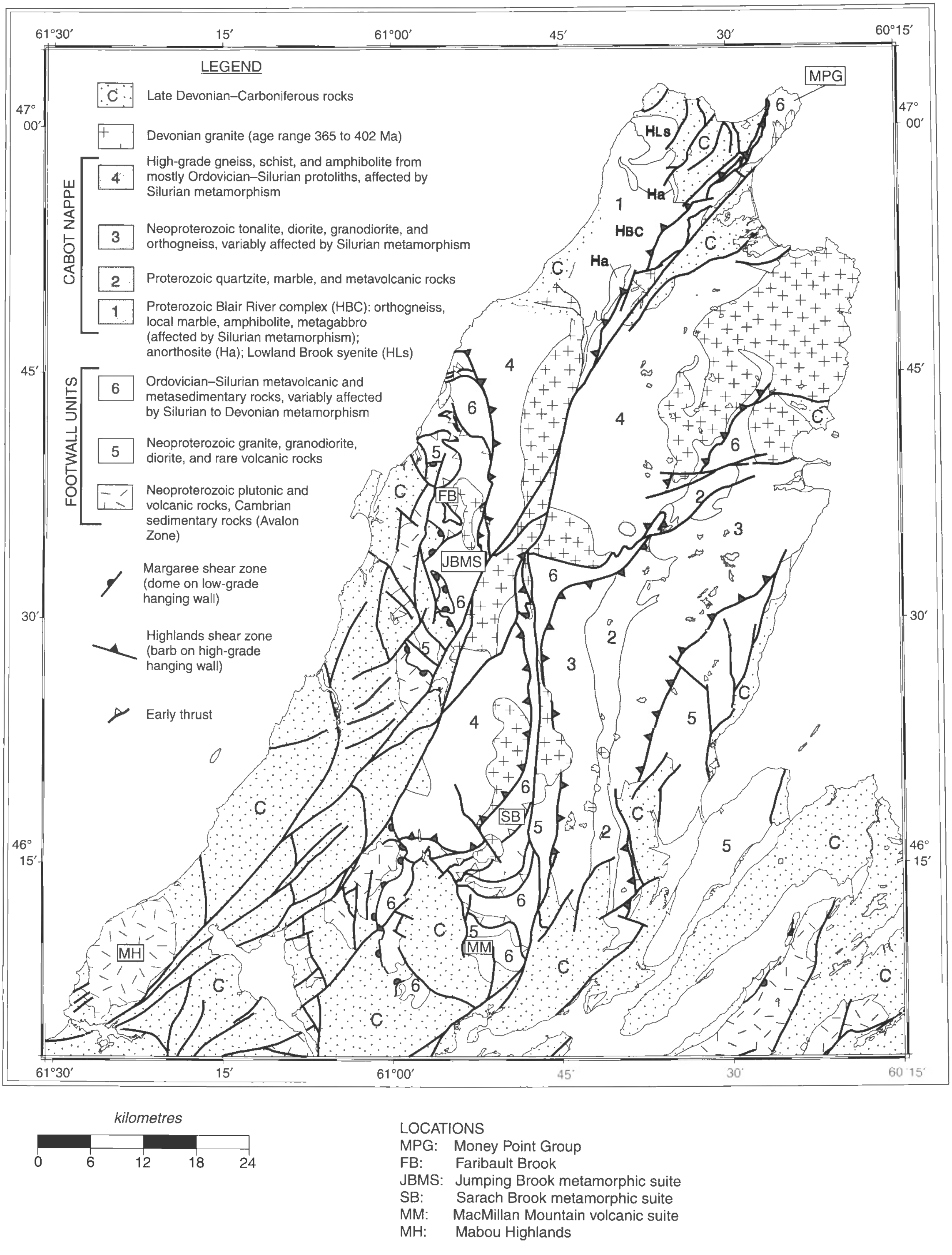


Figure 2. *Geological map of northern Cape Breton Island emphasizing outline of Cabot nappe and trace of Highlands shear zone.*

Jamieson, 1991) with conditions reaching 700–750°C at 8–10 kbar pressure during maximum burial (Doucet, 1983; Craw, 1984; Currie, 1987; Plint and Jamieson, 1989; Currie and Lynch, 1997). Uplift and cooling of amphibolite-grade rocks occurred at approximately 400–380 Ma (Doucet, 1983; Plint, 1987; Plint and Jamieson, 1989; Reynolds et al., 1989; Keppie et al., 1992; Dallmeyer and Keppie, 1993; Lynch and Mengel, 1995). Further uplift at lower metamorphic grades is recorded by muscovite and biotite cooling ages between 383 Ma and 360 Ma.

Post-tectonic Late Devonian high-level granitic plutons are widely dispersed throughout the region (Barr, 1990). These crosscut the principal structural fabrics and stitch fault zones placing an upper limit on deep-seated thrusting (373 ± 2 Ma, U-Pb monazite, for the Black Brook granite; Dunning et al. (1990a)). Bimodal basalt-rhyolite and interbedded redbeds of the Fisset Brook Formation form the basal members to the Maritimes Basin (Howie and Barss, 1975; Blanchard et al., 1984) and are cogenetic with the Late Devonian plutons (Jamieson et al., 1986). The area covered by the western Maritimes Basin is shown in Figure 1. The Maritimes Basin is a vast Devonian–Carboniferous accumulation of marine and nonmarine strata which unconformably overlies the principal tectonic zones of the northern Appalachian Mountains. The Maritimes Basin is proposed to have been initiated during a period of late orogenic to postorogenic extension and crustal thinning accommodated in part by faults such as the low-angle Margaree shear zone of western Cape Breton Island (Lynch and Tremblay, 1994). Coarse clastic sedimentation in the Tournaisian Horton Group occurred during horst-and-graben formation producing alluvial fan and fluviatile deposits (Hamblin and Rust, 1989). The overlying Viséan Windsor Group is dominated by extensive marine carbonate, evaporite, siltstone, and shale, which mark a general transgression and marine flooding (Giles, 1981), in relation to regional crustal sag of a tectonically thinned crust (Lynch and Tremblay, 1994). The Mabou and Cumberland groups consist of Late Carboniferous successions of predominantly nonmarine clastic rocks.

Ordovician–Silurian overlap assemblages

The distribution of Ordovician–Silurian rocks on Cape Breton Island is shown on Figure 2 (unit 6). The main groupings are described here and compared with well known correlative units of the Arisaig Group and White Rock Formation to the south. These rocks are significant to the structural history of the region as they characterize the relatively low-grade footwall to the overthrust Cabot nappe described in a the section 'Early thrusting, Highlands shear zone, and Cabot nappe'.

Money Point Group

The Money Point Group (Macdonald and Smith, 1980) of northern Cape Breton Island has been dated at 427 ± 4 Ma by the U-Pb method with zircon extracted from a metarhyolite (Keppie et al., 1992). An upper metasedimentary unit of pelitic to psammitic schist and conglomerate, overlies a lower volcanic unit dominated by schistose mafic to felsic pyroclastic rocks and rhyolite (Macdonald and Smith, 1980). Well preserved sections show clear primary bedding between members of tuff, basalt, andesite, rhyolite, and sedimentary rocks. Tuffaceous and pyroclastic units include fine-grained (?ash) tuffs, crystal tuffs, and a wide variety of coarse lithic tuffs dominated by lapilli and bombs of andesite and basalt as well as occasional pumice. Metasedimentary units consist of phyllite, schist, quartzite, quartz-pebble conglomerate, and calc-silicate rocks. Quartzite is gradational to wacke which is more feldspathic. Clasts of quartzite or polycrystalline quartz in the conglomerate are up to 5 cm in length, and individual beds are up to 5 m thick. The basal contact of the Money Point Group has not been identified.

Jumping Brook metamorphic suite

The Jumping Brook metamorphic suite is an informal unit name which includes a volcanic-sedimentary succession of rocks in western Cape Breton Island thought to be Ordovician–Silurian (Currie, 1982; Jamieson et al., 1987, 1990). The unit is apparently equivalent to the Money Point Group of northern Cape Breton Island (Macdonald and Smith, 1980), as well as to the Sarach Brook metamorphic suite of central Cape Breton Island (Barr and Jamieson, 1991; Lynch et al., 1993a), both of which have been dated by the U-Pb zircon method to be Late Ordovician to Early Silurian (Dunning et al., 1990a; Keppie et al., 1992). Bedding-concordant quartz-porphyry along Faribault Brook (Jumping Brook metamorphic suite) yielded a U-Pb zircon age of 432 Ma (Lin et al., 1997), similar to adjacent rhyolite dykes in basement units thought to be feeders to the volcanic rocks which have been dated at 439 ± 7 Ma by U-Pb on zircon (Currie, 1982); however, a tuffaceous bed in sedimentary rocks yielded a U-Pb zircon age of 551 Ma (Lin et al., 1997), but it is uncertain if this reflects the age of the tuff or if the zircon was derived from adjacent granitic basement units of similar age (e.g. Chéticamp pluton, 550 Ma). Recent investigations have described the lithological, metamorphic, and structural characteristics of the Jumping Brook metamorphic suite (Connors, 1986; Currie, 1987; Jamieson et al., 1987; Plint and Jamieson, 1989; Jamieson et al., 1990; Barr and Jamieson, 1991; Lynch and Tremblay, 1992a; Lynch et al., 1993a; Lynch and Mengel, 1995). The rocks are moderately to strongly metamorphosed, however volcanological and geochemical aspects of the rocks have been interpreted to indicate that volcanic activity was likely subduction-related in an arc setting (Jamieson et al., 1990; Barr and Jamieson, 1991), and that rocks from Faribault Brook have a relatively primitive geochemical character (Connors, 1986).

Two thick units or members are generally recognized in the Jumping Brook metamorphic suite; a lower unit which is predominantly volcanic in nature (Faribault Brook metavolcanic rocks, Jamieson et al. (1987)), and an upper clastic unit (includes the Barren Brook schist and Corney Brook schist of Jamieson et al. (1987)). A sheared contact separates the units. The volcanic unit is dominated by metabasalt which is pervasively sheared, isoclinally folded, and metamorphosed to upper greenschist and lower amphibolite facies. Nonetheless, primary volcanic features such as pillow structures can be observed locally. Interbedded siltstone, wacke, lapilli tuff, and minor rhyolite occur with the basalt. Medium-grained metadiorite sills and intrusions are dispersed throughout the succession and dominate the assemblage in places. The upper clastic unit typically includes well bedded siltstone, dark phyllite, or schist, as well as sandstone and conglomerate. Beds are thinly to thickly bedded, and graded bedding is common. Quartz-pebble conglomerate and wacke are widespread and typified by abundant blue quartz clasts. Arkose and conglomerate containing predominantly granitic clasts with also mafic and felsic volcanic clasts occur. The contact between the lower volcanic unit and upper sedimentary unit is sheared, and the relative stratigraphic positions of the two units remain unclear; however, because of the quartz-rich nature of the sandstone and wacke, and because coarse conglomerate contain a greater abundance of granitic clasts, it is possible that the sedimentary rocks do not lie directly on the mafic volcanic rocks but might be older and have been originally deposited upon a granitic basement.

Sarach Brook metamorphic suite

The Sarach Brook metamorphic suite (Barr et al., 1987b; Barr and Jamieson, 1991; Horne, 1995) is a well bedded succession consisting of fine-grained to aphanitic basalt, porphyritic andesite, lapilli and bomb felsic pyroclastic rocks, crystal tuff, epiclastic sandstone, quartz-rich wacke, conglomerate, and black slate. Zircon from felsic crystal tuff has been dated by the U-Pb method, indicating a crystallization age of 433 +7/-4 Ma (Dunning et al., 1990a). Well preserved portions show only moderate degrees of metamorphism. Conglomerate in the higher grade rocks from the western portion of the Sarach Brook metamorphic suite are highly strained and have been affected by upper greenschist-facies (biotite) metamorphism. Moderately sorted, well rounded clasts up to 5 cm in diameter are dominated by granitic clasts and epidotized diorite. Detrital feldspar and vitreous blue quartz clasts are abundant in some layers where the rock is arkosic in composition. A sharp contact was observed between the conglomerate and a large body of medium- to coarse-grained foliated diorite; the age of the diorite is unknown, but the contact is likely an unconformity. The conglomerate near the contact is approximately 4–10 m thick, and progresses upward to a thick succession of chloritic metawacke containing dispersed quartz or feldspar pebbles.

MacMillan Mountain Volcanic Suite

In the southern Cape Breton Highlands the MacMillan Mountain Volcanic Suite consists of relatively fresh, weakly deformed mafic, intermediate, felsic volcanic rocks, and pyroclastic rocks, with associated sedimentary rock types. Based on the similarities and distinctiveness of the volcanic assemblages Jamieson (1981) originally correlated the MacMillan Mountain Volcanic Suite with the Sarach Brook metamorphic suite. Collectively these were referred to as the 'Crowdis Mountain volcanics' (Jamieson, 1981).

In the MacMillan Mountain unit, dark green to brick red andesite is common, with, locally, varieties of trachyte. Basalt is closely associated with andesite and may be amygdaloidal. Basalt and andesite form thick, crudely bedded, blocky flows with vitrophyre along contacts. Rhyolite is fine grained to aphanitic, with occasional spherulites. Pyroclastic deposits are highly variable but typically well bedded. Laminated ash-flow tuffs and crystal tuffs are common. Welded lapilli tuffs occur as distinctive light coloured horizons containing pumice fragments. Sedimentary rocks include siltstone, coarse sand arkose, and lithic arkose. Pebble conglomerate with quartzite, and granitic clasts as well as quartz and feldspar clasts occurs along the northeastern limit of the MacMillan Mountain Volcanic Suite immediately adjacent to foliated granite of uncertain age as well as diorite and granodiorite (ca. 614 Ma, Jamieson et al., 1986). The exact nature of the contact is unknown, but the distribution and nature of rock assemblages suggest a likely unconformable relationship.

Arisaig Group, Antigonish Highlands

The Arisaig Group of the Antigonish Highlands on the mainland of Nova Scotia to the immediate southwest of Cape Breton Island consists of a 1400–1500 m thick stratigraphic succession dominated by siliciclastic rocks with lesser volcanic rocks. These unconformably overly a Cambrian–Ordovician succession of bimodal, intercontinental volcanic rocks, and terrestrial and shallow marine clastic rocks and limestone that contains typically Avalonian–Baltic fauna (Theokritoff, 1979; Landing and Murphy, 1991). At the base of the Arisaig Group, the Dunn Point Formation consists of approximately 80 m of bimodal intercontinental volcanic rocks. The Beechill Cove Formation disconformably overlies the Dunn Point Formation and consists of shallow marine conglomerate, sandstone, and siltstone. Paleontological evidence indicates that the Beechill Cove Formation is Llandoverian (Boucot et al., 1974; Pickerill and Hurst, 1983), and thus, is similar in age to the Money Point Group and Sarach Brook metamorphic suite of Cape Breton Island described above. Sedimentary facies recorded in the formation include conglomerate derived from progressive marine lag, red shale deposited in a near shore environment, bioturbated mudstone from a shallow subtidal domain, turbidite, lenticular facies derived from storm conditions within the subtidal environment, and laminated shale representing low-energy suspension deposition (Pickerill and Hurst, 1983). The succession grades upward into shale, siltstone, tuff, and arenaceous limestone of the Ross Brook Formation likely Early Silurian (Hurst and Pickerill, 1986), which is conformably overlain by a thick succession of middle to late Silurian interbedded mudstone, siltstone, and sandstone deposited in a near-shore environment (Boucot et al., 1974).

White Rock Formation

The White Rock Formation (Crosby, 1962; Lane, 1976) is an Upper Ordovician to Lower Silurian clastic and volcanic rock unit which directly overlies the Halifax Formation of the Meguma Group, covering a wide area in southwestern Nova

Scotia along the periphery of the Middle Devonian South Mountain Batholith. The White Rock Formation is up to approximately 500 m in thickness, and has been subdivided into upper and lower members. The lower member is composed of a laterally variable complex of arenite, siltstone, and shale, with the base of the member locally defined by a laterally extensive volcanic unit. The volcanic rocks are composed of bimodal dacite and rhyolite overlain locally by basalt. The felsic rocks, which commonly overly shale of the Halifax Formation, occur as massive devitrified ash-flow tuff and flow banded tuff, that vary laterally to water-laid bedded tuff. The basalt occurs as vesicular pillow lava and water-laid mafic tuff. The upper member consists of two distinctive, massive, quartz arenite units, separated by an intervening shale unit. The succession demonstrates a general coarsening-upward tendency, and contains crosslaminated and ripple crosslaminated beds. A regionally extensive felsic tuff horizon occurs along the upper contact of the upper member.

The strata of the White Rock Formation are interpreted to be mostly marine in origin, and a fossil collected from the White Rock Formation indicates a Caradocian or younger age (Lane, 1976). Further constraints on the age of the White Rock Formation are provided by paleontological data from the underlying Halifax and overlying Kentville formations, which restricts deposition of the White Rock Formation to between the Middle Ordovician and the Upper Silurian.

Significance of overlap assemblages

Early Silurian volcanic and sedimentary rocks are widespread in the Canadian Appalachian Mountains (Barr and Jamieson, 1991) and overlap all of the main lithotectonic zones which amalgamated in the Ordovician. Correspondingly these may comprise an overlap assemblage (Williams, 1979; Chandler et al., 1987; O'Brien et al., 1991), placing an upper age limit on ocean closure and tectonic accretion. The overlapping nature of these rocks is well displayed in Nova Scotia where the White Rock Formation is unconformable on the Meguma Group, whereas the Arisaig Group is unconformable on Avalonia rocks, and in Cape Breton Island the Money Point Group and correlative units including the Sarach Brook metamorphic suite, Jumping Brook metamorphic suite, and MacMillan Mountain Volcanic Suite are apparently unconformable above Neoproterozoic rocks of likely Avalonia affinity. Although Early Silurian volcanic rocks are not found within the Blair River complex at the northern tip of Cape Breton, Early Silurian plutonic rocks which intrude the Blair River complex (Miller et al., 1996) demonstrate that this lithotectonic domain (assigned to the Grenville by Miller et al. (1996)) was also affected by the magmatic event which characterizes the overlap assemblage. Furthermore geochemical characterization of the Early Silurian Beechill Cove Formation of the Arisaig Group has established the earliest Grenville-derived sediment input to Avalonia, which may be interpreted to indicate continent-continent amalgamation by the latest Early Silurian (Murphy et al., 1996).

Early thrusting, Highlands shear zone, and Cabot nappe

High-grade rocks are exposed in the Cape Breton Highlands where two kinematically and temporally distinct phases of intense Silurian and Devonian thrusting, imbrication, and crustal shortening characterize the main deformation events, designated the early thrusting event and the later Highlands shear zone. Different strands of the Highlands shear zone are described from geographically distinct areas and the outline of the Cabot nappe is shown in Figure 2. The Cabot nappe consists of a broad panel of mainly amphibolite-grade rocks, bound by zones of intense shear, and metamorphosed in the Late Silurian–Early Devonian. Arguments are presented to correlated fault segments from different areas which were previously thought to be different faults, into one large-scale folded shear zone referred to here as the Highlands shear zone.

Structurally significant study areas

Western Cape Breton Island

The Chéticamp area of western Cape Breton Island is situated along the western Cape Breton Highlands and is underlain by Late Proterozoic granitic rocks as well as Ordovician–Silurian volcanic and sedimentary rocks of the Jumping Brook metamorphic suite. The Chéticamp pluton is a large medium-grained granitic body consisting of biotite granite and biotite-muscovite granite. The pluton dominates ridge-tops (Fig. 3a) of the stream-incised topography and is bounded along its lower contact by the Chéticamp thrust, which is marked by granite mylonite (Fig. 3b). Footwall rocks consist of the younger Ordovician–Silurian metavolcanic and sedimentary rocks of the Jumping Brook metamorphic suite. The fault juxtaposes older rocks over younger units in a blanket-like manner, and the fault can be traced for 15 km from Jumping Brook in the north to the Chéticamp River in the south. The Chéticamp thrust is the oldest recognized fault structure in these rocks, and is a prominent feature of the early thrusting event. Mylonite along the thrust is approximately 100–200 m thick and comprises thinly laminated, light-coloured, very fine-grained mylonitized granite, containing varying proportions of augen, typically 5–15% isolated feldspar crystals (Fig. 3b). The mylonite has a strong north-trending lineation. Feldspar augen are typically segmented along crystallographic cleavages producing shingle structures, with polygonal quartz filling the interstices between porphyroclasts and minor amounts of albite rimming the augen. Locally rotated feldspar porphyroclasts are enveloped by highly asymmetric in-plane mineral trails demonstrating a strong component of simple shear (Fig. 3b); these typically show a top side to the south sense of motion. Minor amounts of epidote, chlorite, calcite, and muscovite have formed along segments of the Chéticamp thrust, in association with the development of C-S fabrics and shear bands within the fault.

A second fault, here interpreted as a thrust, approximately parallels the Chéticamp thrust and is located structurally beneath the Chéticamp thrust within underlying units of the Jumping Brook metamorphic suite. The fault juxtaposes the clastic unit consisting of pelitic schist, quartz and feldspar

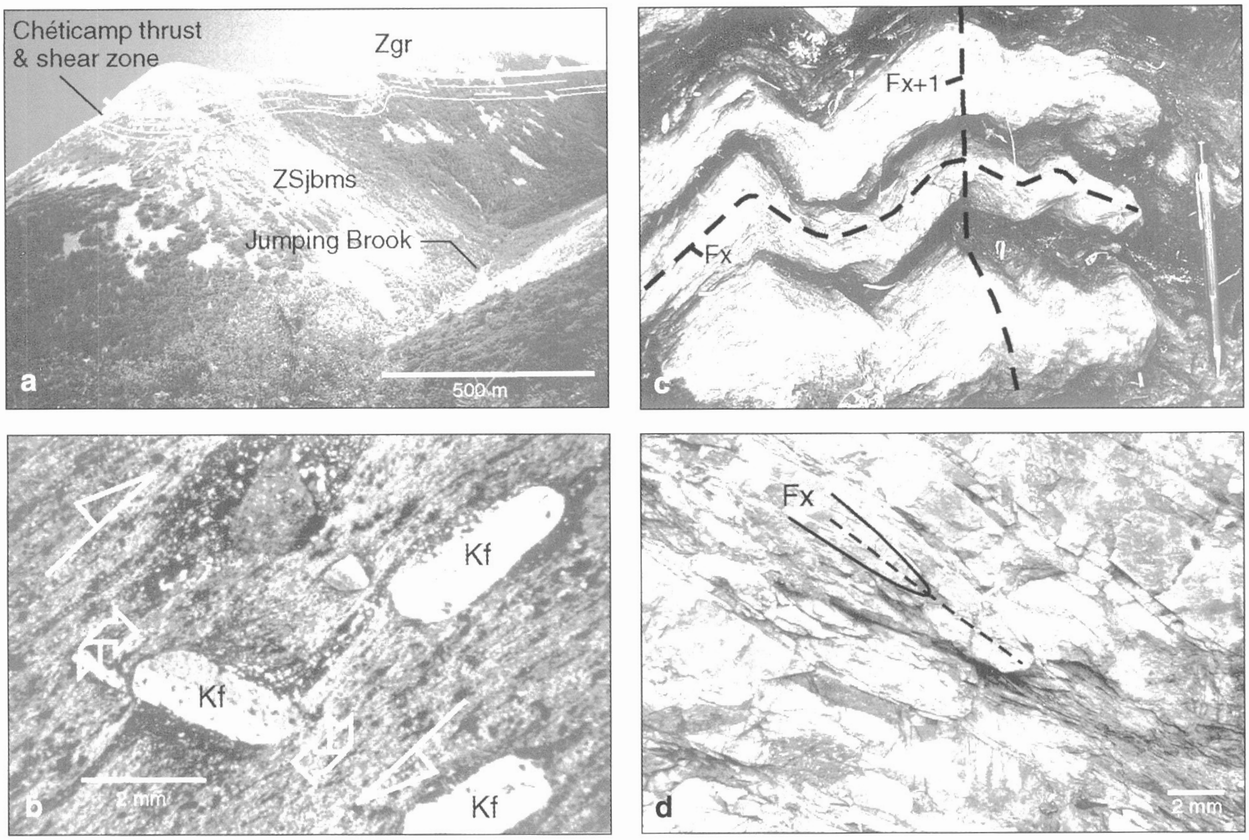
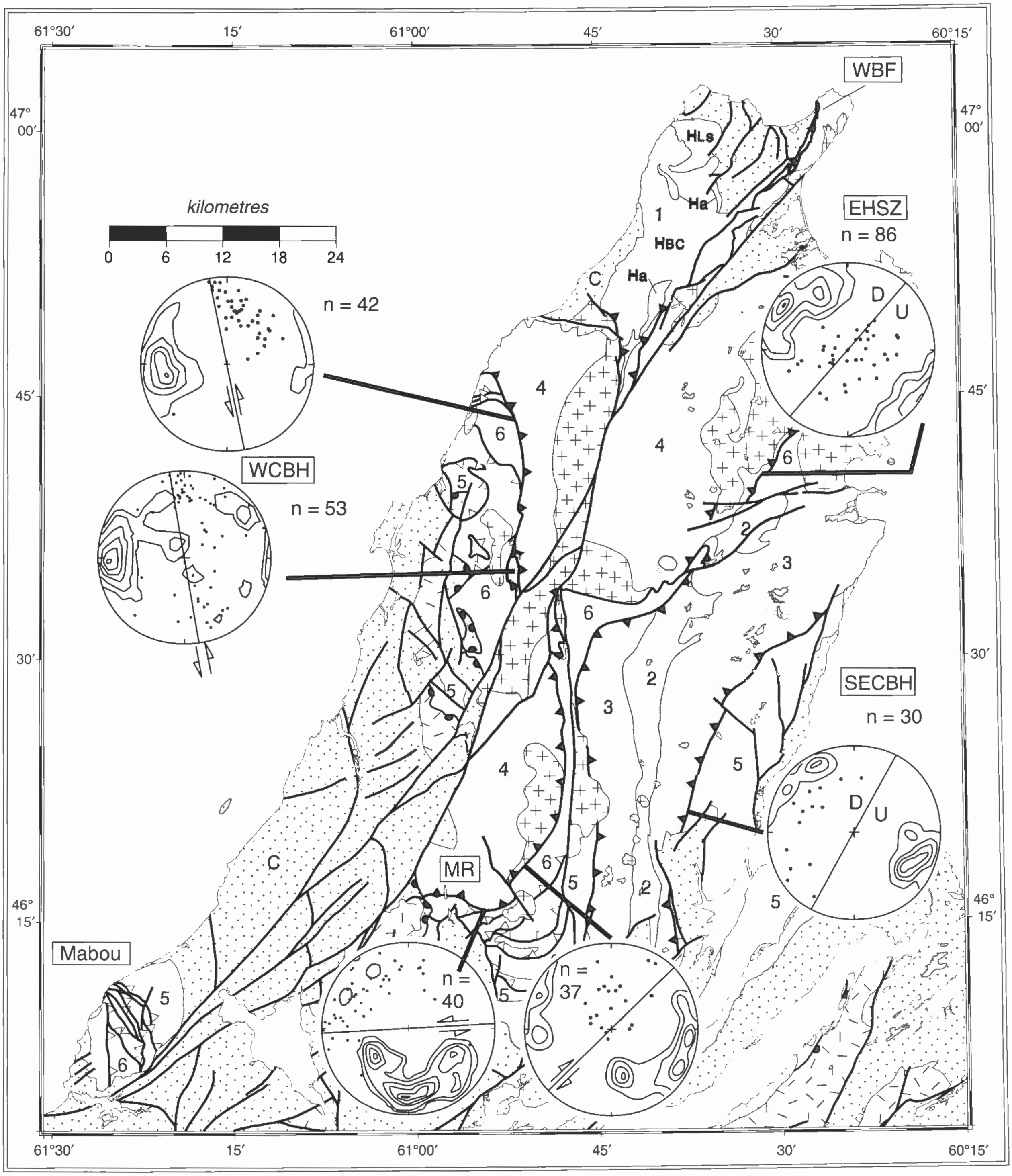


***Figure 3. a)** View looking north across Jumping Brook showing Chéticamp thrust placing Neoproterozoic granite of the Chéticamp pluton (*Zgr*) onto the Jumping Brook metamorphic suite (*ZSjbms*) base of photo to top of ridge approximately 1000 m (photograph by G. Lynch; GSC 2000-041A); **b)** thin section of mylonitized Chéticamp pluton along the Chéticamp Thrust displaying intense grain size reduction and preservation of potassium feldspar (Kf) porphyroclasts with asymmetric minerals trails indicating noncoaxial rotational shear (shear sense indicated by thick arrows and vorticity by small arrows) porphyroclasts approximately 2 mm long; **c)** isoclinally folded (F_x first axial plane) pelitic schist of the Jumping Brook metamorphic suite along Faribault Brook, overprinted by later, upright, north-trending folds (F_{x+1} second axial plane)(photograph by G. Lynch; GSC 2000-041C); **d)** isoclinally folded (F_x axial plane) psammitic schist of the Jumping Brook metamorphic suite situated in immediate footwall of Chéticamp thrust along the Cabot Trail next to Jumping Brook, rock hammer resting on nose of fold approximately 30 cm long (photograph by G. Lynch; GSC 2000-041B).*

pebble conglomerate, and wacke in the hanging wall, with the footwall unit dominated by mafic volcanic rocks, metadiorite, and chloritic schist. Shear is intensely developed along the contact between the units, and shear-sense indicators suggest south-directed transport of the hanging wall. A penetrative shallow-dipping schistosity occurs throughout the assemblages and is accompanied by a north-trending lineation. Recumbent isoclinal folds are widely distributed and feature an axial plane which parallels schistosity (Fig. 3c, d). Fabrics within and adjacent to the proposed thrust surface contain synkinematic garnet with sigmoidal inclusion trails (Fig. 3b), and locally well developed C-S fabrics defined by muscovite or chlorite. Detrital quartz clasts form asymmetrically winged inclusions and have been dynamically recrystallized. Quartz ribbons follow schistosity and contain crystals with a preferred crystallographic orientation and oblique extinction. Synkinematic quartz veins crosscut the shear fabric but are folded with axial planes parallel to the schistosity or occur as segmented boudins along shear planes. Map-scale and outcrop-scale tight to close folds have overprinted the mylonite, and the folds have upright, north-striking axial planes with gently north-plunging fold axes (Fig. 3d).

In western Cape Breton Island the Highlands shear zone defines a major north-trending shear zone, which juxtaposes amphibolite-grade metamorphic rocks of the Pleasant Bay complex (Currie, 1987) to the east with the greenschist-grade Ordovician–Silurian rocks of the Jumping Brook metamorphic suite to the west. Earlier east-trending structures and thrusts including the Chéticamp thrust are truncated by the Highlands shear zone. The Highlands shear zone dips steeply to the east (Fig. 4) separating highly contrasting metamorphic domains, and features a complex array of shear structures distributed across a width of several hundreds of metres. High-grade rocks in the Highlands shear zone include



LOCATIONS: WBF, Wilkie Brook fault; EHSZ, eastern Highlands shear zone; WCBH, western Cape Breton Highlands; SECBH, southeastern Cape Breton Highlands; MR, Middle River; Mabou, Mabou Highlands

Figure 4. *Lower hemisphere equal area projections of poles to mylonite fabrics (contoured) and stretching lineations (points) from the Highlands shear zone along the periphery of the Cabot nappe. General strike of planes and shear sense of fault shown by solid lines and arrows drawn on projections, with 'U' and 'D' indicating up and down, whereas 'n' is the number of planar fabric measurements. Note girdle defined by contours in southern domain (MR, Middle River) emphasize position in hinge of synclinorium or folded klippe. Contours are in per cent: WCBH north 1, 4, 8, 12, 16%; WCBH south 1, 2, 4, 6, 8, 10%; MR west 1, 2, 4, 6, 8, 10%; MR east 1, 3, 6, 9, 12%; SECBH 1, 2, 4, 6, 8, 10%; EHSZ 1, 5, 10, 15, 20%. Map legend same as in Figure 2* (modified from *Lynch (1996a)).*

kyanite-silimanite-bearing granodiorite orthogneiss, garnet amphibolite, pegmatitic augen gneiss, calcic-paragneiss, and various stages of mylonite development (Currie and Lynch, 1997).

Mylonitic gneiss and augen gneiss along the Highlands shear zone display varying degrees of retrogression. Locally, feldspars from the gneissic rocks have been altered to medium-grained mica, defining well developed C-S fabrics (Fig. 5a). Boudinaged pegmatite lenses, gneissic layers, and feldspar crystals form asymmetrically winged porphyroclasts (Fig. 5b) or are enveloped by rotated finer grained feldspathic mineral trails. Medium- to fine-grained quartz ribbons and feldspar ribbons typically define the mylonitic foliation. Synthetic or Riedel shear bands (R *in* Fig. 5b) occur en echelon at a low angle to the principal shear planes. Lineations are defined by mineral trails, lineated black amphibole crystals, or segmented tourmaline and plunge moderately to the north-northeast (Fig. 4). Shear-sense indicators, such as winged porphyroclasts, C-S fabrics, Riedel shear structures, and asymmetric folds, along the Highlands shear zone in western Cape Breton Island consistently demonstrate sinistral motion along the shear zone. Intrafolial folds occur between shear bands and display an 'S' geometry indicating sinistral motion within the shear zone (Fig. 5c). Pegmatite dykes occur along portions of the shear zone, occasionally crosscutting the shear fabrics, but more commonly the dykes are sheared and folded and in some cases are preserved only as very coarse-grained pegmatitic porphyroclast fragments. Very fine-grained laminated mylonite reflects intense grain-size reduction during retrogressive shear and is subordinate in abundance to the coarser grained mica-rich mylonitic gneiss.

A distinctive feature of the Highlands shear zone is the steep metamorphic gradient adjacent to the fault away from the gneissic rocks from within the Cabot nappe. Details of the metamorphism from western Cape Breton Island have been described by a number of authors (Craw, 1984; Currie, 1987; Plint, 1987; Plint and Jamieson, 1989; Lynch and Mengel, 1995; Currie and Lynch, 1997). High-grade, medium-grade, and low-grade metamorphic zones are recognized. The high-grade rocks consist of amphibolite-facies assemblages, with coexisting kyanite and orthoclase within the gneissic bodies (Barr and Jamieson, 1991). Medium-grade rocks to the west and adjacent to the Highlands shear zone contain coarse-grained staurolite porphyroblasts occurring mostly within pelitic metasedimentary units (Fig. 6) and form an aureole extending up to 2 km west of the Highlands shear zone (Fig. 7). The low-grade belt consists of greenschist-grade rocks west of the Highlands shear zone (Fig 6a), with locally chloritoid, garnet, and one occurrence of kyanite. Staurolite in pelitic units from the medium-grade belt occurs as coarse porphyroblasts up to 3 cm (Fig. 6b) and is found with large biotite porphyroblasts, as well as with garnet porphyroblasts which are up to 2 cm. The staurolite overgrows earlier schistosity related to the early thrusting event, and garnet has developed as inclusion-free mantles around garnet crystals defined by inclusion-rich cores. The metamorphic zonation clearly represents a gradient of decreasing pressure and temperature away from the gneissic bodies (e.g. Plint and Jamieson, 1989; Lynch and Mengel, 1995; Currie and Lynch, 1997), with the juxtaposition of the different crustal levels accommodated by the Highlands shear zone. Metamorphism of the Jumping Brook metamorphic suite is spatially related to the Highlands shear zone (Craw, 1984; Plint and Jamieson, 1989), and is characterized by inverted isograds (e.g. Ruppel and Hodges, 1994) within the Jumping Brook metamorphic suite which defines the footwall to the overthrust high-grade gneiss sheet or Cabot nappe.

Tight to close folds overprint the Highlands shear zone (Craw, 1984). The folds are upright and plunge moderately to the north or are noncylindrical and doubly plunging to the north and south. Locally, folds are polyclinal or ptygmatic. Finely laminated mylonite forms kink and box folds.

Peak metamorphism in the gneissic complex of the Cape Breton Highlands was determined from U-Pb (zircon) and $^{40}Ar/^{39}Ar$ (hornblende) to have occurred at ca. 424–396 Ma (Jamieson et al., 1986; Reynolds et al., 1989; Dunning et al., 1990a; Barr and Jamieson, 1991). Biotite growth was prolific near to the Highlands shear zone during shearing, occurring as synkinematic C-S surfaces within the shear zone or forming coarse porphyroblasts in the adjacent medium-grade metamorphic zone. Hornblende defines the lineation along portions of the shear zone. Four $^{40}Ar/^{39}Ar$ age determinations, for biotite and hornblende, in the proximity of the western Highlands shear zone (Fig. 8a) indicate ages of approximately 390–383 Ma for the metamorphism (Plint, 1987) and may provide an estimate for movement on the Highlands shear zone through the blocking temperatures of these minerals. To expand this database, three samples were collected across the Highlands shear zone (Fig. 8a), from the gneissic domain (sample 93193), to mylonite (sample 93194), to footwall pelitic schist (sample GL478), which yielded $^{40}Ar/^{39}Ar$ biotite ages of 373.1 ± 3.4 Ma, 374.1 ± 3.4 Ma, and 380.5 ± 3.5 Ma (Fig. 9), respectively. These dates may reflect the age of the shearing event since shear heating and subsequent shear cooling against low-grade footwall rocks is typical of large overthrust nappes (e.g. Ruppel and Hodges, 1994). An upper limit for deformation is given by the postkinematic high-level Salmon Pool pluton which has been dated at 365+10/-5 Ma U-Pb zircon (Jamieson et al., 1986).

Middle River area

In the Middle River area at the south-central end of the Cape Breton Highlands (Fig. 2), the sequence of diorite basement, clastic cover, and volcanic assemblage from the Early Silurian rocks of the Sarach Brook metamorphic suite, is partially repeated four times from north to south by an imbricate system of east-striking ductile faults. The imbricate stack has a structural thickness of approximately 3–4 km. Very fine-grained laminated light-coloured mylonite along the faults is as much as 100 m thick and contains a well developed north-plunging stretching lineation. Map-scale and outcrop-scale tight to close folds have overprinted the mylonite; the folds have upright north-striking axial planes with gently north-plunging fold axes. The mylonitic rocks are similar to the mylonite rocks developed during the early thrusting event in the Chéticamp area of western Cape Breton Island, and the

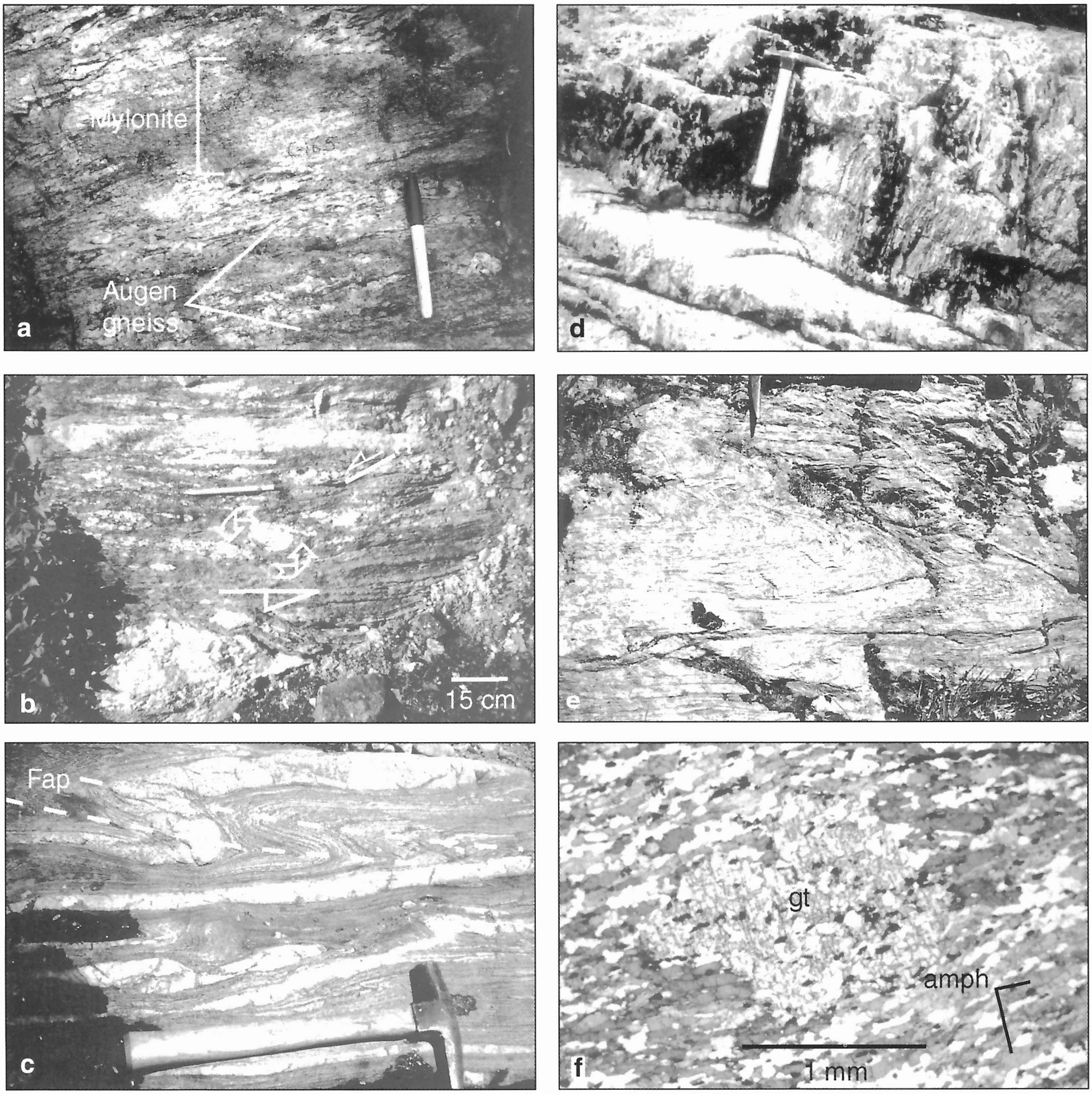


Figure 5. *Photographs of features from Highlands shear zone:* ***a)*** *mylonitic gneiss displaying heterogeneous distribution of shear with fabrics ranging from layers of fine-grained mylonite to augen mylonite containing coarse porphyroclasts of the gneissic protolith (Belle Côte Road, western Cape Breton Highlands)(photograph by G. Lynch; GSC 2000-041D);* ***b)*** *mylonitic augen gneiss containing coarse gneissic or pegmatitic porphyroclast with asymmetric mineral trails indicating noncoaxial rotational shear (shear sense indicated by thick arrows and vorticity by small arrows), and development of low-angle Riedel shears (labelled R), pencil for scale is 15 cm (Belle Côte Road, western Cape Breton Highlands)(photograph by G. Lynch; GSC 2000-041E);* ***c)*** *detached intrafolial folds in ductile shear zone, dashed lines are fold axial planes (Fap) (Belle Côte Road, western Cape Breton Highlands)(photograph by G. Lynch; GSC 2000-041F);* ***d)*** *laminated very fine-grained mylonite and ultramylonite overprinted by late upright postkinematic kink folds (Ingonish River)(photograph by G. Lynch; GSC 2000-041G);* ***e)*** *early tight to isoclinal ductile folds developed in amphibolite immediately adjacent to Highlands shear zone (photograph by G. Lynch; GSC 2000-041H);* ***f)*** *photomicrograph of high pressure and temperature garnet-amphibolite (gt) from Cabot nappe metamorphosed to*

style of deformation is also similar by virtue of the imbrication of the Ordovician–Silurian sedimentary- volcanic succession with its basement. There is little or no metamorphic contrast between the hanging wall and footwall of individual thrusts. The Southern Highlands shear zone in the Middle River area (Mengel et al., 1991) is the southern continuation of the Highlands shear zone described above. In making this correlation the Southern Highlands shear zone in the Middle River area is herein referred to as the Highlands shear zone, and here it bounds early imbricate thrusts and separates relatively low-grade rocks in the south from amphibolite-facies rocks and gneiss of the Middle River metamorphic suite (Doucet, 1983; Horne, 1995) to the north. Mylonite along the Highlands shear zone in this area displays a range of textures from high-grade ductile mylonitic gneiss and augen gneiss to laminated very fine-grained low-grade mylonite and ultramylonite. In the high-grade assemblages, hornblende defines the principal lineation, and coarse-grained biotite and muscovite define well developed C-S fabrics between feldspar porphyroclasts. Lineations plunge from north to northwest (Fig. 4), and shear-sense indicators consistently display tops down to the north-northwest sense of shear with a strong sinistral component; however, here the Highlands shear zone is situated within the hinge zone of a north-plunging syncline and mylonite strikes from north-northeast to east along a strike length of several kilometres (Fig. 4).

Staurolite is confined to a relatively narrow zone of pelitic rocks along the Highlands shear zone in the Middle River area. Microscopically, staurolite porphyroblasts overgrow the principal schistose fabric suggesting growth during or after the formation of the matrix foliation. Staurolite also overgrows garnet, which contains inclusion trails that are oblique to those in the staurolite and surrounding matrix,

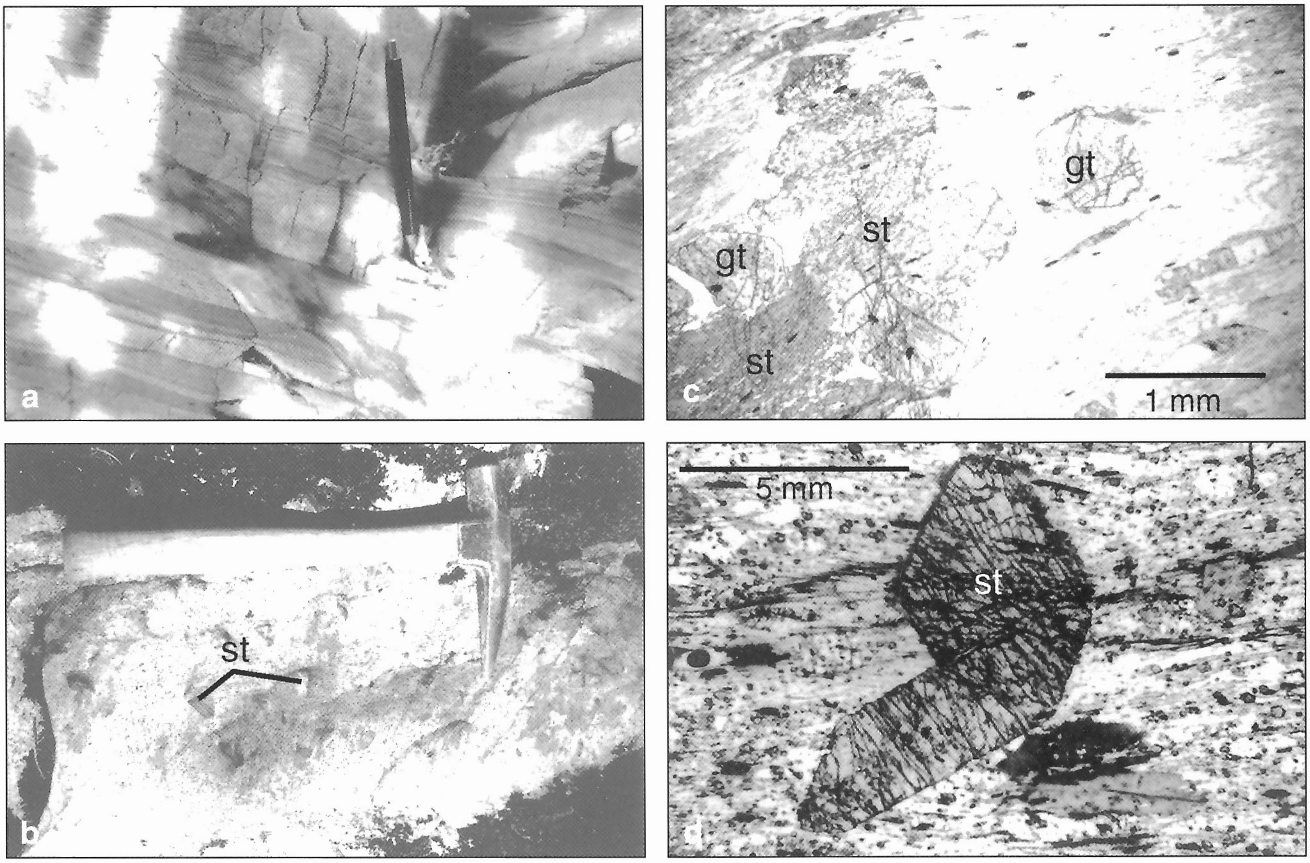


*Figure 6. Photographs showing features from staurolite halo surrounding Cabot nappe: **a)** thinly bedded greenschist-grade quartz-rich wacke with well developed graded bedding from Jumping Brook metamorphic suite is protolith to staurolite schist adjacent to Cabot nappe seen in b (photograph by G. Lynch; GSC 2000-041I); **b)** staurolite-garnet schist near headwaters of Corney Brook adjacent to Cabot nappe and Highlands shear zone, displaying typical coarse porphyroblasts of staurolite (st)(photograph by G. Lynch; GSC 2000-041J); **c)** photomicrograph of staurolite-garnet schist demonstrating early finer grained synkinematic garnets (gt) containing rotated fabrics (inclusions) overgrown by coarser grained postkinematic staurolite (st) containing nonrotated fabric (inclusions) suggesting growth during passive heating after overthrusting of Cabot nappe; **d)** euhedral staurolite (st) porphyroblasts formed after development of principal schistosity related to early thrusting event which predates emplacement of Cabot nappe and growth of staurolite crystals along margins of nappe.*

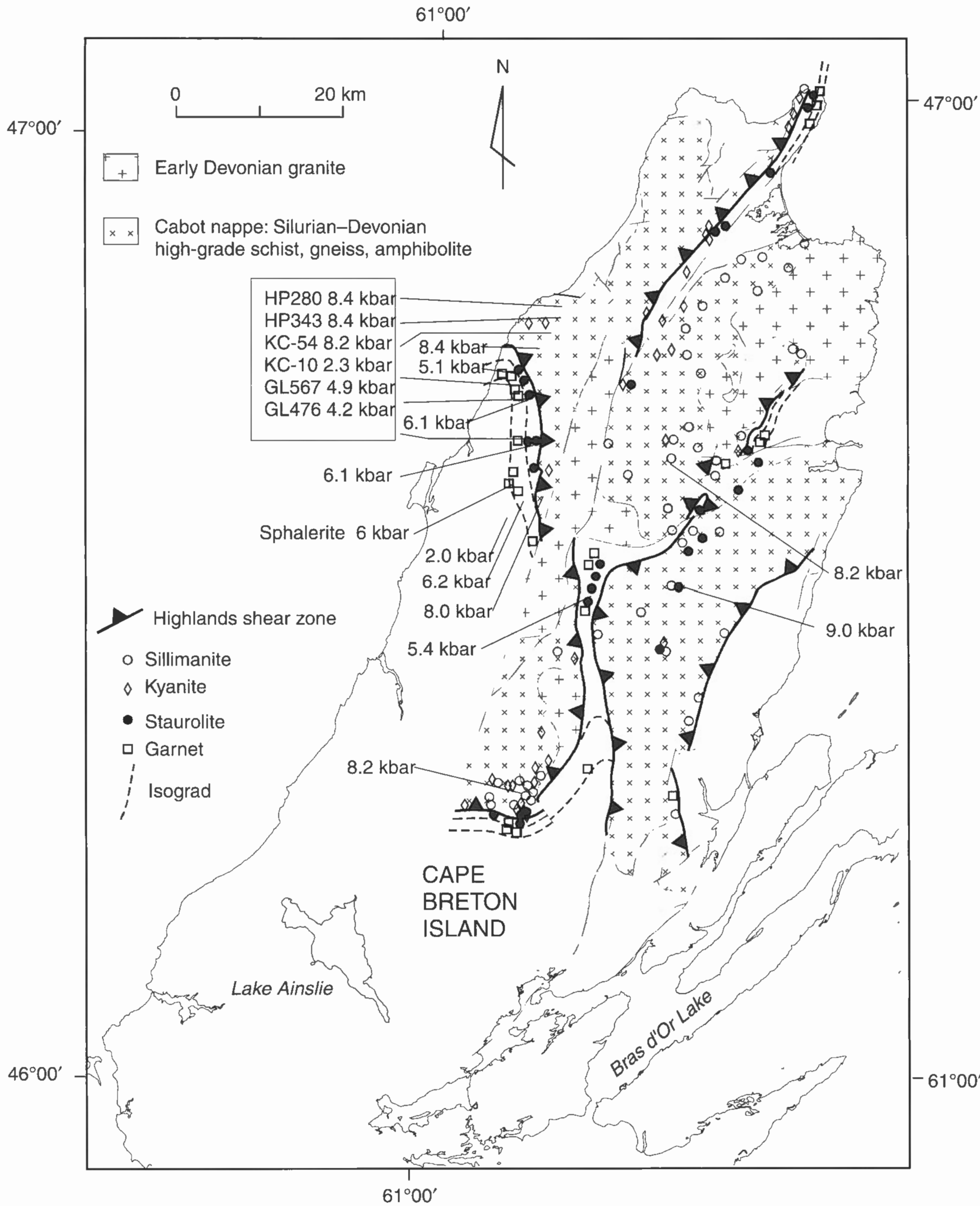


***Figure 7.** Isograd map including quantitative pressure and temperature determinations from Cape Breton Highlands compiled from Currie and Lynch (1997). New determinations highlighted with box, from mineralogical data presented in Tables 1 and 2 for samples GL476 and GL567, whereas mineralogical data for KC and HP samples are from Currie (1987) and Plint (1987), respectively. Paleopressure data is given in kilobars (kbar).* (modified from *Lynch (1996a), and Currie and Lynch (1997)).*

indicating the existence of an older fabric. Garnet porphyroblasts with sigmoidal inclusion trails typically have inclusion-free rims demonstrating a distinct two-stage growth (*see* descriptions by Doucet (1983) and Horne (1995)). With regards to the above descriptions, the microtextures for garnet and staurolite within and adjacent to the Highlands shear zone in the Middle River area are remarkably similar to those for garnet and staurolite in the Chéticamp area in footwall units immediately adjacent to the Highlands shear zone.

The $^{40}Ar/^{39}Ar$ age determinations from high-grade rocks of the Middle River metamorphic suite in the hanging wall of the Highlands shear zone have produced ages of 390 Ma and 388 Ma for hornblende and 370 Ma for biotite (Reynolds et al., 1989), as well as 386 Ma and 377 Ma for hornblende (Doucet, 1983). Since the $^{40}Ar/^{39}Ar$ age determinations represent cooling through the closing temperatures of hornblende or biotite and since the Highlands shear zone is responsible for the juxtaposition of high-grade rocks with cooler greenschist-grade rocks, then the age determinations likely correspond to a stage of movement on the Highlands shear zone within the interval 390–377 Ma. A further constraint on the age of the Highlands shear zone in the Middle River area is provided by Horne (1995), who describes the timing relationships between the 376 Ma old synkinematic Botham Brook pluton and the Highlands shear zone.

Mabou area

The Mabou Highlands consist of a large isolated horst block exposing basement units through the Carboniferous cover in southwestern Cape Breton Island (Barr and Macdonald, 1989). In this horst block (Fig. 3) abundant and thick steeply dipping mylonite zones occur. These include a significant volume of mylonitized plutonic and gneissic rocks, as well as metavolcanic and metasedimentary units of likely Neoproterozoic age (Lin et al., 1997). Gneissic rocks include amphibolite, granitic orthogneiss ranging in composition from granite to diorite, a migmatitic injection gneiss complex (Barr and Macdonald, 1989), and minor marble and possibly quartzite. The predominant structural feature within the Mabou Highlands is the imbrication of the volcano-sedimentary package in a style similar to that described in the previous sections from the Chéticamp and Middle River areas for the early thrusting event.

The shear zones define moderately to steeply dipping faults which strike north to northwest, outlining a northward-tapering imbricate stack bounded to the west by a relatively straight, north-striking mylonitic fault. Lineations plunge northward along mylonite zones. Shear-sense indicators (winged inclusions and C-S fabrics) indicate combined dextral-reverse motion transporting blocks upward and toward the south. Small to outcrop-scale polyclinal, disharmonic intrafolial box folds within laminated mylonite occur between planar strands of mylonite within shear zones. Some folds, however, are apparently not bound by planar shear surfaces and may be due to a later stage of folding overprinting the shear zones. Larger scale, folded mylonitic shear zones are detached along and were apparently transported by adjacent mylonitic faults.

Southeastern Cape Breton Highlands

In the southeastern Cape Breton Highlands, in the region just north of St. Anns Harbour, the Barachois River augen gneiss (Barr et al., 1987b; Raeside and Barr, 1992) is a north- to northeast-trending high-grade shear zone which separates granitic rocks, low-grade lapilli tuff, and subvolcanic dacite and andesite of the Precambrian Price Point Formation to the east-southeast, from granodiorite, diorite, and tonalite as well as high-grade metasedimentary and gneissic units to the west-northwest. The shear zone is exposed along Mill Cove Brook, Quarry Brook, Goose Cove Brook, Neil MacLeods Brook, Timber Brook, Barachois River, and East Branch Indian Brook, as well as other areas, and can be traced from north to south for approximately 30 km. Units confined to the high-grade side of the fault include an extensive package of quartzite, marble, metawacke, and amphibolitic metabasalt which also includes the high-grade MacMillan Flowage Formation (Raeside and Barr, 1992), as well as deep-seated Late Proterozoic tonalite intrusions (Farrow and Barr, 1992), migmatitic tonalite, and the gneissic Late Proterozoic Kathy Road diorite (Dunning et al., 1990a). Most of these units have been affected to varying degrees by Silurian metamorphism (e.g. Reynolds et al., 1989; Dallmeyer and Keppie, 1993) (Fig. 8a). Two stages of folding are widely observed within the high grade rocks (Fig. 10) producing folded boudins (Fig. 10b) hook-shaped (type 3) and crescent-shaped (type 2) (Fig. 10d) fold interference patterns (Ramsay, 1967) at the outcrop scale; early recumbent tight to isoclinal folds (Fig. 10c) have been overprinted by a later stage of tight to close upright folds (Fig. 10a) with north-northeast-striking axial planes.

The shear zone is several hundred metres wide and dips moderately to steeply towards the west-northwest. Shearing affects granitic rocks adjacent to the shear zone, and fault rocks include potassium feldspar mylonitic gneiss and augen gneiss that have been sheared into mica-rich semipelitic gneiss and schist containing abundant augen. Lineated amphibolite also occurs. Finely laminated light-coloured mylonite and ultramylonite are found along the fault in areas where the shear zone truncates Precambrian quartzite and metawacke from hanging wall units. Stretching lineations and shear-sense indicators including winged inclusions, rotated porphyroclasts, shear bands, and C-S structures in mica-rich units, indicate east-side-up and west-side-down motion along the shear zone (Fig. 4). Small-scale intrafolial sheath folds are found in the gneissic rocks as well as within the ultramylonite. Small-scale moderately plunging open folds with upright north-striking axial planes locally overprint the primary shear fabrics. Metamorphic grade decreases

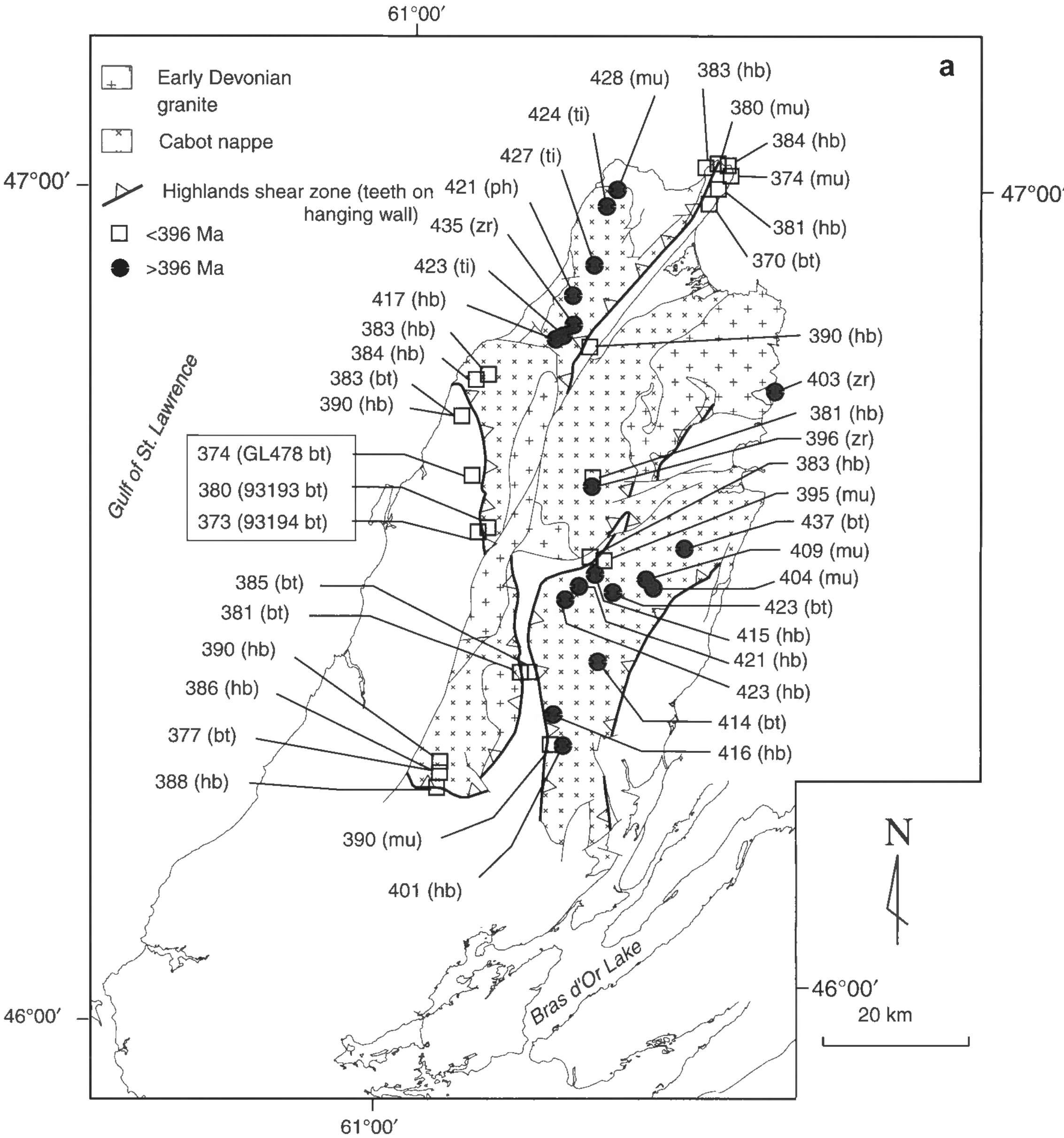


Figure 8. ***a)*** *Compilation of Silurian and Devonian* $^{40}Ar/^{39}Ar$ *cooling ages and some U-Pb age determinations from the Cabot nappe. Mineral abbreviations are as follows: hb, hornblende; bt, biotite; mu, muscovite; ph, phlogopite; ti, titanite; and zr, zircon. Data are from Jamieson et al. (1986), Reynolds et al. (1989), Dunning et al. (1990a), Keppie et al. (1992), Dallmeyer and Keppie (1993), and Miller et al. (1996). New biotite data from determinations presented in Figure 9 highlighted with a box* (modified from *Lynch (1996a)).* ***b)*** *Histogram compilation of Ar/Ar and K-Ar age determinations for hornblende and mica in rocks from the Cape Breton Highlands and western Newfoundland, demonstrating Silurian and Devonian thermal events during time of uplift and erosion for these regions. Diagram depicts the Gaspé and Chaleurs groups of Gaspésie as possible basins or sinks which may have captured detritus from uplifted and eroded areas. Original sources of geochronological data referenced in caption of Figure 8a, and in Cawood (1993). Stratigraphic column of Gaspésie* modified from *Malo and Bourque (1993), and incorporates interpretation of major transgressive (T I, T II) and regressive R I, R II, R III) cycles. Abbreviations: CBI, Cape Breton Island; W. Nfld, western Newfoundland; and Carbonife., Carboniferous.*

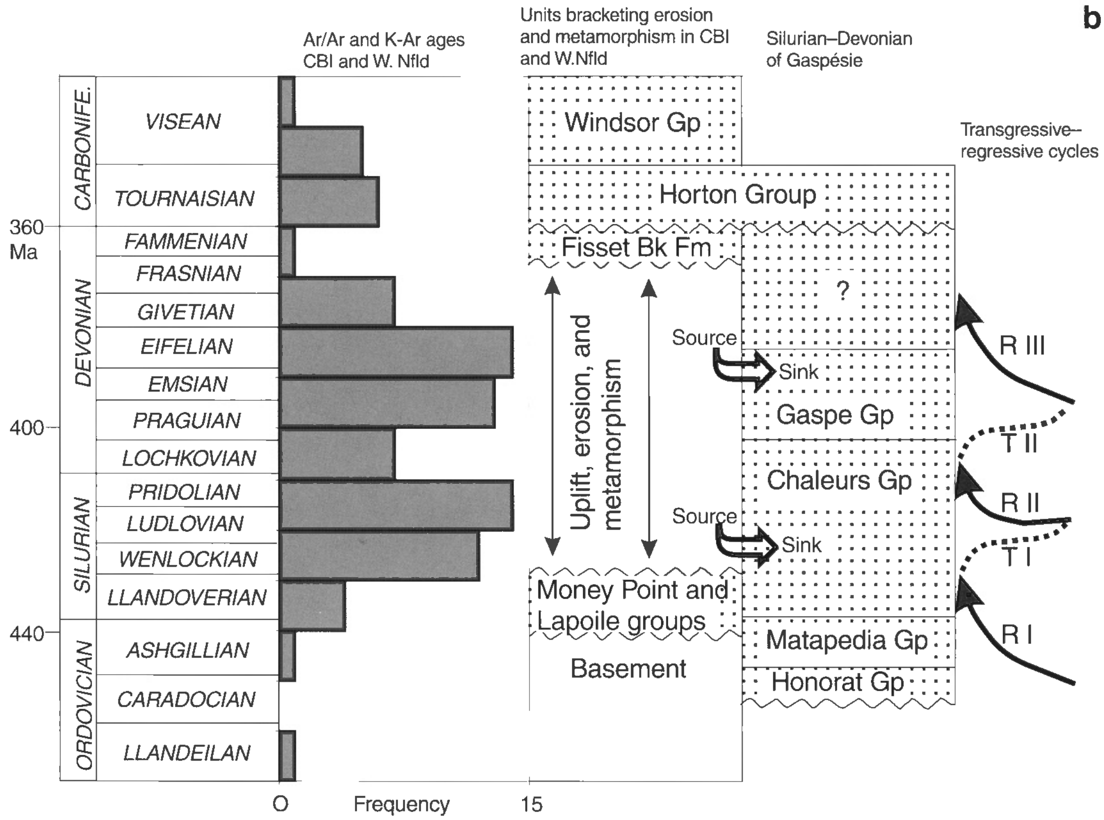


Figure 8b.

abruptly across the trace of the shear zone from relatively higher grade gneissic and schistose assemblages on the northwestern side of the fault to low-grade units toward the southeast, with the transition marked by the sillimanite, garnet, and biotite isograds in the area of Goose Cove Brook (Barr et al., 1987b).

Northeastern Cape Breton Island

In the northeastern Cape Breton Highlands near the headwaters of Clyburn Brook, the Eastern Highlands shear zone (Barr et al., 1987b; Lin, 1992) consists of a thick mylonitic zone approximately 500 m wide, which strikes northeast, and dips steeply to the northwest (Fig. 4). The shear zone juxtaposes high-grade gneiss and orthogneiss on its northwestern side, against low-grade volcanic and clastic rocks to the east which are correlated here with the Ordovician–Silurian Jumping Brook metamorphic suite. In the low-grade units black shale, siltstone, and wacke are widespread, and are interbedded with porphyritic andesite, tuff, chloritic basalt, and basaltic volcanic breccia. A pronounced metamorphic gradient is recorded within the volcanic-sedimentary succession, with the metamorphic grade increasing from east to west towards the shear zone, successively through garnet-biotite and staurolite isograds. Gneissic units west of the fault include granitic orthogneiss of the Chéticamp Lake gneissic complex dated at 396 ± 2 Ma by U-Pb zircon (Dunning et al., 1990a) as well as amphibolite, paragneiss, schist, and Early Ordovician gneissic conglomerate (Chen et al., 1995). Mylonite from the Eastern Highlands shear zone is overprinted by late folding producing noncylindrical tight upright folds, plunging moderately to the northeast or southwest, as well as abundant polyclinal box folds within thinly laminated mylonite (Fig. 5d). Shear-sense analysis of the Eastern Highlands shear zone by Lin (1992) demonstrates relative east-side-up and west-side-down vertical movement along the fault in its present position, however it is important to consider in a kinematic analysis that the shear zone is folded.

The northern extremity of the Eastern Highlands shear zone is truncated by the large Black Brook granite body dated at 373 ± 2 Ma, U-Pb monazite (Dunning et al., 1990a), which places an upper age limit for movement and shearing on the Eastern Highlands shear zone. The shear zone is truncated to the south by a series of steep brittle east-striking faults of likely Carboniferous age.

Northern Cape Breton Island

The Wilkie Brook fault (Fig. 4) strikes north-northeast and can be traced for approximately 40 km along strike from Cape North southwesterly to the Margaree pluton in the north-central Cape Breton Highlands. The fault zone features a thick section of intensely mylonitized rock, flanked on its eastern side by gneiss, amphibolite, mylonitic gneiss, marble, and schist of the Cape North Group (Macdonald and Smith, 1980). Most of these flanking units demonstrate shear fabrics of varying intensity, and include widespread protomylonite, in which case they may be included as part of the shear zone, though the limits are not well defined as the boundaries to the shear zone appear gradational across several kilometres. The fault forms the eastern boundary of the Blair River complex which extends all the way to the coast on the northwestern side of the island. As well as the Cape North Group, rocks east of the fault include Early Silurian volcanic and sedimentary rocks of the Money Point Group. The metamorphic grade increases progressively from east to west through the Money Point and Cape North groups toward the Wilkie Brook fault, through a number of well defined isograds (Macdonald and Smith, 1980). In the east, the Money Point Group is only weakly metamorphosed, and increases in metamorphic grade toward the west through the cordierite, staurolite, kyanite, and silimanite isograds. Metamorphic $^{40}Ar/^{39}Ar$ cooling ages for hornblende within the rocks east of the Wilkie Brook fault range from 390 Ma to 381 Ma (Keppie et al., 1992) (Fig. 8a).

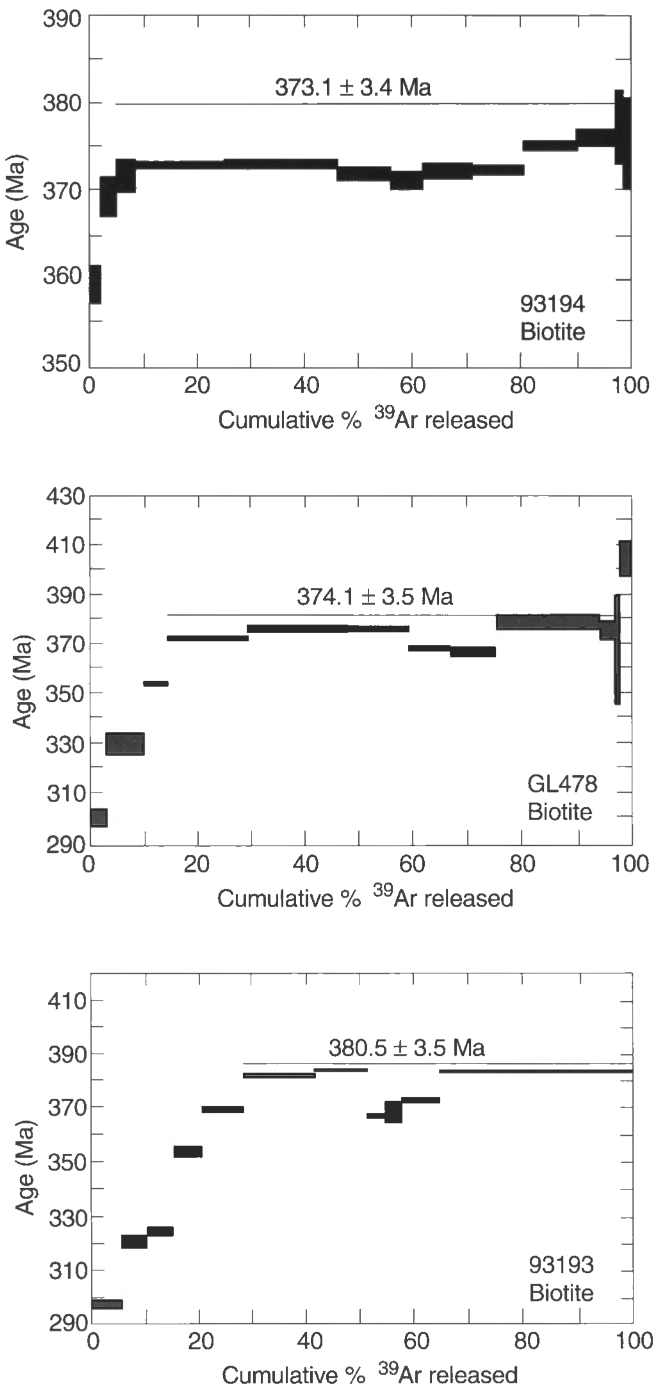


Figure 9. *Biotite $^{40}Ar/^{39}Ar$ ages for schistose samples from locations shown in Figure 8 near margin of Cabot nappe and Highlands shear zone. Analyses and age determinations done in geochronology laboratory of the Geological Survey of Canada, Ottawa.*

Five deformation phases (D_1–D_5) are described and discussed for the Money Point and Cape North groups by Macdonald and Smith (1980), of which D_1 and D_2 feature tight to isoclinal coaxial folding and a penetrative stretching lineation. D_3 deformation features upright northeast-trending kink and chevron folding, which also overprints the mylonite of the Wilkie Brook fault. Limited kinematic data are available for the mylonite of the Wilkie Brook fault. Observations made at a number of points along the fault indicate east-side-up and west-side-down sense of shear based on shear bands, C-S fabrics, and a steep to oblique downplunge stretching lineation. D_3 folds which overprint the mylonite indicate that the mylonite planes and their contained shear-sense indicators have likely been rotated from their original positions.

Outline of Cabot nappe and correlation of fault segments from the Highlands shear zone

In the Cape Breton Highlands a number of major shear zones have been identified by different authors which separate high-grade gneissic rocks from low-grade units and domains. Although these are major fault structures, as evidenced by very thick mylonite and strongly contrasting metamorphic domains across the faults, they are discontinuous structures which do not have a simple map trace at the surface. The difficulties in mapping these faults along strike can be partly attributed to truncation and offset along late brittle faults, to large crosscutting Late Devonian plutons, and to Carboniferous deposits which hide portions of the structures. Another reason for the irregular surface trace of these shear zones is because they have been folded. Hand sample and outcrop-scale folds, as well as map-scale folds which overprint the mylonite are observed in all areas. Although the shear zones are predominantly exposed on steep northeast-striking limbs of folds, map-scale, moderately north-plunging fold hinges containing major shear zones are exposed along Ingonish River (anticlinal), Middle River (synclinal), and near

Pleasant Bay (anticlinal). It is the actual shear zones which bound the gneissic bodies that provide a useful marker for tracing the map-scale folds. Furthermore, all of these shear zones share similar structural characteristics and bound similar rock units across Cape Breton Island; most areas studied contain correlative greenschist-facies, Ordovician–Silurian sedimentary and volcanic rocks on one side of the fault, and have been overprinted by the staurolite isograd next to the fault. The distribution of these greenschist units as well as the staurolite isograd are also useful in mapping large-scale folds. The nature of the rocks on the high-grade side of the shear zones is more irregular, but most are gneissic in texture and have been variably affected by amphibolite-facies metamorphism in the Late Silurian to Early Devonian. As such, an integrated view of the shear zones can be adopted, with the evidence indicating that the various shear zones considered in this section actually constitute a single folded shear zone on a large scale, which is referred to in this analysis as the Highlands shear zone. Specifically the Wilkie Brook fault, the Eastern Highlands shear zone, the unnamed shear zones bounding gneiss described from the St. Anns area, as well as the segments described in the western and southern Cape Breton Highlands can be correlated on structural, lithological, and metamorphic grounds as being the same fault. Correspondingly the gneissic domain can be viewed as a high-grade allochthonous sheouet bound by the Highlands shear zone, referred to as the Cabot nappe, and whose outline with the Highlands shear zone is shown in Figures 2, 4, 7, and 8.

A kinematic interpretation of the Highlands shear zone must take into account the effects of folding, and that the orientation of the mylonite has been altered. Metamorphic contrasts provide the most reliable means of assessing the relative displacements of the units across the Highlands shear zone. As a generalization, the deep-seated gneissic rocks affected by Silurian metamorphism have moved upward

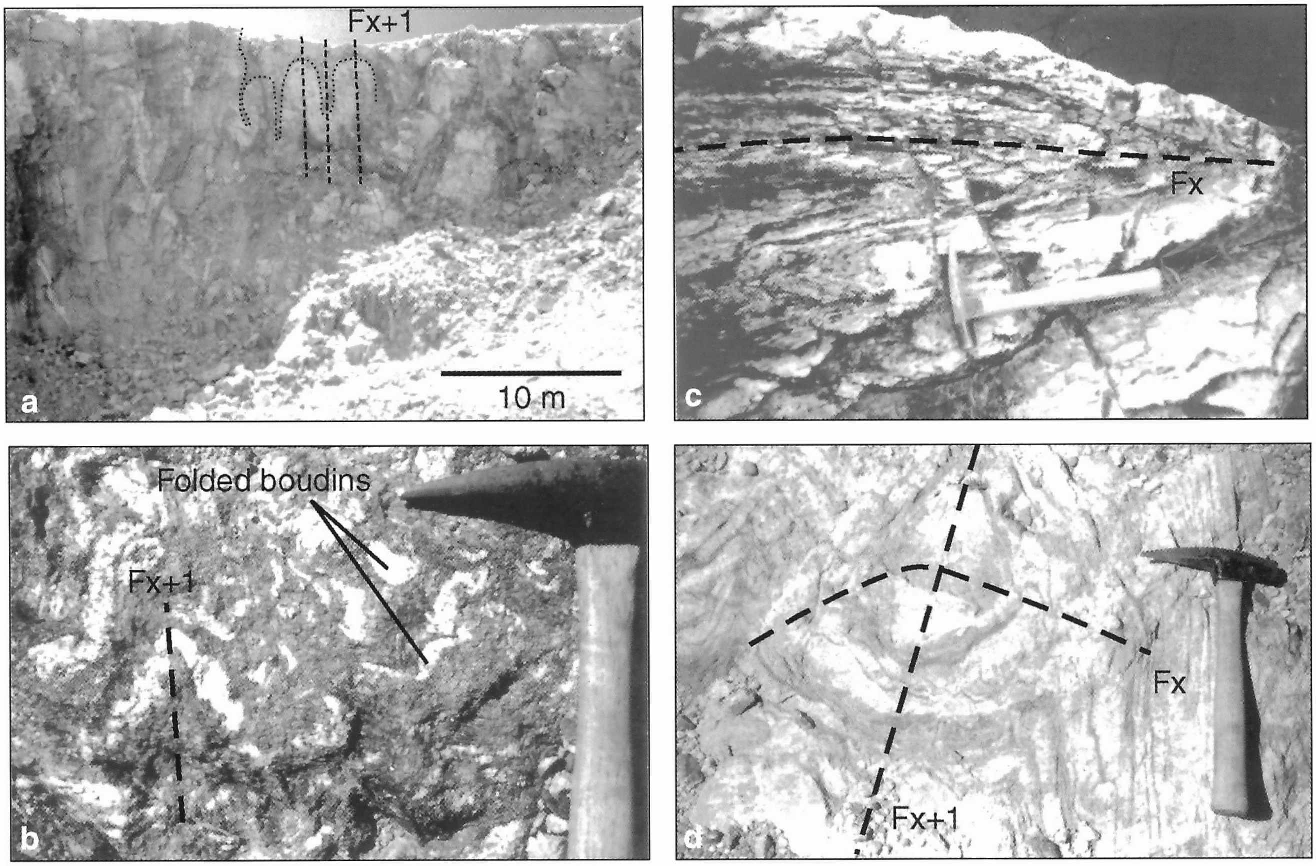


***Figure 10. a)** Large, tight, upright folds overprinting early gneissic fabric from rocks of the Cabot nappe exposed in quarry south of Gisborn Flowage, typical of late-stage folding overprinting the nappe. Quarry wall is approximately 8 m high, and axial planes of folds shown with dashed ligne (F_{x+1})(photograph by G. Lynch; GSC 2000-041K); **b)** folded quartzofeldspathic gneiss boudins from Cabot nappe also highlighting late-stage of folding (axial plane F_{x+1}) overprinting earlier gneissic fabric and stretching (photograph by G. Lynch; GSC 2000-041L); **c)** quartzite from southeastern lobe of the Cabot nappe showing early stage of recumbent isoclinal folding, likely developed during early nappe transport and overthrusting (photograph by G. Lynch; GSC 2000-041M); **d)** same quartzite unit showing two stages of folding and crescent-shaped fold interference pattern, demonstrating late folding event which overprints Cabot nappe (photograph by G. Lynch; GSC 2000-041N).*

relative to the greenschist-grade units. As such the gneissic domain is interpreted as occupying the hanging wall of the Highlands shear zone, and the greenschist-facies rocks including Ordovician–Silurian units are in the footwall. Other evidence of this relative motion is the thermal imprint that overthrusting of the gneissic body has had on the footwall units producing the regional staurolite isograd in the immediate footwall to the Highlands shear zone, reflecting an inverted metamorphic gradient. In considering the structural position of the Cabot nappe in the hanging wall of the Highlands shear zone, the folded nappe defines an isolated klippe in the core of a regional synclinorium. Outcrop-scale and mineral-scale kinematic data from mylonite around the margins of the klippe suggest that, prior to folding, the Cabot nappe was thrust in a general northwesterly direction.

Although presently isolated as a partial klippe, the Highlands shear zone was a deeply rooted fault which is characterized by the juxtaposition of highly contrasting metamorphic domains, with the thrust emplacement of the main gneissic nappe resulting in the regional development of inverted isograds. Four broad lithological subdivisions within the Cabot nappe include: from north to south 1) the Blair River complex of Grenvillian derivation (Barr et al., 1987a; Barr and Raeside, 1989) (unit 1 on Fig. 2); 2) the Pleasant Bay complex of Currie (1987), including the Chéticamp Lake gneiss and Ordovician–Silurian gneissic conglomerate (Chen et al., 1995) (unit 4 on Fig. 2); as well as the southeastern lobe of the Cabot nappe containing 3) Proterozoic quartzite, metawacke, marble, and amphibolite (unit 2 on Fig. 2), intruded by 4) large Neoproterozoic granite, granodiorite, diorite, and tonalite bodies (unit 3 on Fig. 2). Although the nappe has preserved mica and amphibole cooling ages from older igneous and metamorphic events (Reynolds et al., 1989; Dallmeyer and Keppie, 1993), both the Silurian and Devonian metamorphic ages that characterize the nappe are recorded in all portions of the thrust sheet (Fig. 8a).

In the north the cooling temperatures within the Blair River complex are slightly older than other parts of the Cabot nappe, recording Silurian denudation (ca. 427–421 Ma for titanite, and 417 Ma for hornblende, Fig. 3); however cooling ages in the proximity of the Highlands shear zone along the eastern margin of the Blair River complex record Early Devonian cooling periods (ca. 390–381 Ma $^{40}Ar/^{39}Ar$ for amphibole, Keppie et al., 1992). Similar diachronous patterns are recorded elsewhere within the Cabot nappe (Fig. 8a). Evidently periods of peak metamorphism and partial denudation occurred in the Late Silurian, however geochronological data, particularly that which is distributed along the margins of the Cabot nappe in proximity of the Highlands shear zone, or in areas intruded by Devonian plutons in the central portion of the Cabot nappe, demonstrate that the thermal effects related to denudation persisted into the Early Devonian. Shear heating has been proposed as an effective means of causing metamorphism and anomalous heat levels adjacent to major shear zones (Molnar and England, 1990; England et al., 1993; Ruppel and Hodges, 1994). Such a mechanism is considered here to have been a possible factor in generating Early Devonian metamorphism along the Highlands shear zone, even though portions of the core of the thrusted nappe began to cool in the Silurian.

Metamorphism and pressure-temperature conditions

An understanding of metamorphism is central to comprehending the geology of the Cape Breton Highlands because of the widespread distribution of metamorphic rocks in this region. Gneiss and associated high-grade schist define the Cabot nappe, and place it in sharp contrast with surrounding units. The transition from high-grade to low-grade rocks is typically abrupt and occurs across mylonite of the Highlands shear zone which bound the nappe. This section investigates in more detail pressure, temperature, and geochronological aspects of the high-grade Cabot nappe and adjacent lower grade units.

High-grade metamorphism in the Cabot nappe

In the central Cape Breton Highlands, high-grade metamorphic units affected by Silurian–Devonian metamorphism include rocks of the Pleasant Bay Complex in the west (Currie, 1987; Plint and Jamieson, 1989), the Chéticamp Lake gneissic complex in the east (Barr and Jamieson, 1991; Raeside and Barr, 1992), and the Middle River metamorphic suite in the south (Doucet, 1983). To the north, the Blair River complex was affected by high-grade metamorphism in the Silurian (Miller et al., 1996), as were units within the southeastern lobe of the Cabot nappe (Reynolds et al., 1989; Dallmeyer and Keppie, 1993). Rocks in the Pleasant Bay Complex are dominated by tonalitic to granodioritic orthogneiss, amphibolite, pelitic gneiss, quartz-rich calcareous gneiss, highly foliated granitic rocks and variably foliated concordant to discordant pegmatite dykes. Orthogneiss has been dated at 433 +20/-10 Ma by the U-Pb zircon method, and has been interpreted as a magmatic age (Jamieson et al., 1986), whereas a date of 411 ± 2 by the U-Pb monazite method from a rock of the same suite has been interpreted as an age likely reflecting the peak conditions of metamorphism (Barr and Jamieson, 1991).

Rocks of the Chéticamp Lake gneiss are dominated by monzogranitic to granodioritic orthogneiss, with lesser semipelitic gneiss, biotite schist, amphibolite, and calcareous gneiss. Dating has yielded an age of 396 ± 2 Ma by the U-Pb zircon method, which has been interpreted as the age of high-grade metamorphism (Dunning et al., 1990a). The Middle River metamorphic suite (Doucet, 1983) on the other hand, is characterized by a greater abundance of amphibolite, pelitic to semipelitic schist, as well as distinct marble layers. In the southeastern Cape Breton Highlands, the Cabot nappe is characterized by an extensive north-trending unit of thick quartzite and marble, with adjacent units of psammitic schist and chlotitic schist (metawacke), and includes paragneiss and marble of the MacMillan Flowage Formation (Raeside and Barr, 1992). This segment of the nappe also features large bodies of dioritic orthogneiss and variably foliated tonalite and granodiorite.

An extensive geochronological database exists for high-grade rocks of the Cape Breton Highlands (e.g. Fig. 8). The $^{40}Ar/^{39}Ar$ cooling ages for high-grade rocks within the Cabot nappe span a 40 Ma interval from 423 Ma to 381 Ma (Fig. 8a); however, the cooling ages are unevenly distributed, with younger ages typically found in the central portion of the nappe where large late orogenic to postorogenic Early Devonian plutons have intruded (Fig. 8a), likely affecting the local heat budget and cooling ages of amphibole.

Several publications describe the petrographic relationships of metamorphism within high-grade rocks in the Cape Breton Highlands (e.g. Macdonald and Smith, 1980; Doucet, 1983; Craw, 1984; Currie, 1987; Plint and Jamieson, 1989; Farrow and Barr, 1992; Raeside and Barr, 1992). Until recently, very little quantitative data have been available on the specific pressure and temperature conditions of Silurian–Devonian metamorphism for high-grade rocks in the Cape Breton Highlands. Using garnet-biotite-plagioclase thermobarometry on mineral analyses from two samples, Plint and Jamieson (1989) have determined that peak metamorphism for high-grade rocks in the western Cape Breton Highlands reached conditions of approximately 700–750°C and 8–10 kbar, and it was pointed out that such conditions are necessary for the coexistence of potassium feldspar and kyanite as observed (Plint and Jamieson, 1989; Barr and Jamieson, 1991; Raeside and Barr, 1992). In the Middle River area Doucet (1983) bracketed metamorphic temperatures for the high-grade rocks to approximately 514–634°C using the garnet-biotite thermometer, and 415–612°C using the calcite-dolomite thermometer.

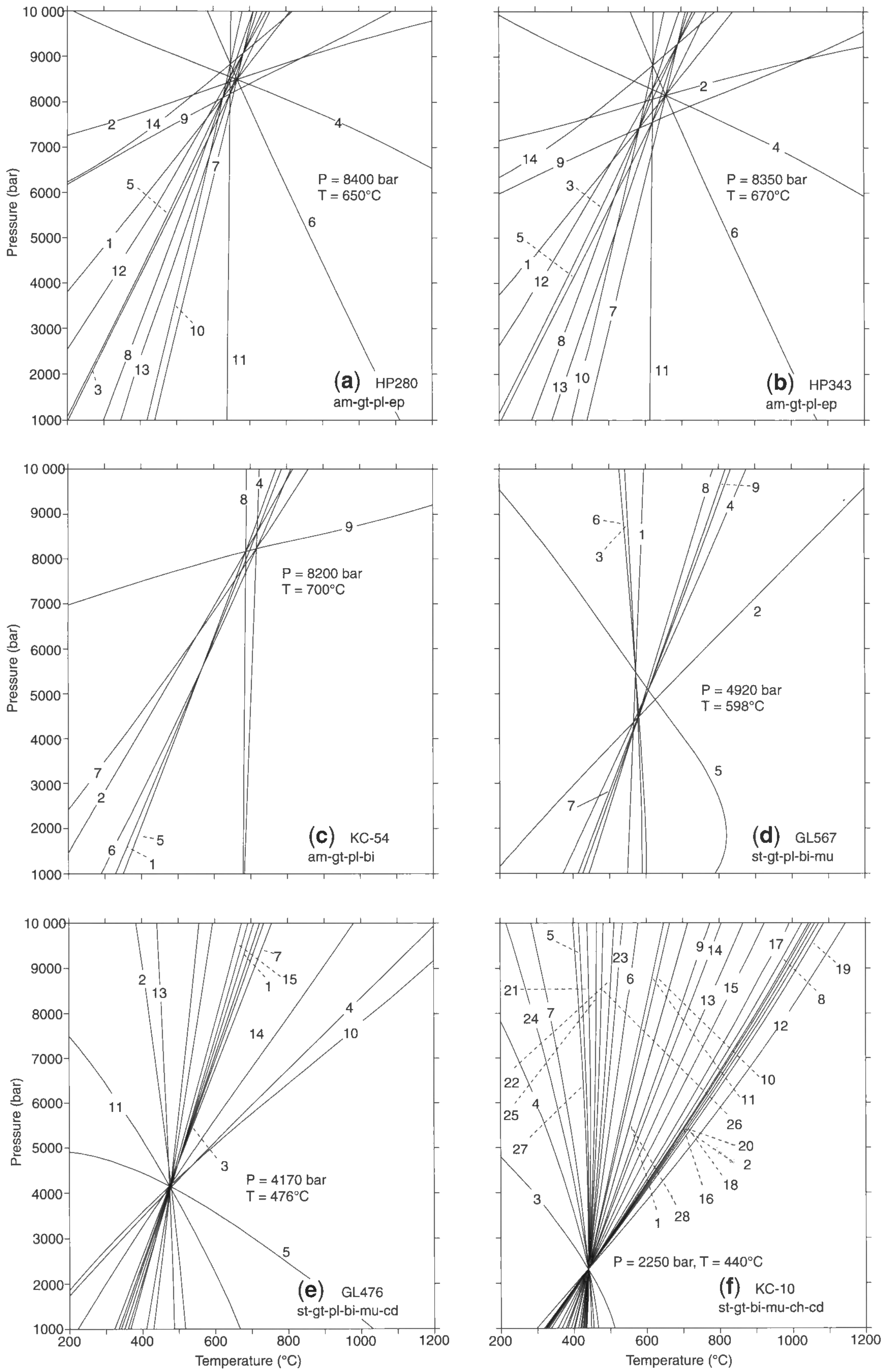
To augment the pressure-temperature database for high-grade rocks in the Cape Breton Highlands Currie and Lynch (1997) sampled and analyzed garnet-amphibolites (Fig. 5e, 5f) from the Pleasant Bay Complex and the Belle Côte Road gneissic complex, the Chéticamp Lake gneiss, the Middle River metamorphic suite, as well as paragneiss from the southeastern Cabot nappe (Fig. 7). Assemblages which include coexisting garnet and amphibole were carefully chosen because garnet-amphibolite grade rocks have proven to be reliable thermobarometers for high-grade rocks (e.g. Kohn and Spear, 1989, 1990; Mader and Berman, 1992). Similar quartz-rich calcareous gneissic rocks were sampled from six different sites. Quantitative mineralogical data from three garnet-amphibolite samples of Plint (1987) and Currie (1987) were also processed for pressure-temperature determinations, using methods described in Berman (1991) and Mader and Berman (1992). The two key assemblages for thermobarometry are 1) amphibole-epidote-garnet-plagioclase-quartz±biotite, and 2) amphibole-diopside-garnet-plagioclase-quartz-biotite. Biotite was not recorded in the two garnet-amphibolite samples from Plint (1987). Also, in the southeastern Cape Breton Highlands garnet has not been found within amphibolite or quartz-rich calcic paragneiss, although the coexistence of rutile and titanite within the key assemblage of amphibole-epidote- plagioclase-quartz- titanite-rutile provides a means for accurate pressure-temperature determinations as well (Currie and Lynch, 1997). Typically amphibole shows a strong lineation and occurs with abundant quartz and variable proportions of epidote, plagioclase, and biotite. The amphibole composition is that of a calcic hornblende and exhibits blue-green pleochroism in thin section. Garnet is typically small and usually inclusion-free, and is preferentially concentrated along individual laminae. Diopside, where present, forms symplectic intergrowths with epidote. Calcite and titanite are accessory minerals within diopside-bearing samples.

Pressure-temperature determinations based on the intersection point of three independent reactions (e.g. Berman, 1991) for garnet-amphibolite consistently fall within the interval 8–8.5 kbar and 675–740°C (Fig. 7) (Currie and Lynch, 1997), in areas where rocks demonstrate Late Silurian to Early Devonian $^{40}Ar/^{39}Ar$ amphibole plateau ages (Fig. 8). The biotite-free garnet-amphibolite samples of Plint (1987) recorded slightly cooler temperatures, at approximately 605°C, but a similar high pressure of 8 kbar (Fig. 7, 11). For the southeastern Cabot nappe, a high-quality pressure-temperature determination has been made on the garnet-free calcic paragneiss based on the intersection point of three independent reactions in the key assemblage amphibole-epidote-plagioclase-quartz-titanite-rutile, yielding a value of 9.0 kbar and 776°C, in an area where rocks preserve $^{40}Ar/^{39}Ar$ amphibole plateau ages of 423–415 Ma (Fig. 7). These results effectively place the Cabot nappe within the upper amphibolite facies of metamorphism (Fig. 12), and demonstrate that peak conditions and uplift occurred from Late Silurian to Early Devonian.

Metamorphism adjacent to the Cabot nappe

Pressure-temperature determinations were compiled and made using previously published and new (Table 1) mineralogical data, for rocks immediately adjacent to the Cabot nappe and Highlands shear zone in order to compare and contrast the crustal levels for metamorphism within and outside of the nappe (Fig. 11, 12). Early Silurian supracrustal volcanic-sedimentary rocks outside or adjacent to the Cabot nappe are typically metamorphosed to greenschist grade. Mafic volcanic rocks contain abundant chlorite, epidote, plagioclase, actinolite, and calcite, as well as chloritoid locally and rare hornblende. Clastic sedimentary rocks and phyllite contain predominantly metamorphic muscovite as well as chlorite, and rare biotite, garnet, and kyanite; however, as described in previous sections the metamorphic grade increases in these rocks towards the Cabot nappe and Highlands shear zone. Marking this transition are various isograds of which the staurolite isograd is the most conspicuous and mappable on a regional scale. The staurolite isograd parallels the Highlands shear zone on both the eastern and western sides of the Cabot nappe, occurring in pelitic units within less than 1–2 km of the fault (Fig. 7). Staurolite, up to 3 cm, typically overgrows an earlier schistose fabric giving the rock a distinct knobby texture.

Staurolite in pelitic rocks is quite temperature sensitive, with its stability field restricted to the interval 510–675°C (Fig. 12), straddling the upper greenschist to lower amphibolite transition. Equilibrium assemblages of garnet-biotite-muscovite-quartz-plagioclase-staurolite provide a useful thermobarometer. Consequently pressure and temperature



P = 8400 bar
T = 650°C
(a) HP280
am-gt-pl-ep
P = 8350 bar
T = 670°C
(b) HP343
am-gt-pl-ep
P = 8200 bar
T = 700°C
(c) KC-54
am-gt-pl-bi
P = 4920 bar
T = 598°C
(d) GL567
st-gt-pl-bi-mu
P = 4170 bar
T = 476°C
(e) GL476
st-gt-pl-bi-mu-cd
P = 2250 bar, T = 440°C
(f) KC-10
st-gt-bi-mu-ch-cd
Pressure (bar)
Temperature (°C)

(Opposite)

Figure 11. *Pressure-temperature determinations using TWEEQ (Berman, 1991), for mineral analyses contained in Tables 1 and 2 (samples GL476 and GL567), and from Currie (1987) (samples KC-10 and KC-54) and Plint (1987) (samples HP280 and HP343).* ***a), b),*** *and* ***c)*** *are high pressure-temperature determinations from garnet-amphibolite of Cabot nappe (Fig. 7);* ***d), e),*** *and* ***f)*** *are from the staurolite halo and include two cordierite-bearing samples. Mineral abbreviations are as follows: am, amphibole; gt, garnet; pl, plagioclase; ep, epidote; bi, biotite; st, staurolite; mu, muscovite; cd, cordierite; ch, chlorite; a full listing of reactions is provided in the Appendix.*

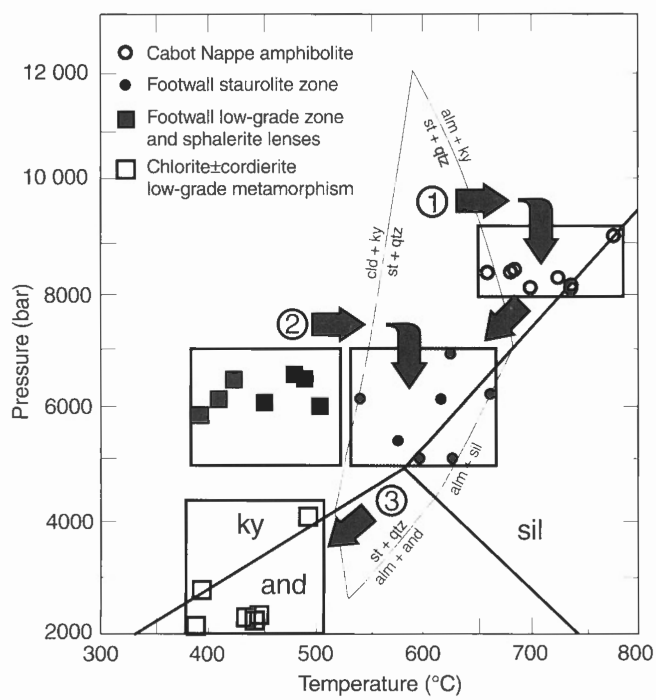


Figure 12. *Compilation of pressure-temperature determinations from Cabot nappe and footwall compiled from this study, Lynch and Mengel (1995), and Currie and Lynch (1997). Separate burial denudation paths for Cabot nappe and footwall are shown by trends 1 and 2, respectively, and juxtaposition occurring where two trends merge during Early Devonian overthrusting and transport of nappe. Final exhumation of nappe and footwall panel together during late-stage extension shown by trend 3. Mineral abbreviations are as follows as follows: alm, almandine; and, andalusite; cld, chloritoid; ky, kyanite; qtz, quartz; sil, sillimanite; st, staurolite.*

determinations were made on staurolite samples from along the western and southern margins of the Cabot nappe (Fig. 6, 11). Pelitic and psammitic schist containing the key assemblages of garnet-biotite-muscovite-quartz- plagioclase-sillimanite and/or kyanite or garnet-biotite-muscovite-quartz-plagioclase-staurolite indicate metamorphic conditions of 4.9–6.2 kbar and 536–661°C (Fig. 7, 11) (Currie and Lynch, 1997). These thermometry results compare well with those of Plint and Jamieson (1989) for the staurolite zone in western Cape Breton Island. Plint and Jamieson (1989) established thermometric values of 550–574°C for matrix biotite-garnet pairs according to the calibrations of Ferry and Spear (1978), and 600–640°C for biotite inclusions and garnet cores; however, the barometric determinations of Plint and Jamieson (1989) for these same rock units are somewhat higher; using garnet-plagioclase-kyanite-quartz, values of 5.9–7.5 kbar were determined on the one hand with the calibration of Newton and Haselton (1981), whereas 6.9–7.9 kbar conditions were determined using the calibration of Koziol and Newton (1988). Note, that as well as staurolite, cordierite defines a mappable isograd within the Money Point Group in the Cape North area at the northern tip of Cape Breton Island (Macdonald and Smith, 1980), adjacent to the Cabot nappe. Cordierite also occurs in the Jumping Brook metamorphic suite in western Cape Breton Island (Currie, 1987). Analytical data on cordierite, garnet, plagioclase, biotite, and muscovite for sample GL476 (Table 2), as well as for sample KC-10 from Currie (1987), were processed using the method described in Berman (1991), and yielded pressure-temperature estimates of 4.2 kbar and 476°C, and 2.3 kbar and 440°C, respectively (Fig. 11), indicating that cordierite growth represents lower grades of metamorphism than recorded for staurolite samples that do not contain cordierite.

Greenschist metamorphism of sulphide lenses in western Cape Breton Island

Below the staurolite isograd, metamorphic conditions within greenschist-grade rocks can be difficult to determine because minerals appropriate for thermobarometry are not always available; however, polymetallic sulphide lenses containing sphalerite-arsenopyrite-pyrrhotite-pyrite intergrowths occur in clastic and volcanic rocks of the Jumping Brook metamorphic suite in the western Cape Breton Highlands (Fig. 7), providing a mineral assemblage appropriate for pressure-temperature determinations in the low-grade belt (e.g. Lynch and Mengel, 1995). The sulphide minerals occur as 5–30 cm thick, highly sheared and recrystallized lenses concordant to bedding. Peak conditions of metamorphism within the low-grade belt of the Jumping Brook metamorphic suite, which contains the sulphide lenses, reached the upper greenschist facies. Metamorphic silicate assemblages include varying proportions of chlorite, muscovite, biotite, garnet, quartz, and albite, as well as locally chloritoid, amphibole, and rare kyanite. Plint and Jamieson (1989) estimated metamorphic temperatures to have been near 400–450°C in the low-grade belt beyond the staurolite isograd, on the basis of garnet-biotite geothermometry.

Table 1. Electron microprobe analyses of sample GL567.

	Biotite					Muscovite			Staurolite		Garnet			Plagioclase						
	1	2	3	4	5	1	2	3	1	2	1	2	3	1	2	3	4	5	6	7
SiO_2	33.94	33.44	35.00	35.05	34.85	46.27	46.41	46.48	26.59	27.65	36.23	36.55	35.17	63.13	63.87	63.86	63.98	64.49	63.85	62.42
TiO_2	1.30	1.51	1.74	1.88	1.62	0.68	0.39	0.79	0.76	0.78	0.00	0.11	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al_2O_3	17.85	17.85	18.35	19.20	17.58	35.45	34.20	34.96	52.72	53.07	20.55	20.65	20.13	22.73	22.51	22.22	22.51	21.20	22.46	22.60
FeO	23.09	24.57	19.90	19.39	20.91	1.36	1.44	1.46	14.15	14.12	35.65	36.27	36.63	0.08	0.02	0.08	0.04	0.00	0.00	0.19
MnO	0.00	0.03	0.00	0.00	.039	0.00	0.00	0.11	0.00	0.00	2.93	2.38	1.42	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	9.18	8.84	9.82	9.83	9.61	0.67	1.05	0.72	1.73	1.79	2.82	2.90	3.25	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	0.10	0.02	0.09	0.03	0.08	0.04	0.06	0.06	0.00	0.04	1.30	1.18	1.13	4.60	4.09	4.09	4.19	3.10	3.92	4.46
Na_2O	0.01	0.08	0.31	0.34	0.23	1.55	1.24	1.41	0.00	0.00	0.00	0.00	0.00	9.34	9.57	9.58	9.65	10.19	9.60	9.36
K_2O	8.25	7.42	8.77	9.10	8.41	9.13	9.22	9.10	0.02	0.02	0.00	0.00	0.00	0.07	0.09	0.11	0.05	0.08	0.09	0.06
Total	94.48	94.27	94.82	95.47	94.02	95.91	94.85	95.77	96.39	97.76	100.11	100.11	97.78	100.28	100.29	100.29	100.83	99.16	99.91	99.09

Table 2. Electron microprobe analyses of sample GL476.

	Biotite			Muscovite			Plagioclase			Garnet						Cordierite
	1	2	3	1	2	3	1	2	3	1	2	3	4	5	6	1
SiO_2	37.23	37.29	36.68	44.15	45.73	44.29	59.87	59.84	59.38	38.38	38.05	37.69	37.13	37.53	38.08	46.69
TiO_2	1.53	1.60	1.53	0.31	0.35	0.42	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Al_2O_3	19.15	19.38	19.29	33.48	33.39	32.83	25.92	25.74	25.72	20.88	20.96	19.96	19.78	19.79	19.73	33.01
FeO	14.30	14.54	14.21	1.98	1.86	1.91	0.02	0.05	0.05	35.18	34.81	36.41	34.68	35.33	35.88	12.10
MnO	0.14	0.14	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.10	0.16	0.23	0.12	0.16	0.18	1.10
MgO	12.63	12.73	12.78	0.74	0.73	0.69	0.06	0.06	0.05	2.17	2.57	1.90	2.36	2.39	1.87	5.50
CaO	0.00	0.00	0.00	0.00	0.00	0.00	6.80	6.59	6.84	4.18	3.46	3.46	3.77	4.03	3.39	0.01
Na_2O	0.40	0.40	0.36	1.41	1.54	1.34	7.27	7.28	7.30	0.00	0.00	0.00	0.00	0.00	0.00	0.50
K_2O	8.76	8.96	8.84	8.62	8.90	8.70	0.06	0.10	0.06	0.00	0.00	0.00	0.00	0.00	0.00	0.02
Total	94.14	95.04	93.81	90.69	92.50	90.18	100.00	99.66	99.40	100.89	100.01	99.65	97.84	99.23	99.13	98.94

The sulphide deposits are located near the sheared shallow-dipping contact between the lower mafic volcanic member and the upper clastic succession of the Jumping Brook metamorphic suite. A highly sheared rhyolite horizon (dated at 432 Ma; Lin et al., 1997) within the clastic succession hosts six of the occurrences; two others are contained within mafic volcanic rocks, and one within quartz-pebble conglomerate. The rhyolite, which is between 2 m and 20 m thick, is highly sericitized and more strongly sheared than adjacent units. Well preserved, euhedral quartz phenocrysts indicate a volcanic origin, but it is not clear whether the rhyolite was intrusive or extrusive (e.g. Lin et al., 1997). Recrystallized quartz and sulphide minerals form massive competent lenses from 5 cm to 30 cm thick and are oriented parallel to schistosity and bedding. Lenses are typically lineated and boudinaged along north-trending axes. Fracture-controlled sulphide minerals occur in planar quartz veins, and segmented boudinaged quartz veins up to 10 cm thick are also found. Arsenopyrite and pyrite in approximately equal proportions are the most abundant sulphide minerals forming up to 50–80% of the lenses, and are associated with variable proportions of sphalerite, galena, and chalcopyrite, as well as minor bornite, pyrrhotite, and tetrahedrite. Rarely sphalerite or galena are more abundant than arsenopyrite and pyrite. Sphalerite and chalcopyrite show an inverse distribution; samples rich in sphalerite are poor in chalcopyrite. Three of the sites visited contained samples with intergrowths of arsenopyrite-pyrite-pyrrhotite-sphalerite suitable for thermometric and barometric study. The other occurrences were lacking in either sphalerite or pyrrhotite. Detailed microstructural descriptions of the lenses are provided in Jamieson et al. (1987) and Lynch and Mengel (1995). To further characterize the lenses, analyses from nine different occurrences were plotted on a Cu-Pb-Zn ternary diagram showing subdivisions typical of different massive sulphide classes (Franklin et al., 1981) (Fig. 13) (A.L. Sangster, unpub. data, 1992). Most occurrences fall within the Cu-Zn field with some overlap into the Zn-Pb-Cu field. The 'Core shack' occurrence on the other hand displays more scatter overlapping into all fields including Pb-Zn (Fig. 13). The Cu-Zn-rich nature and the association with the Early Silurian rhyolite horizon is consistent with the deposits being characterized as volcanogenic massive sulphide occurrences. Intense shearing however partially masks the characteristics of volcanogenic massive sulphide occurrences.

Electron-microprobe analyses of arsenopyrite, sphalerite, and pyrrhotite are presented in Lynch and Mengel (1995). The arsenopyrite composition is relatively As deficient or S rich. In the stoichiometric formula $FeAs_{1-X}S_{1+X}$, the value for X ranges between 0 and 0.13. Crystals are typically homogeneous in their composition, though one sample showed a marked As enrichment toward the rim. Arsenopyrite is a useful geothermometer in rocks that have been metamorphosed to the greenschist or lower amphibolite facies (Sharp et al., 1985). For the samples analyzed here, temperature determinations show a wide range, from 300°C to 510°C. Accordingly the temperature variation is accompanied by a pronounced change in the sulphur fugacity in the mineralizing environment because of the buffering capacity of coexisting pyrite and pyrrhotite.

Sphalerite displays a restricted composition; the iron content is between 11 and 14 cation per cent, with one value at 16.1 cation per cent, and the greatest frequency lies between 13.0 and 13.5 cation per cent. Such restricted iron content is typical of sphalerite occurring along the pyrite-pyrrhotite solvus, where the composition is approximately constrained to the interval 10–20 cation per cent Fe depending on the conditions of formation (Barton and Skinner, 1979). The iron content of sphalerite is typical for sphalerite which is in equilibrium with pyrite and hexagonal pyrrhotite, with values distinctly higher than that which would be imposed by the pyrite-monoclinic pyrrhotite buffer (11–12 mole per cent FeS in sphalerite, Fig. 6), indicating that analyses likely reflect compositions established at high temperature. Experimental calibration of the sphalerite geobarometer at high pressure (Scott, 1973; Hutchison and Scott, 1981) has revealed that the composition of sphalerite is sensitive to pressure but relatively insensitive to temperature. Estimates of pressure in this study for sphalerite-pyrite-pyrrhotite-arsenopyrite intergrowths range from 5.6 kbar to 7.0 kbar for the sulphide lenses which contained closely intergrown sphalerite-pyrrhotite-pyrite-arsenopyrite, based on the geobarometer of Hutchison and Scott (1981). A compositional range of 13–14 mole per cent FeS in sphalerite was most frequently recorded which corresponds to pressure determinations of 5.6–6.7 kbar; however, the thermodynamic calculations of Toulmin et al. (1991) suggest that pressure determinations are more strongly influenced by temperature, which for this study produce estimates of 5.5–6.9 kbar when combined with arsenopyrite temperature determinations. Although there is some divergence by the application of the two methods, similar results are attained. The range in composition of arsenopyrite

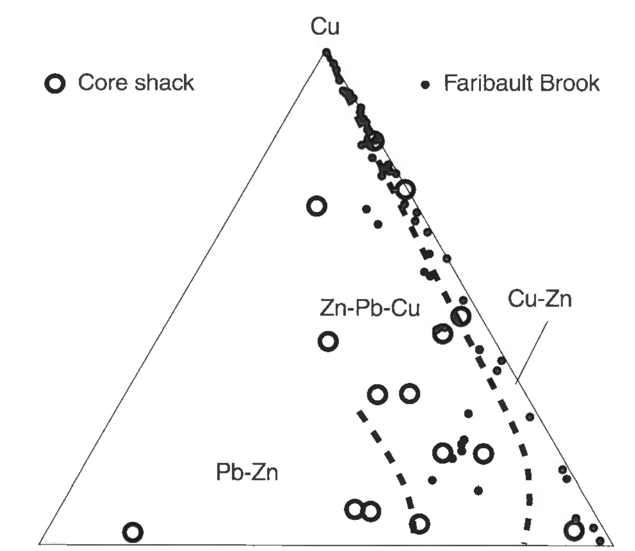


Figure 13. *Ternary Cu-Pb-Zn plot from concordant mineralized lenses within the Jumping Brook metamorphic suite along Faribault Brook (site labelled "sphalerite" on Figure 7, and* see also *Lynch and Mengel (1995) for location). Metal groupings for different types of volcanogenic massive sulphide deposits according to Franklin et al. (1981) and A.L. Sangster (unpub. data, 1992).*

produced while buffered along the pyrite-pyrrhotite solvus provides clear evidence of retrograde conditions during metamorphism. In contrast, sphalerite has a more uniform composition, possibly due to the fact that sphalerite isopleths follow certain geothermal gradients in pressure-temperature space (Fig. 8a) (Toulmin et al., 1991). In the low-grade belt, Plint and Jamieson (1989) bracketed the temperature of metamorphism using the garnet-biotite geothermometer, at approximately 400–450°C which agrees well with the higher temperature arsenopyrite determinations (Fig. 5). Pressure estimates from sphalerite are placed at between 5 and 7 kbar, similar to those for the staurolite zone, providing useful information on the relative motion of thrust blocks; the adjacent high-grade Cabot nappe (Fig. 2) was metamorphosed to 8–9 kbar and subsequently thrust over the low-grade belt along the Highlands shear zone, moving upward by the vertical equivalent of approximately 1–4 kbar, or 3–12 km (Fig. 12). The nappe was subsequently folded and isolated as an erosional klippe.

Maragaree shear zone and Late Devonian extension

Cape Breton Island provides a window through the Devonian–Carboniferous Maritimes Basin (Fig. 1), exposing basement rocks, and providing an opportunity for establishing links between basement processes and basin formation. The early evolution of the large Late Devonian–Carboniferous Maritimes Basin is evidently linked to the erosion of the Canadian Appalachian Mountains and Acadian orogeny, ultimately resulting in exposure of the Cabot nappe of northern Nova Scotia; however, the juxtaposition of deep-seated rocks with the nonmetamorphosed units of the Maritimes Basin, and short time frame between youngest metamorphism at depth within the Cabot nappe (ca. 380 Ma), as well as early sedimentation and volcanism at the surface in the Maritimes Basin (ca. 375 Ma; Barr et al., 1995), indicate the likely influence of extensional faults such as the Margaree shear zone, described in this section, in the rapid upward transport of the high-grade units. With the characterization of metamorphic core complexes (e.g. Crittenden et al., 1980; Armstrong, 1982) and recognition of low-angle extensional faults in the western Cordillera of North America (Armstrong, 1972; Davis et al., 1980; Wernicke, 1981), new examples of shallow-dipping extensional faults are increasingly being recognized in older orogens. Such detachment faults are in part responsible for the destruction of mountain belts, the formation of large basins, crustal thinning, and the denudation of mid-crustal levels. Recent studies in northern Europe have demonstrated that large Devonian and Carboniferous basins evolved above regional detachment faults following the Caledonian orogenic event (McClay et al., 1986; Séranne and Séguret, 1987; Gibbs, 1987; Malavieille et al., 1990; Fossen, 1992). Middle to Late Paleozoic extensional complexes and detachment faults are also reported in the southern Appalachian Mountains (Snoke and Frost, 1990), as well as in the New England Appalachian Mountains (O'Hara and Gromet, 1983, 1985; Goldstein, 1989; Getty and Gromet, 1992a, b).

The initiation of sedimentation and volcanism within the Maritimes Basin in the Late Devonian followed Early Devonian Acadian deformation and metamorphism. Strike-slip models for the origin of the Maritimes Basin are widely proposed (Bradley, 1982; Gibling et al., 1987); however, McCutcheon and Robinson (1987) alternatively suggested on the basis of geophysical magnetic data and geological constraints that the Maritimes Basin originated during extension and crustal thinning as an isostatic response to Acadian crustal thickening. An extensional regime was also made evident through the sedimentological investigations of Hamblin and Rust (1989) on early basin fill. Deep seismic, gravity, and aeromagnetic profiles from the Gulf of St. Lawrence are interpreted to indicate that the centre of the Maritimes Basin is underplated by gabbro derived from the mantle, and that the lower crust was thinned in the Late Devonian–Carboniferous (Marillier and Verhoef, 1989) (Fig. 14).

Extension and early evolution of the Maritimes Basin

In Cape Breton Island, the Margaree shear zone is a regional low-angle extensional shear zone fringing the south and southeastern margins of the Cape Breton Highlands (Fig. 14, 15), forming a structural boundary between underlying basement units affected by Silurian–Devonian metamorphism and plutonism, from the overlying basal member to the Devonian–Carboniferous Maritimes Basin. The Margaree shear zone crosscuts and has affected early basin fill, and is interpreted to have played an important role in the denudation of the underlying high-grade metamorphic rocks and Cabot nappe. The term detachment is used here in the manner of Lister et al. (1991), referring to a shallow-dipping brittle-ductile shear zone formed in an extensional regime. Sedimentary and volcanic rocks in the hanging wall are not necessarily detached directly along bedding planes, but are crosscut by the shear zone. The Late Paleozoic Maritimes Basin consists of a thick marine and nonmarine sedimentary succession (Fig. 1). Rock types include clastic rocks, carbonate, evaporite, and minor volcanic units. The Maritimes Basin is centered in the Gulf of St. Lawrence, but overlaps extensive regions onshore as well. From north to south the Maritimes Basin is approximately 300–400 km wide, and from east to west it is 900 km long. Estimates of the basin thickness indicate locally as much as 9000 m of sedimentary fill onshore (Howie and Barss, 1975), with offshore accumulations of up to 12 km (Marillier et al., 1989). Six groups ranging from Late Devonian to Lower Permian define the Maritimes Basin (Fig. 16). These are from oldest to youngest the Fountain Lake, Horton, Windsor, Mabou, Cumberland, and Pictou groups (Ryan et al., 1991; Fig. 5). The principal clastic cycles occur within the Horton, Cumberland, and Pictou groups, whereas the intervening Windsor and Mabou groups are dominated by marine carbonate, evaporite, and shale, which mark a general transgression in the middle to late Viséan (Giles, 1981). Volcanic rocks of the Fountain Lake Group and Fisset Brook Formation are widespread at the base of the Maritimes Basin and mark the onset of rifting early in the basin history.

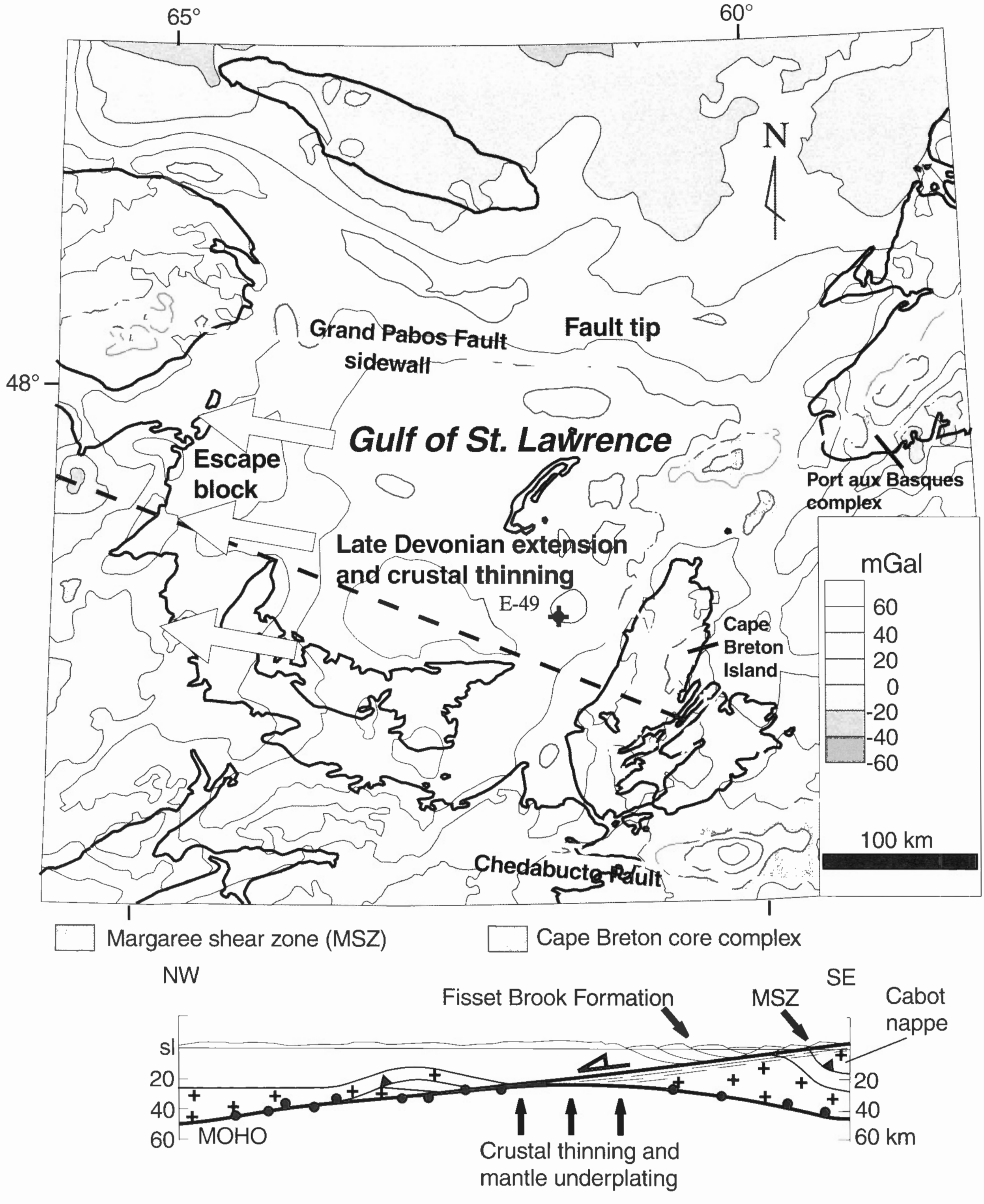


Figure 14. *Regional gravity map* modified from *Marillier and Verhoef (1989) to show relationship between crustal thinning in Gulf of St. Lawrence and extensional exhumation of Cabot nappe from beneath the Margaree shear zone, between strike-slip faults bounding the northern and southern limits of crustal thinning, and escape of upper panel towards the west (large arrows). Position of schematic southeast-northwest cross-section corresponds with dashed line (black), and position of offshore well E-49 shown by dotted cross* (modified from *Lynch and Keller (1998)).*

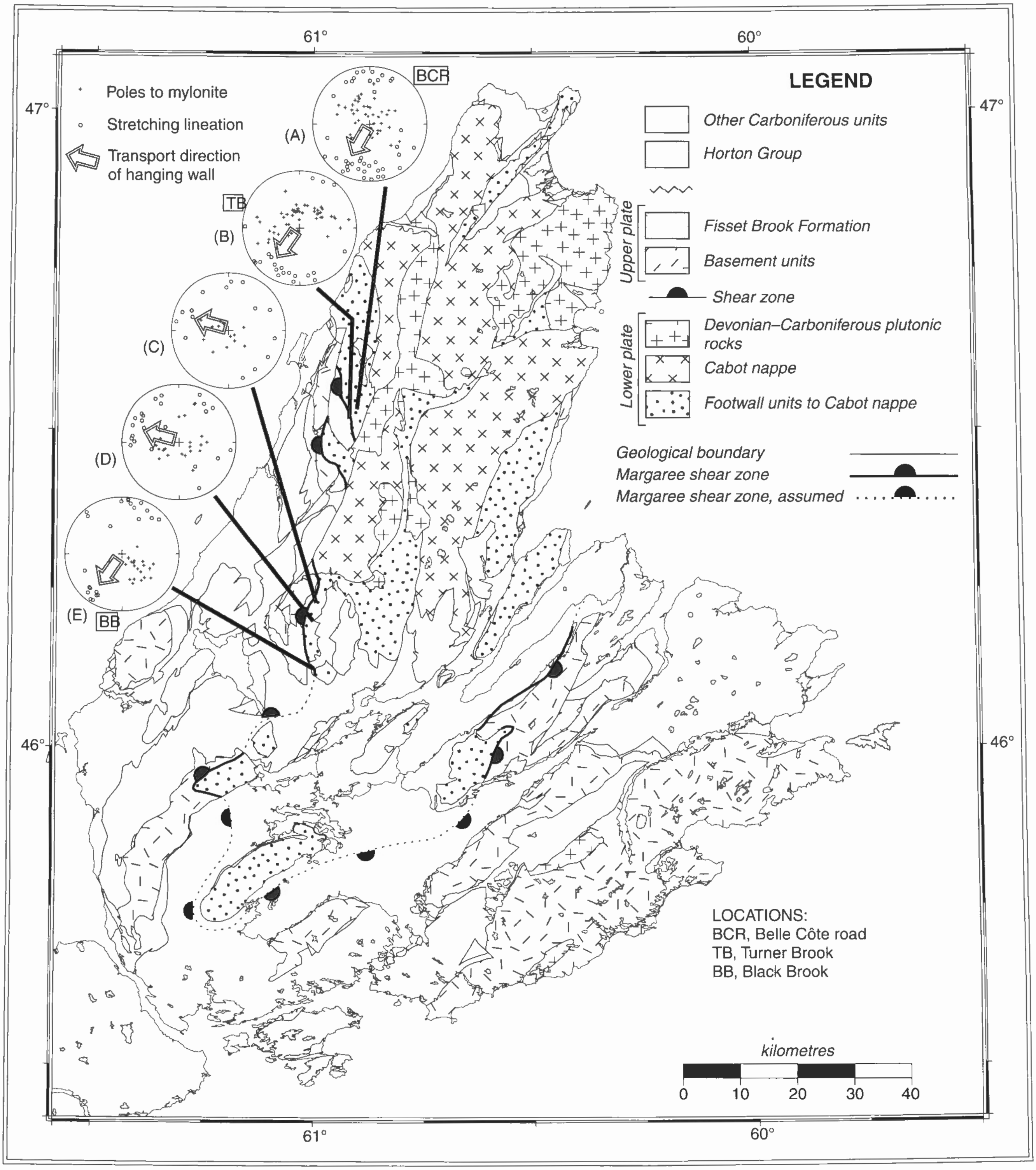


Figure 15. Lower hemisphere stereonet projections of mylonite fabrics from western Margaree shear zone, and subdivision of upper plate and lower plate units. Note Carboniferous (Tournaisian) Horton Group (grey pattern) unconformably overlaps Margaree shear zone, whereas Fisset Brook Formation (mid-Devonian) is crosscut by Margaree shear zone and restricted to hanging wall (modified from *Lynch and Tremblay (1994)*).

Fisset Brook Formation

Bimodal basalt-rhyolite and interbedded siliciclastic rocks of the Fisset Brook Formation occur in western Cape Breton Island (Fig. 15), and can be compared generally to other volcanic-bearing rock units across the Maritime provinces, including the Fountain Lake Group, forming the basal member to the Maritimes Basin (Kelley and MacKasey, 1964; Howie and Barss, 1975; Blanchard et al., 1984). The Fisset Brook Formation is a key unit to understanding the inception of the Maritimes Basin and relationship of the Margaree shear zone with the formation of the Maritimes Basin, because the Fisset Brook Formation is restricted to the immediate hanging wall of the Margaree shear zone in western Cape Breton Island. The age of the Fisset Brook Formation is constrained by radiometric and paleontological data which indicate a Middle to Late Devonian age (Blanchard et al., 1984; Barr et al., 1995) which may extend into the early Carboniferous (Kelley and MacKasey, 1964; Blanchard et al., 1984). A U-Pb age determination on zircon from rhyolite in the Gillanders Mountain area established an age of 373 ± 4 Ma (Barr et al., 1995). The volcanic rocks are thought to be the eruptive equivalents to Late Devonian subvolcanic intrusions of similar age that are widespread in the Cape Breton Highlands, such as the Salmon Pool pluton (Jamieson et al., 1986), or the Black Brook granite (Dunning et al., 1990a). Howie and Barss (1975) include the Fisset Brook Formation within the Horton Group, likely due to the presence of coarse conglomerate within the formation and apparent stratigraphic overlap with the Horton Group. Alternatively, the Fisset Brook Formation might be included within the Fountain Lake Group of western Nova Scotia, which consists of correlative volcanic and clastic rocks of similar age (Blanchard et al., 1984). Volcanic rock geochemical characteristics suggest a within-plate continental setting for the Fisset Brook Formation (Currie, 1982; Dostal et al., 1983; Blanchard et al., 1984; Barr et al., 1995). This is consistent with field observations

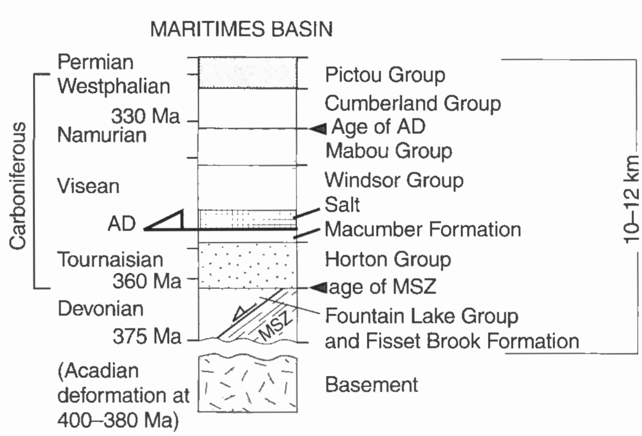


Figure 16. *Simplified stratigraphic column of units from Maritimes Basin from northern Nova Scotia, and relative ages of the Margaree shear zone (MSZ) and Ainslie detachment (AD).*

which show an unconformable contact with underlying Precambrian granite and other basement units (Lynch and Tremblay, 1992a). The depositional environment proposed by Blanchard et al. (1984) is one of early explosive volcanism and clastic alluvial fan-type deposits, followed by basaltic eruptions with coeval fluvial and lacustrine sedimentation. Deposition apparently occurred within a horst-and-graben setting. Although a thickness of up to approximately 500 m is reported for the Fisset Brook Formation (Blanchard et al., 1984), at least one and possibly several important intraformational unconformities, as well as an unconformity with the overlying Horton Group suggest that original thicknesses are not preserved. In the Gillanders Mountain area, Barr et al. (1995) described a lowermost sedimentary unit overlain by basaltic and rhyolitic units of the Fisset Brook Formation. Conglomerate horizons in the lower sedimentary unit consist of polymictic pebble, cobble, and boulder conglomerate, which are interbedded with arkose, lithic arkose, and varicoloured siltstone. Small gabbroic plutons and dykes crosscut the sedimentary rocks. Overlying basalt lavas consist of flows and minor vitric and lapilli tuff. Flow tops and bottoms are typically vesicular and amygdaloidal. Rhyolite may be vuggy, and often consists of eutaxitic to spherulitic flows with well developed flow banding, or consists of more massive lithic lapilli tuff or crystal-lithic tuff. Welded rhyolitic ash flows and accretionary lapilli tuff also occur. Quartz and alkali feldspar phenocrysts are common. The groundmass is locally altered to sericite and radiating bundles of clay minerals.

In detail, basalt contains phenocrysts of plagioclase or olivine with augite, within a groundmass of fine-grained calcic plagioclase. Flow textures are common as is massive basalt, locally with ophitic to subophitic crystal aggregates. Flows are typically vesicular (Fig. 17a). Volcanic breccia and mafic pyroclastic deposits are also common. The basalt is variably metamorphosed from relatively unaffected units to those which contain abundant chlorite, epidote, actinolite, calcite, as well as minor hematite or pyrite. Metamorphism may include epidote with andradite (Blanchard et al., 1984). Zeolite minerals are reported to occur within vesicles (Blanchard et al., 1984). Typically however, vesicles are partially or completely filled with fine-grained epidote, chlorite, and calcite giving the rock a green spotted texture.

The Fisset Brook Formation is unconformably overlain by the Tournaisian Horton Group, consisting predominantly of a thick succession of coarse terrestrial sandstone and conglomerate. Sedimentation in the Horton Group (Bell, 1929) is interpreted to have occurred during active faulting in association with grabens and half grabens containing proximal alluvial fan, medial fluvial, and distal meandering fluvial and/or floodplain deposits, as well as local braidplain and mudflat and/or playa deposits (Hamblin and Rust, 1989). The thickness for the Horton Group is highly variable, and is up to and possibly greater than 3350 m in western Cape Breton Island.

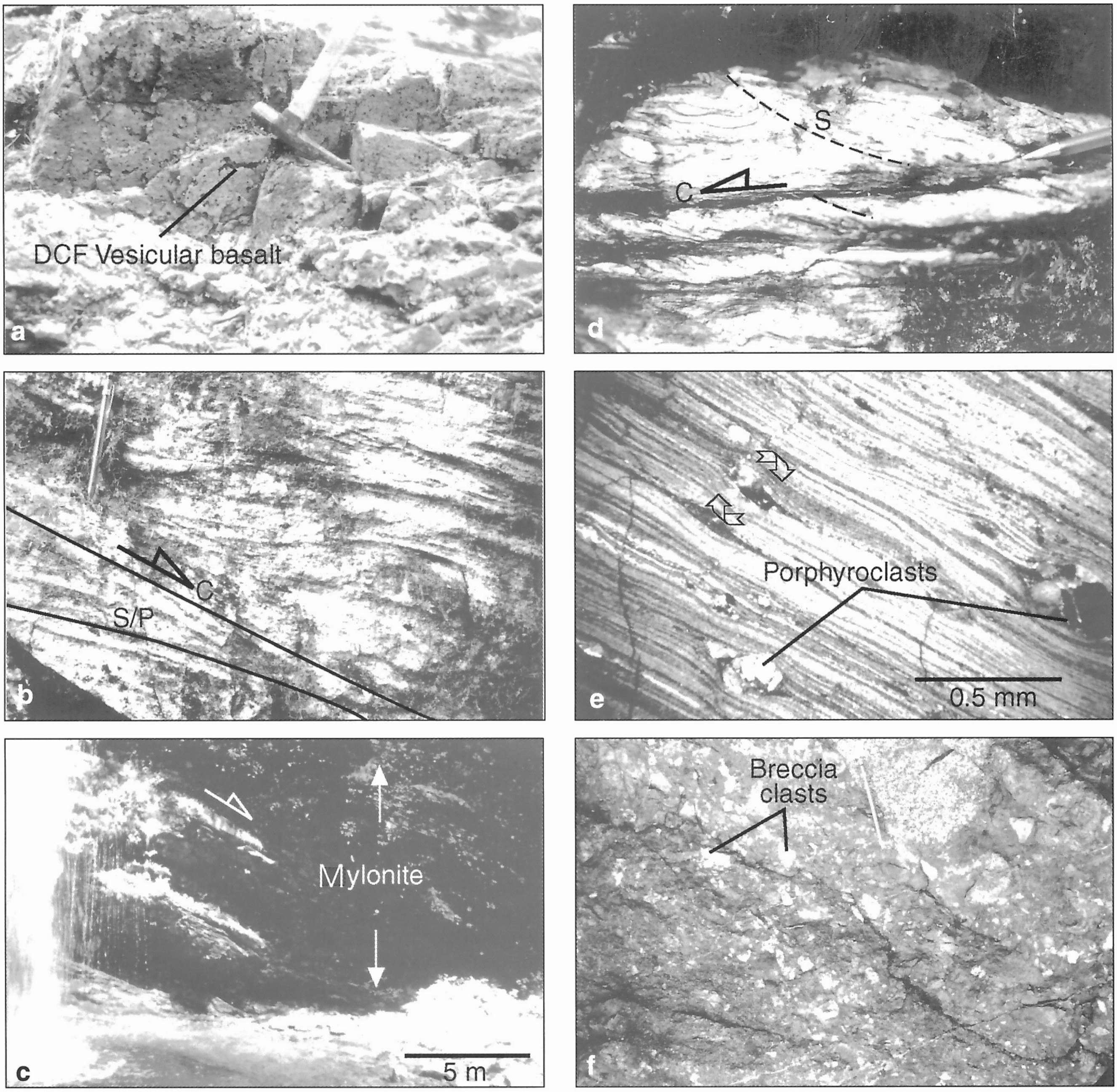


Figure 17. *Photographs from Margaree shear zone:* ***a)*** *vesicular Fisset Brook Formation (DCF) basalt in immediate hanging wall of shear zone along Belle Côte road (photograph by G. Lynch; GSC 2000-041O);* ***b)*** *mylonitized basalt from the Fisset Brook Formation along Turner Brook showing moderately dipping principal shear planes (C-plane) bounding sigmoidal flattening S-planes and subparallel P-shears (photograph by G. Lynch; GSC 2000-041P);* ***c)*** *6 m high outcrop showing portion of Margaree shear zone along Turner Brook, note waterfall on left of photo (photograph by G. Lynch; GSC 2000-041Q);* ***d)*** *shallow-dipping mylonite from Margaree shear zone north of Black Brook in Gillanders Mountain area with well developed C-shears and S flattening planes (photograph by G. Lynch; GSC 2000-041R);* ***e)*** *photomicrograph of mylonitized granitic rock from Margaree shear zone with porphyroclasts of feldspar demonstrating noncoaxial shear;* ***f)*** *tectonic breccia along the top of the Margaree shear zone along Belle Côte road with granitic clasts and clasts of basalt in matrix of chlorite, goethite, hematite and cataclasite, pencil for scale 15 cm (photograph by G. Lynch; GSC 2000-041S).*

Margaree shear zone

Typically, the Margaree shear zone is 200–300 m thick, but may reach up to 1 km in true thickness. Evidence of retrogression is recorded in the mineralogical paragenesis and in the evolution of fault textures; ductile feldspathic mylonite and mylonitic gneiss are overprinted by quartz-muscovite mylonite, which is overprinted by cataclasite, concordant brittle shear, and chloritic breccia. Brittle detachment faults have incised and cut the ductile mylonite. Mylonite is shallow dipping to nearly flat lying and rarely steeply inclined (Fig. 15, 17b, c, d). Stretching lineations trend and plunge to the south-southwest, west, and north-northwest (Fig. 15). Shear-sense indicators (e.g. Hanmer and Passchier, 1991) such as C-S fabrics, winged inclusions, rotated mineral trails, and asymmetric extensional shear bands have been documented in the field as well as from oriented thin sections, and consistently demonstrate westerly transport of the hanging wall (Fig. 15, 17).

Indirect evidence for the age of unroofing from beneath the Margaree shear zone is widespread. The $^{40}Ar/^{39}Ar$ and K-Ar data for biotite and muscovite from basement rocks in northwestern Cape Breton Island indicate that much of this region was still buried to relatively high-temperature conditions in the Middle Devonian (ca. 375–365 Ma) (Currie, 1987; Plint and Jamieson, 1989; Reynolds et al., 1989; Keppie et al., 1992). Footwall rocks to the Margaree shear zone include high-grade gneiss and upper greenschist-facies assemblages from metasedimentary and metavolcanic rocks metamorphosed in the Early Devonian. Age constraints on faulting are provided by crosscutting relationships with sedimentary fill, and are detailed in the next section. The Late Devonian Fisset Brook Formation is crosscut by the Margaree shear zone and is restricted to the hanging wall of the zone (Fig. 15); fresh vesicular basalt and redbeds of the Fisset Brook Formation can be traced downward into the Margaree shear zone across a strain gradient featuring progressive shear and mylonitization (Fig. 17b), with the production of greenschist-facies mineral assemblages. An upper limit on faulting is obtained from carbonate rocks of the Viséan Windsor Group which unconformably overly the Margaree shear zone. Tournaisian coarse clastic rocks of the Horton Group are also unconformable on the Margaree shear zone and contain clasts of mylonite. These relative age constraints for the Margaree shear zone are schematically illustrated in Figure 16.

Relatively high-temperature mylonite evolved in basement rocks within the footwall of the Margaree shear zone, and is discontinuously distributed along the Margaree shear zone. The high-temperature mylonite is characterized by feldspathic mylonite, and microstructures within feldspathic mylonite demonstrate the development of laminated orthoclase ribbons, separated by thin shear planes of fine-grained rotated subgrains. Strain and undulatory extinction are preserved within feldspar. Typically, layers of feldspathic mylonite are boudinaged and segregated into augen mantled by muscovite, quartz, minor biotite, and fine-grained trails of feldspathic subgrains. Feldspar porphyroblasts are common (Fig. 17e). Asymmetry of mineral trails indicates a strong component of noncoaxial shear. The feldspathic boudins acted as rigid bodies within the mylonite during retrogression, forming winged inclusions containing asymmetric quartz-muscovite pressure shadows; C-S fabrics defined by muscovite and biotite typically occur. Polycrystalline monomineralic quartz ribbons locally define the mylonitic fabric; strained and recrystallized quartz show parallel extinction at a moderate angle to the flow plane, with flattening at a high angle to the apparent bulk minimum instantaneous stretching axis. Layers of strain-free recrystallized quartz are also observed. Quartz may display core and mantle textures featuring irregular, serrated grain boundaries developed during grain boundary migration and subgrain rotation.

Although noncoaxial shear is evident within the Margaree shear zone by the occurrence of asymmetric mineral trails, winged inclusions, quartz ribbons with inclined extinction, core and mantle textures, and C-S fabrics, a component of pure shear is also apparent. Mylonitized granite and volcanic rocks display individual subhedral feldspar crystals of igneous plagioclase which have been segmented along cleavage, stretched, and distributed along the flow plane; limited rotation of these separated lath-like fragments suggests deformation by pure shear. The existence of pure and simple shear components indicate that the Margaree shear zone may have evolved under conditions of general shear (e.g. Hanmer and Passchier, 1991; Simpson and DePaor, 1993).

A consistent temporal evolution of fault-rock types is observed from early ductile fabrics, to brittle-ductile, to late cataclasis. Cataclastic zones form broad domains and irregular map units. The zones contain a weak flat-lying foliation, or more typically, no fabric. Cataclastic faults also occur as thin horizons (50 cm) along the mylonitic foliation. Breccia is quite variable, and may have coarse angular fragments (Fig. 17f), or fault surfaces may be dominated by pea-sized clasts. Microbreccia was also found in some areas rendering protoliths difficult to map. Matrix to the breccia is chloritic, and may locally contain hematite and goethite. Closely spaced sets of late-stage shear planes developed during fluid-assisted cataclastic flow. Brittle deformation has affected the rocks through distributed frictional sliding and cataclasis of polymineralic aggregates. Fine-grained muscovite, chlorite, epidote, and calcite have crystallized along spaced shear planes. Quartz-fibre growth and crack-seal textures developed along microscopic tension gashes. Shear surfaces are transgranular and contain angular abraded fragments of silicate minerals. Late-stage movement is further highlighted by slickensided surfaces and striae overprinting hematite laminae which infiltrated the mylonitic fabric.

Mylonite and associated fault-rocks of the Margaree shear zone are mineralogically and microstructurally variable due to varying degrees of recrystallization, cataclastic abrasion, and grain-size reduction induced during shear under evolving conditions of retrogressive metamorphism. The Margaree shear zone was progressively exhumed during extension producing late-stage brittle shear and lower temperature mineral assemblages.

Mylonitized Fisset Brook Formation and timing of the Margaree shear zone

In western Cape Breton Island field evidence, crosscutting relationships, and overlapping strata demonstrate that the Margaree shear zone was active during the period between deposition of the Devonian Fisset Brook Formation and Carboniferous Horton Group (Fig. 16). Major- and trace-element geochemistry was used by Lynch (1996b) to link mylonitic rocks to formations, the results of which are partly summarized here, and enhanced by the addition of two new geochemical plots further characterizing the volcanic rocks (Fig. 18). The Margaree shear zone is well exposed along three branches of Turner Brook where mylonite, schist, and fault breccia outcrop across a width of 1–2 km. The mylonite dips moderately to the southwest and the true thickness of the shear zone is approximately 500–700 m. Mylonitic basement units include diorite, granitoid rocks, and metabasalt or chloritic schist from the Jumping Brook metamorphic suite. The outcrop section along the northwestern branch of the Turner Brook features a 100 m wide lozenge or fault horse of vesicular Fisset Brook basalt, red siltstone, and clastic rocks, fault bounded above by mylonitized Fisset Brook basalt, and below by mylonitic basement units. The mylonitic Fisset Brook basalt is finer grained and does not have the schistose aspect of mylonitic basalt from the basement units, and is mineralogically characterized by a low-grade assemblage of chlorite-epidote-actinolite-calcite and very fine-grained abraded plagioclase. A stretching lineation is well developed. Vesicular basalt passes downward into mylonitic basalt with well developed C-S fabrics indicating southwest transport (Fig. 17a, b). In thin section porphyroclasts of plagioclase form asymmetrical winged inclusions which, together with asymmetrical shear bands, suggest southwest transport of the hanging wall. Large clots up to 1–2 cm in length and dominated by fine-grained epidote form porphyroclasts, presumably derived from amygdules in the basalt, or replaced plagioclase phenocrysts. Cataclasite and fault breccia cover wide areas and feature a well indurated brown oxidized rock-flour matrix. Breccia clasts of sheared basalt may be as large as 20–30 cm, and are subrounded to angular in shape (Fig. 17f). Thin cataclasite layers may be entirely contained within the mylonite zone, and follow the principal fabric. Hematite is also common within the shear zone where it covers slip planes and has been overprinted by fault striations.

In the area surrounding Turner Brook the Fisset Brook Formation consists of a thick succession of interbedded vesicular basalt, volcanic breccia, epiclastic sandstone, red siltstone, sandstone, and arkosic conglomerate. The succession is separated into upper and lower members by a distinct cobble conglomerate horizon which marks an intraformational unconformity by virtue of vesicular basalt and redbed siltstone clasts derived from subjacent portions of the formation, as well as by the presence of a variety of basement-derived clasts. It is noteworthy that no mylonite clasts were observed within the conglomerate. Bedding within the Fisset Brook Formation also dips moderately to the west-southwest and is nearly parallel to the shear zone. Rhyolitic lapilli tuff occurs at the top of the succession of Fisset Brook Formation volcanic rocks in the hanging wall of the Margaree shear zone.

Geochemistry of fresh and mylonitized Fisset Brook basalt

Geochemical analysis of mylonitized Fisset Brook Formation basalt was undertaken to confirm the presence of the formation in the Margaree shear zone (Lynch, 1996b). This adds to an extensive geochemical database which already exists for the Fisset Brook Formation as well as for basalt from the older Jumping Brook metamorphic suite in western Cape

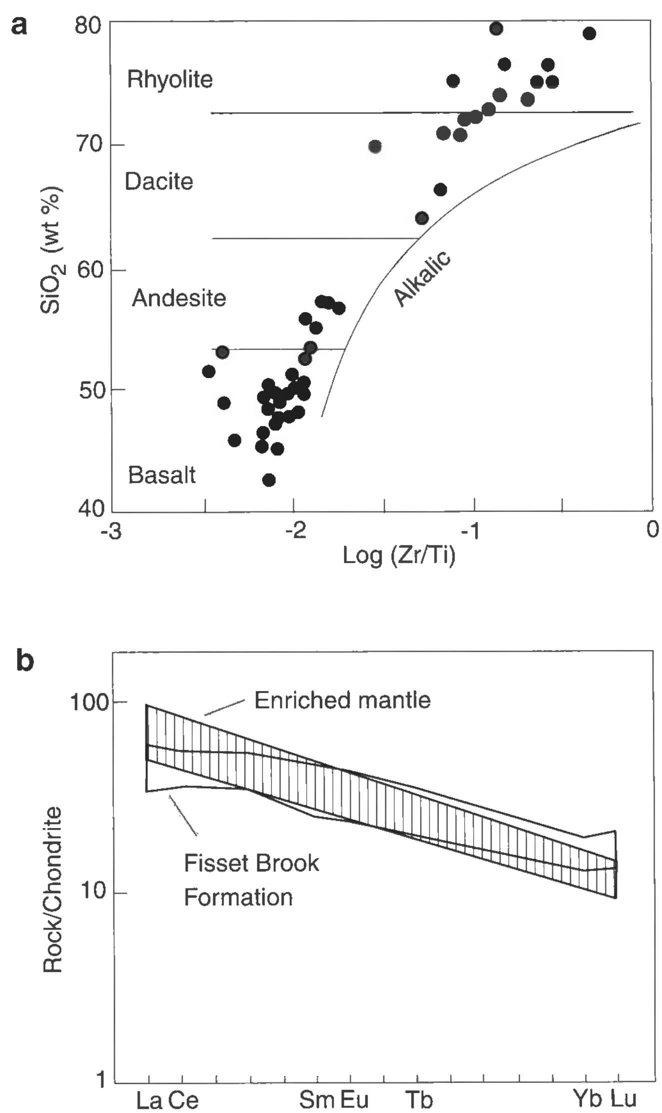


***Figure 18. a)** SiO_2 versus Zr/Ti plot for Fisset Brook Formation compiled from data in Blanchard et al. (1984) and Lynch (1996b) to show bimodal nature of volcanic rocks; **b)** new REE plot which includes fresh basalt and mylonitic basalt from Fisset Brook Formation for samples from Turner Brook (see Lynch (1996b) for detailed sample location), and comparison with REE values for basalt derived from light REE-enriched mantle.*

Breton Island (Blanchard et al., 1984; Connors, 1986; Lynch, 1996b). The bimodal nature of fresh, unstrained Fisset Brook Formation volcanic rocks is well illustrated in the SiO_2 versus Zr/Ti plot of Figure 18a, showing that basalt and rhyolite dominate. This feature is often typical of intercontinental volcanic systems related to extensional environments (e.g. Blanchard et al., 1984). Of further interest is the range of REE values for Fisset Brook Formation basalt which demonstrate significant light REE enrichment (Fig. 18b). For comparison with fresh Fisset Brook Formation basalt from the hanging wall of the Margaree shear zone, and with basalt of the Jumping Brook metamorphic suite in the footwall of the shear zone, six samples of mylonitic basalt from different sites in the Turner Brook area were analyzed for trace- and major-element abundances (Lynch, 1996b). Because of the sheared nature of the rocks, the plots focus on elements which are considered to be relatively immobile, such as Al, Ti, Zr, Y, Ga, and Cr (Fig. 19). In particular Ti (O'Hara and Blackburn, 1989) and Al (Selverstone et al., 1991) are recognized as immobile elements in geological settings where rocks have been mylonitized. Plots are illustrated as elemental ratios in order to eliminate enrichment or dilution factors caused by the loss or introduction of mobile elements.

All samples used are basic in composition, with a range of SiO_2 from 41 to 57 weight per cent for the mylonite, from 44 to 54 weight per cent for the Fisset Brook Formation, and from 44 to 51 weight per cent for the Jumping Brook metamorphic suite (Lynch, 1996b). Distinct geochemical fields can be ascribed to the Fisset Brook Formation and to the Jumping Brook metamorphic suite, with generally limited overlap between the two.

Plots generated as elemental ratios (Fig. 9, 10) have two main advantages: 1) trends may be enhanced and groupings are more clearly established, and 2) enrichment and dilution factors, caused from the leaching or introduction of mobile elements, are reduced if the elements chosen for the plot are relatively immobile, therefore showing the sample's original ratio signature. This second point is particularly important for this study as metamorphosed, nonmetamorphosed, and mylonitized basalt are all compared. Major elements plotted as ratios with Al_2O_3 indicate that the higher Ti and P, and lower Mg in the mylonite, compared to the Jumping Brook metamorphic suite, are likely primary features, and that the mylonite is geochemically similar to basalt from the Fisset Brook Formation (Lynch, 1996b). The same generalization can be made for the trace elements, where ratios of Y/Cr, Zr/Cr, and Ga/Cr for the mylonite clearly plot in the field defined by the Fisset Brook Formation basalt, and are distinct from the ratios recorded in basalt from the Jumping Brook metamorphic suite (Fig. 19).

Evidence for the upper age limit of the Margaree shear zone in the Black Brook area

An upper age limit for the Margaree shear zone is established through contact relations with the Tournaisian Horton Group. Regionally the Horton Group unconformably covers the Margaree shear zone at map scale, resting directly on footwall and hanging wall units east and west of the shear zone in western Cape Breton Island. East of Lake Ainslie along Black Brook, the Margaree shear zone has been mapped where it occurs as a thick shallow-dipping mylonite zone beneath the

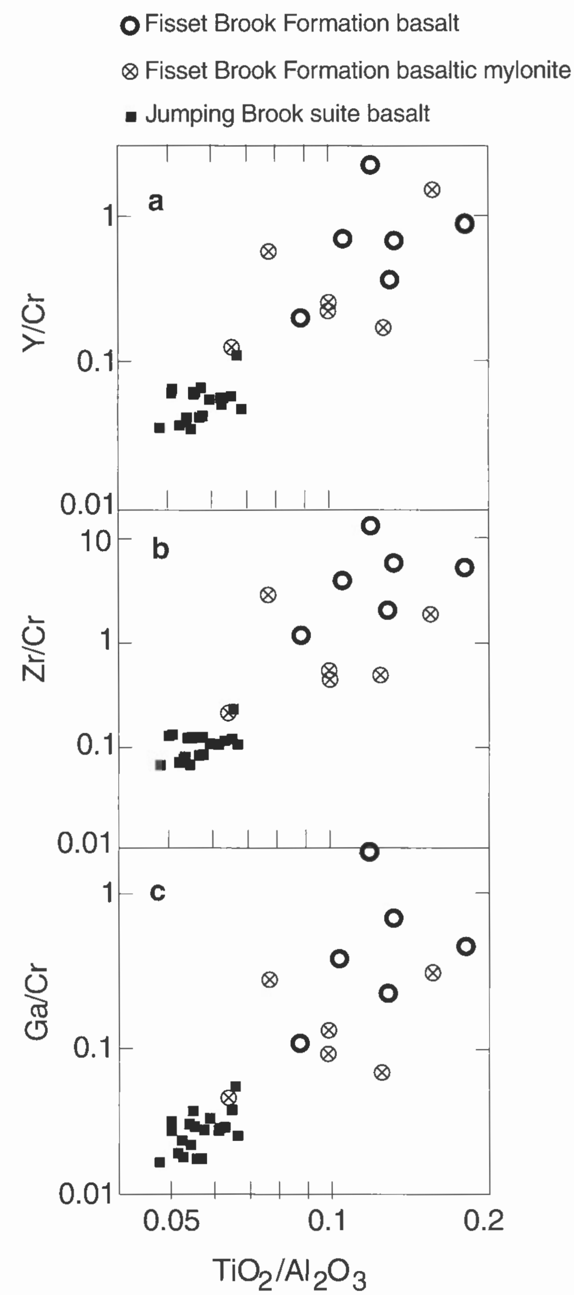


***Figure 19.** Trace element ratio plots for samples from fresh Fisset Brook Formation, mylonitic Fisset Brook Formation, and schistose basalt from the Jumping Brook metamorphic suite in the Turner Brook and Faribault Brook areas, demonstrating that the mylonitic samples can be correlated to the Fisset Brook Formation (*modified from *Lynch (1996b)).*

Fisset Brook Formation (Fig. 15). An unconformity with mylonite from the shear zone is best exposed on Black Brook near Gillanders Mountain (Lynch, 1996b). There, a clean contact can be observed with well sorted cobble conglomerate unconformably overlying thinly laminated mylonite. The clast-supported conglomerate contains rounded clasts up to 15–20 cm, of predominantly white quartzite and granitic fragments in a pebbly arkosic groundmass with abundant detrital muscovite. Mylonite clasts are subordinate but may be up to 20–30 cm and are typically subrounded to subangular. Rounded clasts of fresh to slightly epidotized, highly vesicular basalt derived from the Fisset Brook Formation are dispersed within the conglomerate, as well as subordinate clasts of red siltstone and arkose. The conglomerate is thickly bedded, with bedding running parallel to or at a slight angle to the mylonitic fabric, demonstrating that the mylonite was flat lying at the time of Horton Group sedimentation.

The mylonite in the Black Brook area is exposed across many outcrops with a map width of approximately 2 km in a northeast-trending belt; true thickness of the shear zone is approximately 1 km. Mylonite dips moderately to the west-northwest (Fig. 15), with stretching lineations plunging northeast-southwest (Fig. 15). Shear-sense indicators such as C-S fabrics and winged inclusions demonstrate southwest transport of the hanging wall. The mylonite consists of predominantly sheared conglomerate (of unknown origin) and possibly volcanic breccia or tuff. Fragments within the mylonite are highly strained showing extreme aspect ratios, and smooth rounded outlines. Quartzite is a common and easily identifiable clast type. Other types include granitic and dioritic rocks as well as possibly aphanitic volcanic clasts. The latter clast type is typically highly chloritized and epidotized making unequivocal identification difficult. The size of clasts is highly variable. Although pebble or lapilli sizes are most abundant, intervals occur which contain abundant cobble-sized fragments. The protolith to these rock types cannot be definitively identified, but the rocks clearly include conglomerate and volcanic units, possibly from the Fisset Brook Formation. Fresh Fisset Brook Formation vesicular basalt occurs along the ridge top above the mylonite, stratigraphically immediately above a few outcrops of green chloritic and epidote-rich mylonitic basalt and white mylonitic rhyolite, also possibly of the Fisset Brook Formation.

Regional significance of the Highland shear zone and Margaree shear zone

Rapid denudation of metamorphic rocks on Cape Breton Island is indicated by the relatively short time interval between metamorphism (ca. 380 Ma) and deposition of nonmetamorphosed sedimentary and volcanic rocks (ca. 375 Ma) at the base of the Maritimes Basin. Adjacent metamorphic and nonmetamorphic units highlight the existence of a significant crustal gap. Both the Highlands shear zone and Margaree shear zone, in succession, contributed to the ultimate exhumation of these deep-seated rocks (Fig. 12, 20).

The thrust emplacement of the Cabot nappe and its contained gneissic rocks has resulted in partial denudation of the nappe as evidenced by the greenschist-facies assemblages in the footwall of the nappe and inverted isograds; however, the final exhumation was accommodated during extension and shear along the Margaree shear zone, as indicated by the fact that the Fisset Brook Formation does not overlap onto the high-grade rocks. Although the metamorphic rocks of the Cape Breton Highlands are characterized by very distinct metamorphic and structural features, they are not entirely restricted to this local area and may be correlated with similar rocks elsewhere; high-grade gneiss metamorphosed to high-pressure and high-temperature conditions in the Early to mid-Devonian are exposed to the south on mainland Nova Scotia in the Liscomb Complex (Clarke et al., 1993), to which the Cabot nappe may have been linked at depth prior to its final extensional exhumation; furthermore, to the north in southwestern Newfoundland, gneiss from the Port aux Basques Complex displays similar characteristics to that in the Cabot nappe in its Silurian peak metamorphic signature and associated Devonian granitic intrusions (Burgess et al., 1995) (Fig. 14). Together these gneissic rocks outline a broad discontinuous gneiss belt in the central Canadian Appalachian Mountains. Silurian and Devonian cooling ages similar to those in Cape Breton Island are recorded in western Newfoundland (e.g. Cawood, 1993), with significant periods of uplift and erosion indicated between 400–380 Ma and 430–410 Ma (Fig. 8b). Furthermore, to account for the sedimentological budget of erosion, implied from the denudation of high-grade rocks, units which may have captured detritus during this time period include the Gaspe and Chaleurs groups of Gaspésie (Fig. 8b).

Correlations may also be carried further, and the metamorphic rocks of the Cape Breton Highlands have very similar metamorphic and structural characteristics to those contained in the large thrust nappes of the Scandinavian Caledonides. The Caledonian nappes of Norway provide a close counterpart to the Cabot nappe of the Canadian Appalachian Mountains, in terms of structural style, metamorphism, cooling histories, and exhumation mechanisms. Albeit these nappe complexes on either side of the Atlantic Ocean are oppositely verging toward their respective cratons, such a comparison is useful in order to better place the Cabot nappe into its structural and global tectonic context by viewing the Caledonian and Appalachian orogenic belts as partly contiguous orogens in the Paleozoic. Furthermore these nappe complexes demonstrate how Grenvillian rocks may be incorporated into much younger deformation events. Although links between the Caledonian and Appalachian orogenic belts have long been recognized, the proliferation of new data warrants revisiting. Both the Caledonian and Cabot nappes contain basement rocks derived from Laurentia or equivalent rocks, and include widespread marble, quartzite, and various gneissic rocks metamorphosed to 8–10 kbar and 650–730°C in the Silurian (Hodges and Royden, 1984; Steltenpohl and Bartley, 1987; Currie and Lynch, 1997). Metamorphic cooling ages tracking denudation in the Caledonides span the interval 425–373 Ma (Chauvet and Dallmeyer, 1992; Coker et al., 1995), which is similar to the range of 428–370 Ma recorded in the Cabot nappe, although

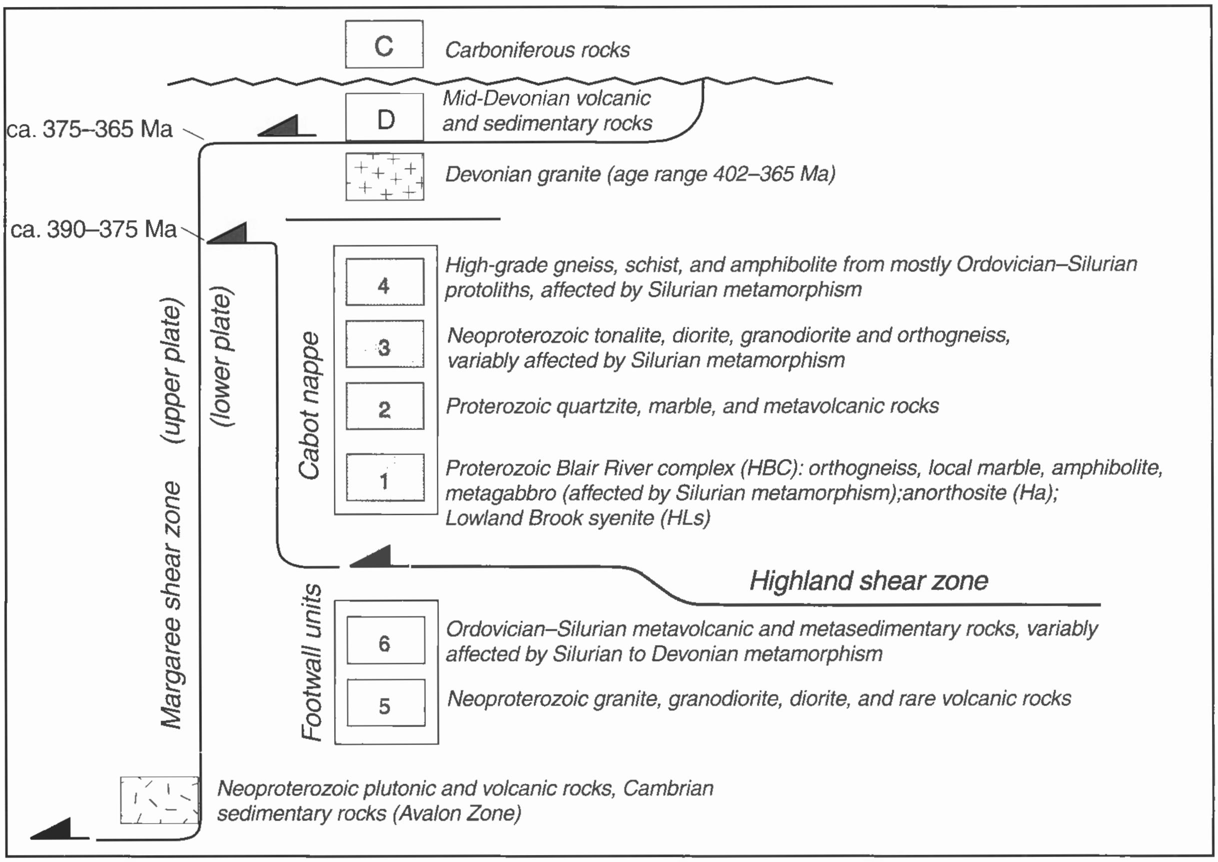


Figure 20. *Schematic representation of faulting events leading to complete exhumation of Cabot nappe; units in Cabot nappe were metamorphosed to lower crustal levels in the Silurian and thrust transported in hanging wall of Highlands shear zone to mid-crustal levels in the Early Devonian as shown by shear symbols climbing up towards the left; further denudation occurred in Upper Devonian time during extension and rise of footwall units from beneath the Margaree shear zone, shown by shear symbol downstepping towards the left, prior to being unconformably covered during the Carboniferous.*

both regions preserve vestiges of older metamorphic and/or igneous events. Inverted metamorphic gradients recorded by well defined isograds along the margins of the overthrust nappes are also characteristic of both regions (e.g. Andreasson and Lagerblad, 1980; Lynch, 1996a). The high-grade nappes in the two regions are folded and preserved as klippes in the hinge zones of synclinal structures, with bounding thrusts and their kinematic indicators rotated away from their original positions (Andreasson and Lagerblad, 1980; Northrop, 1996; Lynch, 1996a). Finally, the Cabot nappe and Caledonian nappes of southwestern Norway are contained within their respective lower plates of low-angle, crustal-scale extensional shear zones (e.g. Chauvet and Séranne, 1994; Lynch and Tremblay, 1994), which were responsible for the final exhumation of the thrust and nappe complexes to surface, while similar mid- to late-Devonian nonmetamorphosed sedimentary and/or volcanic rocks were detached above in the upper plates.

Root zone of the Cabot nappe and extensional exhumation

The Cabot nappe is isolated as an erosional klippe. Apparently, the root zone to the Cabot nappe and Highlands shear zone is not presently exposed, presumably due to truncation by the extensional Margaree shear zone and cover rocks of the Maritimes Basin. It is, however, speculated that the Highlands shear zone splays from an underlying décollement whose surface expression might be represented by a fault such as the Grand Bruit fault zone of southwestern Newfoundland (O'Brien et al., 1993). Detailed descriptions of the Grand Bruit fault zone demonstrate that it was active during several periods of the early Palaeozoic and that it exercised a fundamental tectonic control on the leading edge of the converging Gondwanan continent (O'Brien et al., 1993). Large-scale folding of the Cabot nappe and Highlands shear zone suggests that they were transported over a ramp in an underlying thrust.

The upper plate to the Margaree shear zone includes the Avalon Zone south of the Cabot nappe and Cape Breton Highlands. The Cabot nappe was separated from its root zone during extensional exhumation in late Devonian time, providing a window through to deeper crustal phenomena. This is characteristic of numerous orogens which have undergone extension, and has the effect of disrupting the surface trace of linear belts of rocks, resulting in only "patches" of high-grade rocks at the surface. As such, the units in the footwall of the Margaree shear zone exposed in the Cape Breton Highlands define a metamorphic core complex, similar to the Tertiary core complexes of the western Cordillera of North America (e.g. Armstrong, 1982). Transport of the Cabot nappe with its bounding shear zone and segments of the footwall to the thrust, upward from beneath the Margaree shear zone during extensional exhumation (*see* cross-section in Fig. 14) accounts for why there is currently no surface trace of the root zone to the thrust. With restoration along the Margaree shear zone, the Cabot nappe and its contained Late Devonian plutons were uplifted from a position at depth to the immediate west-southwest of Cape Breton Island during extensional unroofing. Regional geophysical data (Marillier and Verhoef, 1989) indicate that significant crustal thinning and crustal underplating occurred immediately to the west of Cape Breton Island in Late Devonian time (Fig. 14). This provides ample material balance for the extensional unroofing of the Cabot nappe in a crustal-scale simple shear model.

The tectonic evolution of Cape Breton Island from the Silurian to Carboniferous is schematically represented in Figure 20. Both thrusting events described here can be categorized according to well known Appalachian orogenic events; the Early Devonian nappe emplacement is a hinterland feature of the Acadian orogenic episode, whereas the earlier imbricate thrusting and folding more likely correspond to the Silurian Salinic orogenic event (e.g. Dunning et al., 1990b; O'Brien et al., 1991, 1993; Rast and Skeehan, 1993). The Silurian shortening event may reflect convergence following subduction and arc magmatism, whereas the Devonian shortening event likely resulted from the final culmination of convergence due to continent-continent collision between Laurentia and Gondwana. The major compressional phases were succeeded by Late Devonian to Carboniferous extension (Hamblin and Rust, 1989; Lynch and Tremblay, 1994). Complete unroofing of the Cabot nappe occurred during low-angle detachment faulting along the Margaree shear zone of western Cape Breton Island. The Margaree shear zone occurs as an extensional step-over fault to large strike-slip systems such as the Grand Pabo fault (Malo and Bourque, 1993) and Cobequid–Chedabucto fault which form the lateral boundary to the region of crustal thinning in the Gulf of St. Lawrence (Fig. 14), and were also active in the mid- to Late Devonian.

Implications for deep crustal structure and LITHOPROBE seismic data

The documentation of Late Devonian–Carboniferous detachment faults has important implications for deep crustal structure. The retrogressive nature of the Margaree shear zone and unroofing of upper greenschist- to amphibolite-facies rocks in its footwall demonstrate that the Margaree shear zone is part of a large-scale extensional feature, that may be linked to the regional gravity anomaly that dominates the Gulf of St. Lawrence (Fig. 14).

Deep seismic reflection data generated as part of the Canadian LITHOPROBE program (Marillier et al., 1989) provide a view into the lower crust. Data were gathered in a line running northwest through the Gulf of St. Lawrence between Cape Breton Island and Newfoundland (lines 86-4 and 86-5, Marillier et al., 1989). A striking feature highlighted by the data is what appears to be an irregular Moho beneath the Canadian Appalachian Mountains. At the northwest end of the section the Moho rises abruptly from 15 s two-way traveltime to 11 s. The rise of the Moho is nearly symmetrically mirrored by the overlying Late Devonian to Carboniferous Saint George's Basin of southwestern Newfoundland, where the basement has been thinned to 8 s two-way traveltime or approximately 24 km thickness. A reflective surface which dips moderately to the southeast appears to have acted as a detachment beneath the basin, and extends down to the Moho, accommodating crustal thinning in the Late Devonian to Carboniferous.

Much of the seismic section (lines 86-4, 86-5) is underlain by a relatively shallow Moho and broad region of highly reflective lower crust with largely nonreflective upper crust. The principal detachment is interpreted as the interface between the highly reflective lower crust and nonreflective upper crust. The boundary possibly represents the brittle–ductile transition, with the lower crust affected by large-scale penetrative strain and extensional shear in the Late Devonian–Carboniferous. This fault could be the offshore subsurface continuation of the Margaree shear zone.

Summary, Silurian–Devonian structures

This section characterizes the Silurian–Devonian structural evolution of Cape Breton Island, in the central Canadian Appalachian Mountains. Although significant reviews of the bedrock geology for this region have recently been published (e.g. Raeside and Barr, 1992; Horne, 1995) relatively little was known about the nature of major faults and folds in the region, and a structural context was mostly lacking.

Field and laboratory work presented here characterize two major thrusting events in Cape Breton Island which correspond, respectively, to Silurian (425–410 Ma) and Devonian (390–375 Ma) periods of intense shortening in the northern Appalachian Mountains. Furthermore, a period of Late Devonian shear and postorogenic extension has now also been recognized. In Cape Breton Island the Silurian event features imbrication of Ordovician–Silurian volcanic arc assemblages with Neoproterozoic basement slivers during south-directed thrusting following a period of subduction and arc volcanism (ca. 433 Ma), resulting in tectonic burial and high-grade metamorphism. Although significant transport during this phase of thrusting is apparent from the juxtaposition of the Chéticamp granite over younger volcanic and sedimentary units, and by the development of thick mylonite, the thrusting event is restricted to west-central Cape Breton

Island. The second phase of shortening features the thrust emplacement of a single large amphibolite-grade nappe (the Cabot nappe) over greenschist-grade footwall rocks during northwest-directed thrusting in the Early Devonian. The Highlands shear zone at the base of the Cabot nappe is a folded fault surface that outcrops in different areas of Cape Breton Island defining a broad synclinal structure, and the fault separates highly contrasting metamorphic domains. Strain rate apparently exceeded the cooling rate during thrusting producing shear heating and dynamothermal metamorphism of footwall units, with the development of the staurolite isograd producing a useful regional marker for delineating the outer limits of the nappe. The Cabot nappe is a well defined, mappable feature that contrasts strongly with rocks which are adjacent to and outside of the nappe. In particular, the juxtaposition of deep-seated rocks (25–30 km, metamorphosed at 8–9 kar) against significantly shallower rocks (15–20 km, metamorphosed at 5–7 kbar) (Lynch and Mengel, 1995; Currie and Lynch, 1997) across a thick shear zone, which can be traced at surface for many kilometres, clearly defines the limits of the nappe. The metamorphic and structural characteristics indicate the overthrust Cabot nappe that blankets much of the Cape Breton Highlands is allochthonous. Furthermore, recognition of the folded nature of the nappe (Lynch, 1996a) has resulted in the redefinition of what constitutes footwall and hanging wall units along the eastern side of the nappe (e.g. Lin, 1992). The characterization of this thrust sheet is of fundamental importance to our understanding of the Appalachian orogen and orogenic processes in general. The Cabot nappe is of particular interest because it demonstrates how transport of high-grade rocks during exhumation is sequentially partitioned into phases of 1) nappe thrusting and 2) extensional metamorphic core complex formation. Inverted isograds around the Cabot nappe are interpreted to indicate that denudation of the high-grade rocks began during the thrusting phase in the Late Silurian to Early Devonian, characterized by downward metamorphic younging through the nappe, before exhumation was completed during Late Devonian extension along the Margaree shear zone.

A significant result of this study is the recognition that thick extensional mylonite zones are as young as Late Devonian. A lower age limit is placed on the Margaree shear zone by the identification of mylonitic basalt from the Late Devonian Fisset Brook Formation within the shear zone along western Cape Breton Island. The geochemical signature of the mylonite in this area is similar to that of the fresh Fisset Brook Formation basalt. Trends that are recorded in the sheared basalt corresponding to those in the nondeformed vesicular basalt include elevated Ti, P, Zr, Y, and Ga values with lower Ni, Cr, Mg, and Ca values, relative to those reported for basalt from older Ordovician–Silurian units (Lynch, 1996b). Geochemical data suggest that the primary geochemical signature of the mylonite is largely preserved, identifying the mylonite as Fisset Brook Formation basalt. An upper age limit is established by the Tournasian Horton Group conglomerate which unconformably overlies the Margaree shear zone; the age of the shear zone is thus constrained to the interval 375–360 Ma.

The Margaree shear zone juxtaposes underlying rocks which were metamorphosed to relatively high-grade conditions at or before 376 Ma, with overlying low-grade to nonmetamorphosed volcanic and sedimentary rocks deposited at approximately 375 Ma, effectively excising a significant portion of the upper crust. Correspondingly, the Margaree shear zone may be classified as an extensional detachment fault which accommodated important crustal thinning during the early evolution of the Maritimes Basin, with high-grade rocks of the Cape Breton Highlands exposed as a metamorphic core complex.

AINSLIE DETACHMENT AND CARBONIFEROUS SALT TECTONICS

Introduction

The Late Devonian to Lower Permian Maritimes Basin of eastern Canada developed as a late- to post-tectonic extensional basin following the Early Devonian Acadian orogenic event, and is subdivided into five main lithostratigraphic groups. The basin was initiated as a response to crustal thinning and extension, in part to accommodate the Margaree shear zone described in the previous section. The Viséan Windsor Group, which is the only marine-dominated interval within the basin, is a thick accumulation of evaporite, carbonate, and siltstone, deposited during passive regional subsidence and intercontinental submersion of a tectonically thinned crust. Earlier rift deposits consist of the nonmarine coarse clastic sediments of the underlying Tournaisian Horton Group; the Horton to Windsor succession defines a typical rift and sag transgressive event. The Macumber and stratigraphically equivalent Gays River formations are the marine basal carbonate units to the Windsor Group and contain numerous Pb-Zn-Ba occurrences in central and northern Nova Scotia. These carbonate units are overlain by thick evaporite deposits of the lower Windsor Group, consisting of gypsum, anhydrite, and halite. In northern Nova Scotia the evaporite-carbonate contact at the top of the basal carbonate is the locus of intense shear and brecciation, which developed in relation to the Ainslie detachment. The highly contrasting rheologies of the formations created an anisotropic zone of weakness which acted as an upper crustal stress guide, stratigraphically controlling the trajectory of the detachment through the basin. The detachment is characterized by an approximately 3–10 m thick calc-mylonite zone, with an intense planar fabric featuring alternating very fine-grained shear planes and coarser annealed layers. Recumbent isoclinal intrafolial folds are widespread, and coarser layers are boudinaged into pinch-and-swell structures, locally producing segmented augen. Variably strained intraclasts, ooids and peloids, recrystallized carbonate boudins, and carbonate vein segments are included in the calc-mylonite as semirigid inclusions and rotated porphyroclasts. Thick zones of fault breccia straddle portions of the detachment and overprint the mylonite, demonstrating an evolution to brittle conditions during progressive shear.

Listric faults in the hanging wall of the detachment have both ramp and flat geometry, with an upper detachment occurring along the upper contact of the Windsor Group with the overlying Namurian Mabou Group. Locally up to 2 km of the stratigraphic succession has been removed, with faults cutting downsection in a westerly direction on Cape Breton Island. This detachment is a stratigraphically controlled regional-scale, flat-lying extensional fault that affected the hydrodynamic regime and mineralizing environment in the Carboniferous Maritimes Basin. Movement on the detachment has stripped away thick evaporitic units and excised the entire Windsor Group above the Macumber Formation across wide areas, effectively breaching a regional aquiclude. With shearing, permeability was locally enhanced through brecciation, creating a favourable environment for mineralization. Significant thickness variations within underlying coarse clastic rocks and pinch outs of the Horton Group aquifer also focused basin fluids. As well as requiring a regionally extensive planar low-strength or low-viscosity layer, large-scale detachment faults require high fluid pressures to sustain motion of the allochthonous sheets. Abnormal fluid pressures are known to occur beneath evaporite, which is the site of nucleation for the Ainslie detachment. Fluid focusing along the detachment is suggested from the structural style, and evidence is provided by widespread synkinematic calcite±fluorite-barite-pyrite veins which occur in the calc-mylonite and breccia.

Detachment was triggered during basement normal faulting along the southeast margin of the Gulf of St. Lawrence and western Cape Breton Island. Structural evidence from onshore exposures in the uplifted or horsted regions of mainland Nova Scotia and Cape Breton Island indicate that movement was transferred from basement faults horizontally to the Ainslie detachment. Thin-skinned gravitational sliding produced large rafts of allochthonous strata which progressively draped over the edge of basement normal faults during extension, resulting in tectonic salt buildup in the main offshore graben and a salt diapir emplacement. Compression and shortening at the front end of the raft within the graben produced seismically imaged large-amplitude salt pillows and thrusts above the detachment. Pre-, syn-, and postkinematic sedimentation bracket the age of the detachment to Late Carboniferous Westphalian C–D.

New 1:250 000 compilation maps for northern Nova Scotia and southeastern New Brunswick (Lynch et al., 1998), along with the fieldwork that was required to contribute to the mapping, provide the basis for highlighting the regional importance of the Ainslie detachment in the evolution of the Maritimes Basin (Lynch and Giles, 1996). Furthermore, this structure has been clearly imaged in the offshore area of the Gulf of St. Lawrence within an extensive body of seismic data (Lynch and Keller, 1998), including LITHOPROBE line 86-1 (Marillier et al., 1989), allowing for a closer integration of onshore and offshore data sets. The Maritimes Basin (Roliff, 1962; Howie and Barss, 1975) incorporates a widely distributed succession of Upper Devonian to Carboniferous stratified units in Atlantic Canada, centered in the Gulf of St. Lawrence and onlapping large areas of the eastern seaboard (Fig. 1). The Maritimes Basin measures roughly 900 km long by 400 km wide, with accumulated strata reaching a thickness of up to 12 km in the offshore depocenter (Marillier et al., 1989). The stratigraphy of the Maritimes Basin is well defined (Fig. 16, 21) and is dominated by nonmarine clastic rocks; however, a thick interval of marine evaporite, carbonate, and siltstone, which comprise the Viséan Windsor Group, is found in the middle of the stratigraphic succession. Thick evaporite units occur in the lower Windsor Group, and the development of a broad diapir field in the Gulf of St. Lawrence (Howie, 1988) is a distinctive feature of the Maritimes Basin (Fig. 22). Characterization of a regional detachment fault along a stratigraphic level at the base of the evaporite and diapirs (Fig. 23), is the subject of this section. The Macumber Formation and laterally equivalent Gays River Formation together form the basal carbonate units of the Windsor Group; these units mark an abrupt marine transgression in the Maritimes Basin (Giles, 1981). The Macumber and Gays River formations host several significant Pb-Zn-Ba(±Cu, Ag) occurrences, including the Walton, Smithfield, Jubilee, and Gays River deposits (Kirkham, 1974).

Evaporitic basins that have developed extensive diapiric structures may contain important hydrocarbon reserves by virtue of their distinct structural style enhancing the development of traps, and due to their typical association with favourable source rock and/or reservoir lithologies. Although evaporite deposits occur worldwide, those which contain important diapir fields were developed mainly in the Mesozoic and Tertiary along the Atlantic and Mediterranean–Red Sea rift systems (e.g. Burke, 1975; Jackson and Talbot, 1994; Jackson and Vendeville, 1994). In well stratified basins evaporite layers impart a pronounced anisotropic weakness to the sedimentary successions and influence the propagation of faults, resulting in large-scale detachment along the evaporite layers (Kehle, 1970; Gibbs, 1984). In the case of Mesozoic Atlantic continental margins, subsidence creating broad low-angle slopes resulted in gravity slides or spreads, salt buildup, and diapir formation at the front end of large allochtonous sheets (Worrall and Snelson, 1989). In contrast, evaporite detachments in rifts act as horizontal transfer zones between basement normal faults which occur beneath the detachments, and thin-skinned listric normal faults in the horsted blocks above the detachments (Gibbs, 1984; Nalpas and Brun, 1993; Gaullier et al., 1993); consequent salt buildup and diapir emplacement are more favourable in grabens, whereas horsted blocks may be stripped of their evaporite layer and overlying sedimentary units. Also, through association with drape synclines or forced folds it is increasingly recognized that evaporite detachments are linked to steep basement normal faults (Withjack et al., 1990; Jackson and Vendeville, 1994), which generate gravitational instabilities and likely trigger detachment.

Although a vast body of seismic data from numerous basins worldwide provide convincing evidence that detachment faults are linked to evaporite horizons and are in many cases associated with diapir formation, very few extensional salt detachments are actually exposed and remain as geophysical images only. The Maritimes Basin, on the other hand, advances our understanding of such phenomena because the

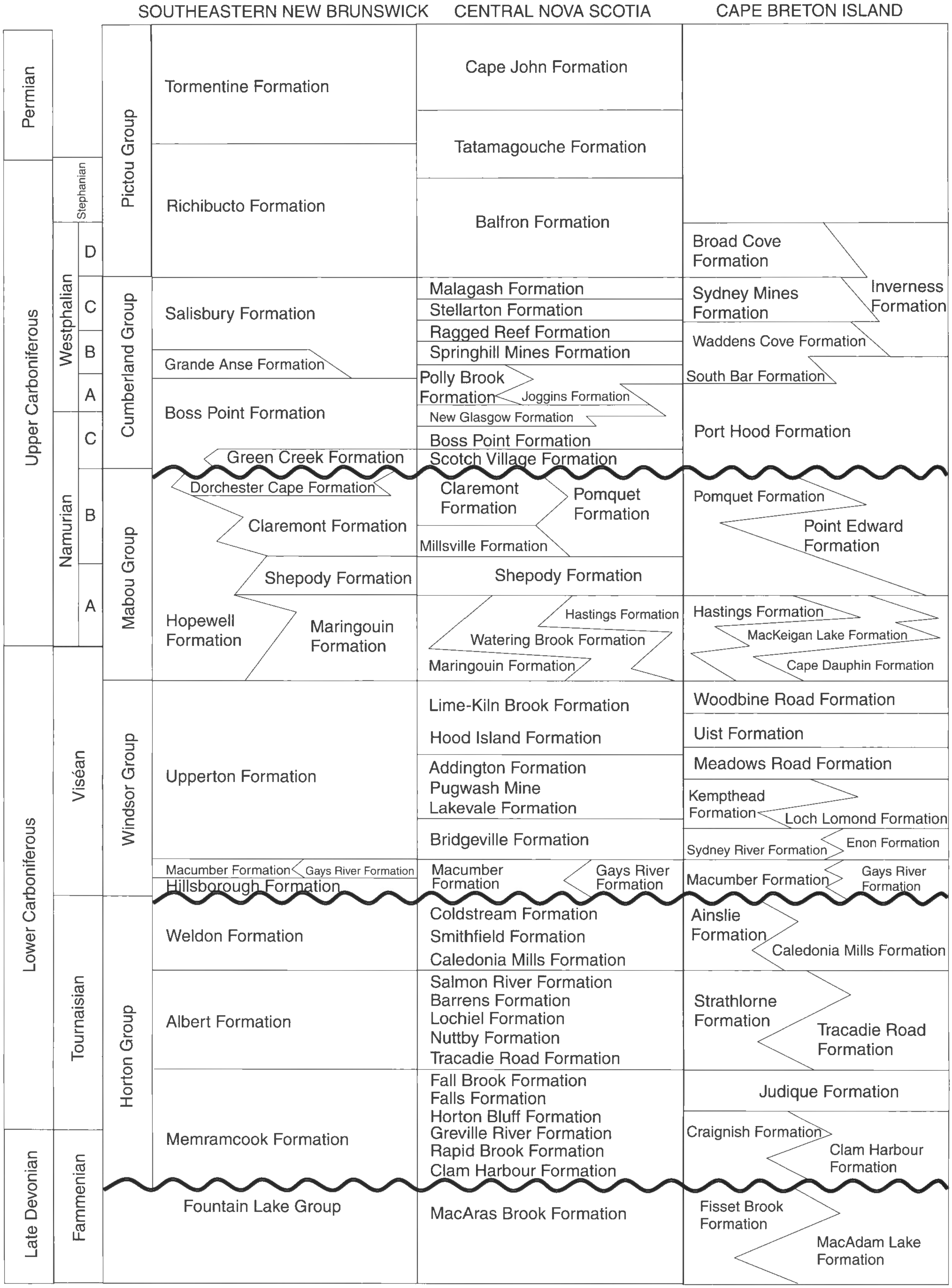


Figure 21. Stratigraphic column of prominent formations from Maritimes Basin in southeastern New Brunswick, central Nova Scotia, and Cape Breton Island. Description of units contained in legend of compilation maps from Lynch et al. (1998).

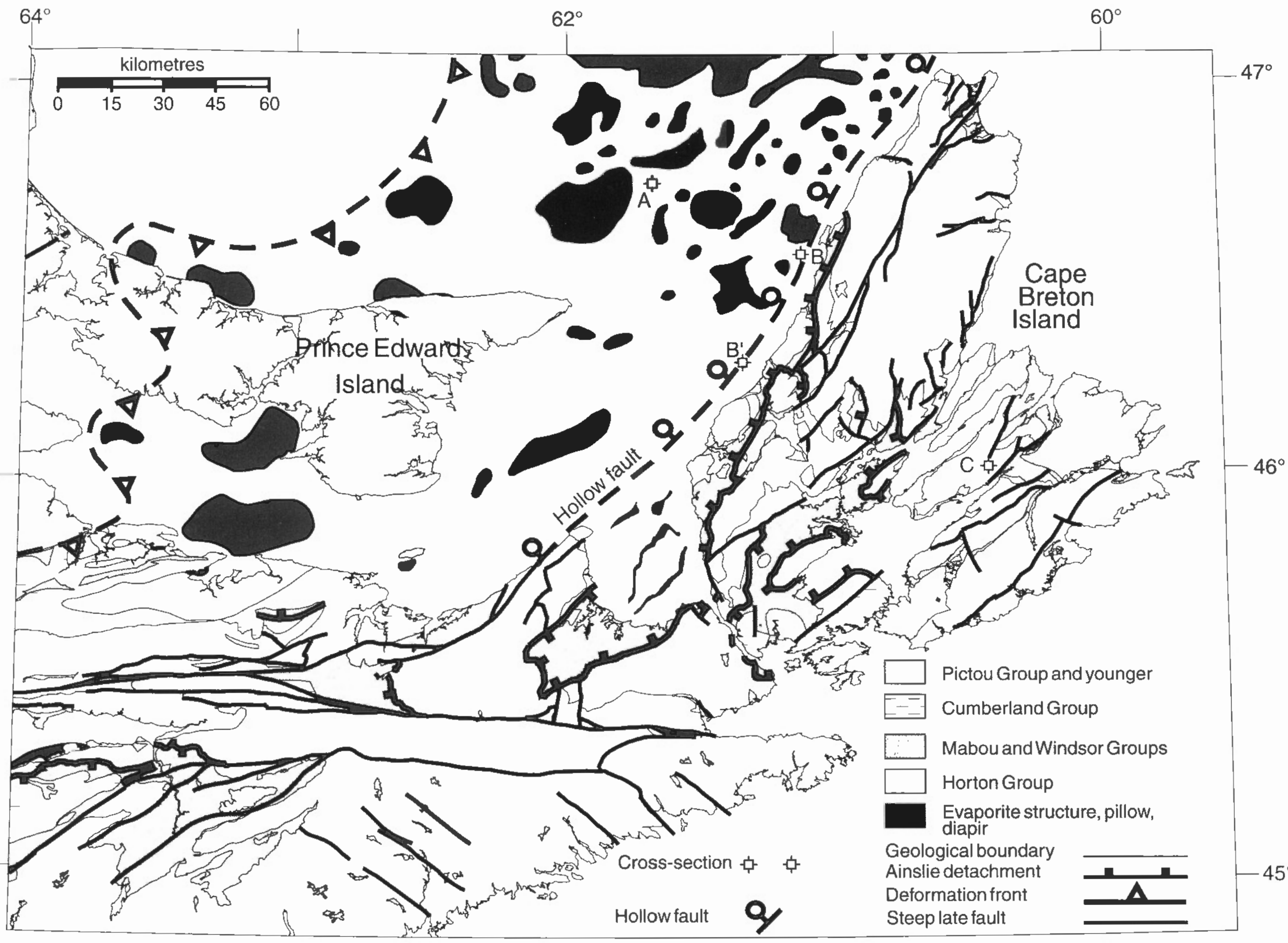


Figure 22. *Map of diapir field in Gulf of St. Lawrence, onland trace of Ainslie detachment, and approximate offshore deformation front at stratigraphic level of the Ainslie detachment, as well as trace of Hollow fault. Cross-section A–B B´C shown in Figure 23.*

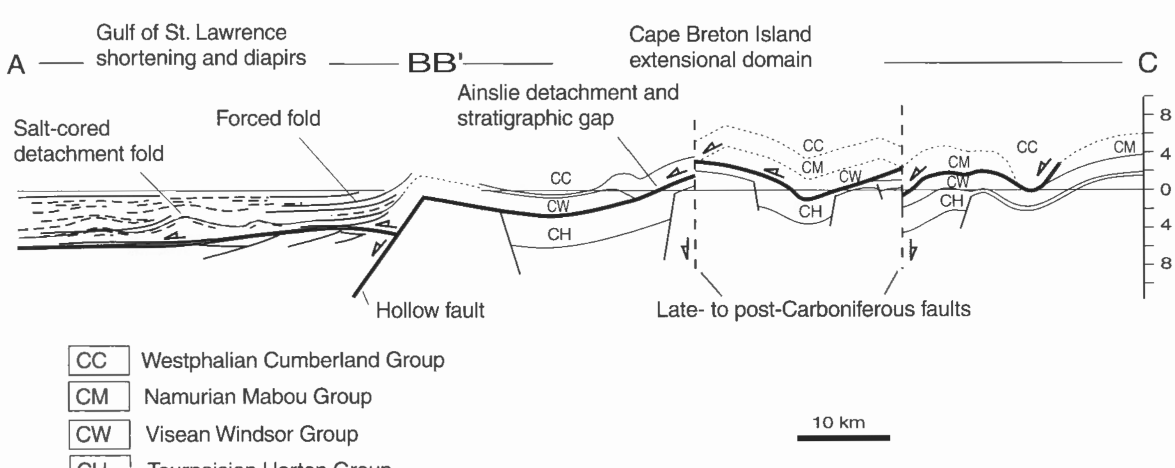


Figure 23. *Cross-section highlighting onshore and offshore features of Ainslie detachment, as well as kinematic relay with Hollow fault and associated drape folding (see text for discussion, and location of section on Fig. 22)* (adapted from *Lynch and Keller (1998)).*

detachment is exposed, and through linking onshore fieldwork in Nova Scotia to offshore seismic data from the Gulf of St. Lawrence, a relationship is demonstrated between basement faulting, detachment faulting, and diapirism.

The basal carbonate units of the Windsor Group are capped by regionally extensive evaporite-dominated deposits. The thick evaporite units of the lower Windsor Group played a key role in influencing basin deformation (Boehner, 1992) and fluid migration (Ravenhurst and Zentilli, 1987). Generally, evaporite deposits form anisotropic low-strength layers at upper crustal levels, which in some regions control the propagation of bedding-parallel detachment faults (Heard and Rubey, 1966; Lehner, 1969; Kehle, 1970; Bishop, 1973; Demercian et al., 1993; Lynch and Tremblay, 1994; Lynch and Giles, 1996). Such detachments horizontally relay displacement between overlying near-surface listric fault systems and underlying high-angle basement normal faults of the mid-crust (Gibbs, 1984; Nalpas and Brun, 1993; Gaullier et al., 1993). Although assisted by the low viscosity of evaporite deposits (Kehle, 1970), large, flat faults require high-fluid pressures for the allochthonous transport of intact detached sheets (Hubert and Rubey, 1959; Crans et al., 1980). Drilling has provided direct evidence of high fluid pressure buildup, immediately beneath evaporite deposits (Lane, 1949; Rubey and Hubert, 1959; Hubert and Rubey, 1959; Dickey et al., 1968). Consequently, the lower contact of evaporite deposits, rather than the upper contact, is usually the preferred site for detachment. Subhorizontal slip planes may be difficult to recognize, either in wells, seismic lines, or during the course of mapping. Strata above the detachments are transported predominantly without being deformed as stratigraphically contiguous rafts, except for the front and back ends of the sheets where compressional buttressing and extensional deformation occur, respectively (Kehle, 1970); however, local salt flow into anticlinal hinges where rafts move over inherited basement structures may progress to diapirism, halokinesis (Talbot, 1978), and penetration of the overlying stratigraphic section (Nalpas and Brun, 1993; Gaullier et al., 1993). Detachment faulting may have local, as well as regional, effects on the circulation of basinal brines and mineralizing fluids; the creation of breccia in the immediate vicinity of the fault enhances permeability, whereas detachment-related disruptions in the stratigraphy may alter the large-scale pattern of fluid migration.

Typically, a genetic link is inferred between carbonate-hosted Pb-Zn deposits and large-scale fluid migration. Ravenhurst and Zentilli (1987), based on geochemical and numerical modelling arguments, proposed a regional-scale fluid migration model for the Maritimes Basin which uses the evaporite deposits at the base of the Windsor Group as an impermeable barrier. This impermeable unit focused fluids along the underlying carbonate of the Macumber and Gays River formations where mineralization occurs. Moreover, a tectonic hydrofracturing mechanism in relation to overpressured fluids was proposed for the origin of some of the mineralized breccia within the Macumber Formation (Ravenhurst and Zentilli, 1987). Although originally flat lying, the Ainslie detachment is now well exposed due to late- to post-Carboniferous folding and brittle reverse faulting which has isolated segments of the Maritimes Basin into erosional subbasins (Fig. 24). Transpression and thrusting documented in some areas have affected the basin after its formation (St. Jean et al., 1993), or are reported as a basin-forming process linked to sedimentation (Rast and Grant, 1973; Ruitenberg and McCutcheon, 1982; Nance, 1987; Hyde et al., 1988).

Stratigraphy of the Maritimes Basin

Five main groups define the Maritimes Basin in Nova Scotia, and range in age from Late Devonian to Permian. From oldest to youngest these are the Horton, Windsor, Mabou, Cumberland, and Pictou groups (Ryan et al., 1991). Volcanic rocks of the Late Devonian Fisset Brook Formation locally underlie the sedimentary succession on Cape Breton Island. Four of the five groups are dominated by nonmarine clastic rocks, whereas the intervening Windsor Group is dominated by marine carbonate, evaporite, and siltstone, which marks a regional transgression in the Viséan (Giles, 1981). A listing of the dominant formations is provided in the stratigraphic column of Figure 21 derived from the 1:250 000 scale compilation maps contained in Lynch et al. (1998), and is divided into three regions highlighting corresponding stratigraphic levels.

Although the five-group stratigraphic framework is well established for the Carboniferous of Nova Scotia, tectonic activity and structures have locally affected the nature of the Windsor–Mabou transition. Stratigraphic relationships which appear to represent the normal succession are now known in some areas to be the result of tectonic juxtaposition due to movements on a large-scale detachment fault and associated listric normal faults (Lynch and Giles, 1996). The effects of these fault systems on the lowermost part of the Windsor Group is significant, not only in their tectonic modifications to the overall stratigraphic sequence, but for their physical modifications of carbonate rocks hosting mineral occurrences. The stratigraphy is first briefly reviewed below, in order to establish the appropriate context for describing the Ainslie detachment.

Horton Group and Fisset Brook Formation

Bimodal basalt-rhyolite and interbedded siliciclastic rocks of the Fisset Brook Formation and broadly correlative units such as the MacAdam Lake Formation, MacAras Brook Formation, and Fountain Lake Group (Fig. 21) are widely dispersed across the Maritime provinces of eastern Canada, locally forming the basal unit to the Maritimes Basin (Kelley and MacKasey, 1964; Howie and Barss, 1975; Blanchard et al., 1984). The age of the Fisset Brook Formation is constrained by radiometric and paleontological data which indicate a Middle to Late Devonian age (Kelley and MacKasey, 1964; Blanchard et al., 1984; Barr et al., 1995). The volcanic rocks are thought to be the eruptive equivalent to widespread Late Devonian subvolcanic intrusions (Jamieson et al., 1986). Volcanic rock geochemical characteristics suggest a within-plate continental setting for the Fisset Brook Formation (Currie, 1982; Dostal et al., 1983; Blanchard et al., 1984;

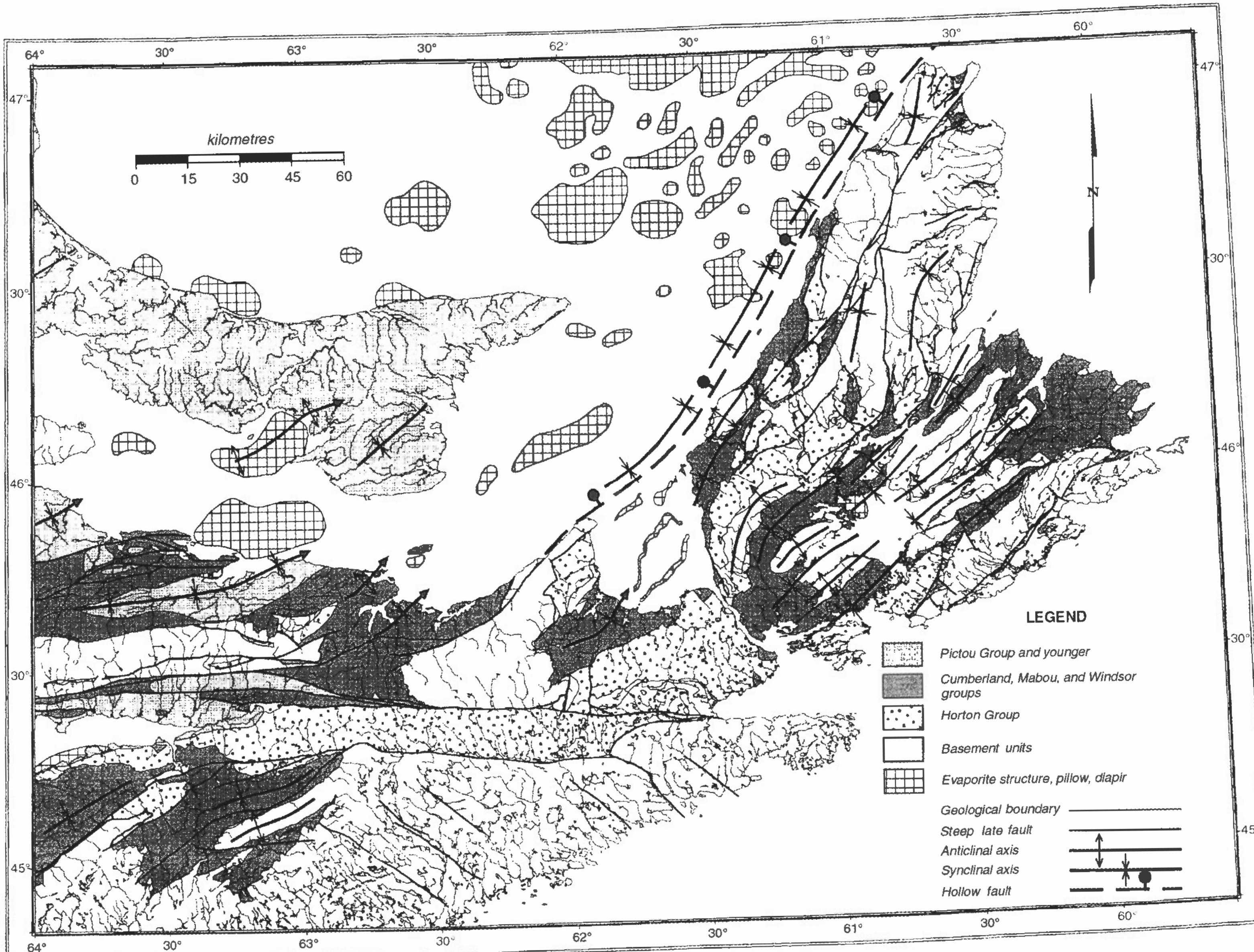


Figure 24. *Distribution of late Carboniferous to post-Carboniferous folds affecting basement and overlying Maritimes Basin stratigraphy. Folds to the northwest of the Hollow fault likely do not affect basement and are possibly detachment folds, diapir structures, and drape folds isolated above the Ainslie detachment.*

Barr et al., 1995). Fresh, dark, highly vesicular basalt is a lithology typical of the Fisset Brook Formation (Fig. 25a), and units are usually well bedded with well defined flow tops and bottoms (Fig. 25a). The depositional environment proposed by Blanchard et al. (1984) is one of early explosive volcanism and clastic alluvial fan-type deposits, followed by basaltic eruptions with coeval fluvial and lacustrine sedimentation. Deposition apparently occurred within a horst-and-graben setting. Although a thickness of up to approximately 500 m is reported for the Fisset Brook Formation (Blanchard et al., 1984), the presence of intraformational unconformities suggest that the original thickness is not preserved (Lynch, 1996a).

The Horton Group is dominated by Late Devonian (Famennian) to Carboniferous (Tournaisian) clastic rocks (Hamblin and Rust, 1989; St. Peter, 1993). Conglomerate rests unconformably on the Fisset Brook Formation locally, or more typically on basement rocks which range from Precambrian to Devonian, overlapping all of the major lithotectonic domains of the Canadian Appalachian Mountains. Clasts within basal Horton Group conglomerate typically reflect local provenances.

The divisions for the Horton Group of Murray (1960) have been widely adopted in northeastern Nova Scotia and Cape Breton Island. Murray (1960) recognized a lower Craignish Formation, a middle Strathlorne Formation, and an upper Ainslie Formation; however, Kelley (1967) noted lateral facies changes and interfingering of the Strathlorne and Ainslie formations. Note, many more subdivisions are defined for the Horton Group in central Nova Scotia (Fig. 21), while a simpler tripartite division of the Horton Group is proposed for southeastern New Brunswick, where units include, in ascending order, the Memramcook, Albert, and Weldon formations (Fig. 21). The depositional environments for the Horton Group on Cape Breton Island have been interpreted by Hamblin and Rust (1989): the Craignish Formation features coarse to fine braidplain sediments near fault-bounded alluvial fans as well as distal mudflat and/or playa deposits, and may include well bedded successions of poorly sorted coarse and fine debris-flow deposits (Fig. 25c, d); the Strathlorne Formation comprises lacustrine, shoreline, and fan-delta facies; whereas the overlying Ainslie Formation is described as an alluvial fan, fluvial, and floodplain system. Facies within the Craignish and Ainslie formations have close similarities, although the Ainslie Formation is generally finer grained. The Strathlorne Formation is distinguished by the presence of abundant grey mudstone, minor redbeds, and locally, conglomerate and rare limestone. The limestone includes thin micritic beds of uncertain origin as well as oolitic rocks in beds 10–20 cm thick. In addition, algal bioherms up to 30 m in thickness (typically 1–2 m) occur locally. Intraformational conglomerate occasionally contains micritic limestone clasts and redbed clasts indicating erosion and cannibalization within the unit.

The thickness of the Horton Group varies greatly; sedimentation is interpreted to have been controlled by a horst-and-graben architecture that formed linear troughs (Hamblin and Rust, 1989; Durling and Marillier, 1990, 1993, 1994). Locally the Horton Group changes from approximately 500 m to as much as 3–5 km in thickness across horst-bounding growth faults. Late folding (Fig. 25e) and steep to flat thrusts (Fig. 25f) overprint the Horton Group.

Windsor Group

The Windsor Group is Viséan and immediately overlies the Tournaisian Horton Group. A comprehensive review of this unit is provided in Giles (1981). The Windsor Group is unique in eastern Canada, representing the sole record of unequivocally marine sedimentation from the Carboniferous. It provides a well documented reference unit in the Maritimes Basin, and is distinct by virtue of its marine carbonate and evaporitic rocks. The carbonate at the base of the Windsor Group has been the focus of extensive base-metal exploration because of known occurrences such as the Gays River Pb-Zn mine in central Nova Scotia, and because of the important Irish carbonate-hosted Pb-Zn metallogenic province which occurs in a similar, if not directly correlative stratigraphic setting (Kelley, 1961; Kirkham, 1974, 1978; Binney and Kirkham, 1975). The relationship with the underlying Horton Group ranges from angular unconformity to disconformity, between the late Tournaisian and early Viséan (Utting et al., 1989).

The Windsor Group ranges in thickness from as little as a few hundred metres, to 2.5–3.0 km in northeastern Nova Scotia and Cape Breton Island. Numerous formations have been defined within the Windsor Group (Fig. 21). For practical purposes the Windsor Group may be divided into lower and upper units, which correspond largely to the "Lower Windsor Zone" and "Upper Windsor Zone" of Bell (1929), and with the lithostratigraphically defined "Upper Windsor" of Moore (1967). The base of the Windsor Group in Nova Scotia is placed at the lower contact of the laterally equivalent Macumber and Gays River formations. The Macumber Formation is the most conspicuous marker unit within the group, comprising regionally uniform laminated peloidal and oolitic limestone which can be traced throughout the Maritimes Basin (Schenk, 1967). Typically, the Macumber Formation is about 10 m thick, but varies from approximately 3 m to more than 25 m thick. The Gays River Formation comprises a biohermal facies of the basal Windsor Group. It was deposited where underlying Horton Group strata were overstepped adjacent to pre-Carboniferous upland areas on the basin periphery (Giles et al., 1979). The Gays River biohermal facies is usually massive with a significant amount of growth porosity. It hosts a deposit of Pb-Zn at the type locality of Gays River. Like the Macumber Formation, the Gays River Formation is overlain by thick evaporite-dominated successions, although in basin-margin areas, marine shale with thin intercalated gypsum may interfinger with anhydrite.

The lower Windsor features a single 600–700 m thick transgressive-regressive (Fig. 2) evaporitic cycle with the Macumber limestone (or equivalents) at its base, passing upwards to thick anhydrite, and ultimately to halite with varying proportions of siltstone near the top (Giles, 1981; Boehner, 1986). Overlying this lower cycle, similar but thinner transgressive-regressive cycles occur repeatedly throughout the upper Windsor Group (Giles, 1981). The recognition of

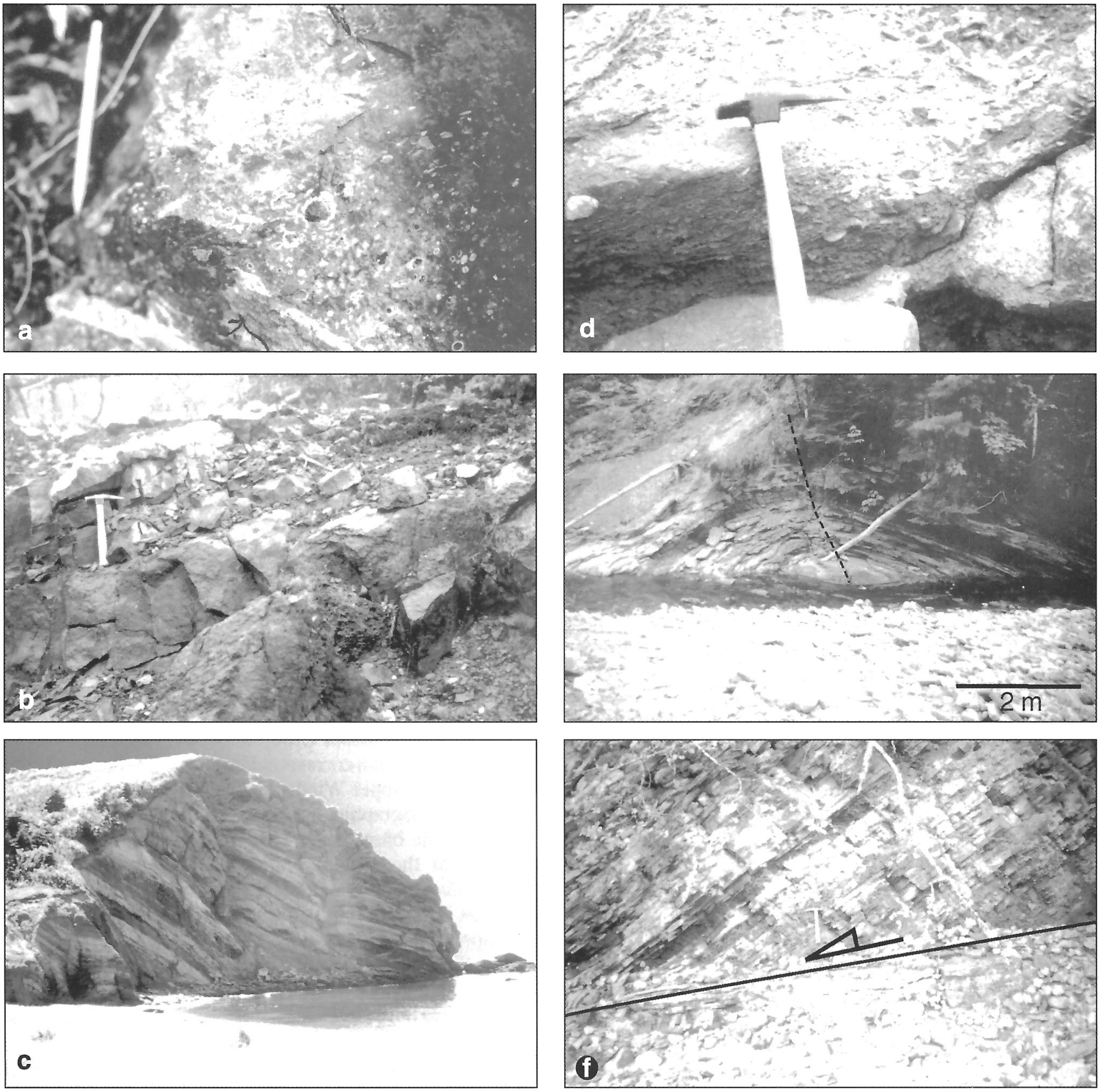


*Figure 25. **a)** Vesicular basalt from the Fisset Brook Formation (photograph by G. Lynch; GSC 2000-041T); **b)** shallow-dipping flow bedding in basalt of the Fisset Brook Formation (photograph by G. Lynch; GSC 2000-041U); **c)** well bedded proximal alluvial fan deposits of the Horton Group including conglomerate, breccia, and sandstone, north of Chéticamp, face of outcrop approximately 12 m high (photograph by G. Lynch; GSC 2000-041V); **d)** poorly sorted debris-flow breccia in mud and silt matrix from site in photograph c (photograph by G. Lynch; GSC 2000-041W); **e)** late upright folds overprinting siltstone and shale from Horton Group (Strathlorne Formation)(photograph by G. Lynch; GSC 2000-041X); **f)** late thrust in Horton Group with cut-off angles between bedding and fault indicating a ramp and flat geometry (photograph by G. Lynch; GSC 2000-041Y).*

these cycles relies on the presence of discrete carbonate marker units that can be correlated over broad regions of central and eastern Nova Scotia in the middle and upper Windsor Group. The proportion of continental siliciclastic rocks increases significantly towards the top of the Windsor Group where red and grey siltstone is the dominant rock type (Moore, 1967; Giles, 1981). Nonetheless, eight regionally distributed marine carbonate marker units can be recognized within the upper Windsor Group (Moore, 1967; Giles and Boehner, 1982).

Mabou, Cumberland, and Pictou groups

In ascending order the Mabou, Cumberland, and Pictou groups comprise the upper stratigraphic units of the Maritimes Basin on the mainland of Nova Scotia and New Brunswick, and mark a return to predominantly nonmarine conditions; however, stromatolites in the lower Mabou Group and regional facies analysis of the Cumberland Group (Rehill, 1996) suggest the presence of some marine incursions. A listing of the principal formations within these three groups is provided in Figure 21. The Pictou Group is nearly absent from Cape Breton Island.

The contact between the upper Windsor Group and the overlying Mabou Group is conformable, with the lower beds of the Mabou Group featuring grey shale and gypsum, as well as stromatolitic and oolitic carbonate rocks assigned to the Hastings Formation (Belt, 1965). Red siltstone and ripple crosslaminated sandstone dominate the Pomquet Formation of the upper Mabou Group (Belt, 1964, 1965). Thickness estimates for the group demonstrate a wide range, from 3000 m (Belt, 1965) to more than 6000 m in the Antigonish Basin of central Nova Scotia (Boehner and Giles, 1982). The age of the Mabou Group ranges from latest Viséan to Namurian (Neves and Belt, 1970; Utting, 1987), with the Viséan– Namurian boundary located near the top of the Hastings Formation on Cape Breton Island (lower Mabou Group).

Locally stratigraphic relationships between the Windsor and Mabou groups reveal a complexity which until recently had not been suspected. Significant portions of the Windsor Group succession are absent in broad areas of northeastern Nova Scotia and in central and western Cape Breton Island. Previously attributed to disconformable or unconformable relationships between the Mabou and Windsor groups, these relationships are now considered to be of tectonic origin (Lynch and Giles, 1996). Detachment of the succession overlying the Macumber Formation has resulted in down-to-the-basin translation of large rafts of younger strata. Where the detachment is represented by former ramps, younger strata ranging from middle Windsor to Mabou Group may be in tectonic contact with the Macumber Formation (Lynch and Giles, 1996).

The basal Cumberland Group (Ryan et al., 1991) unconformably overlies upper portions of the Mabou Group where thick arkosic or quartz-arenite channel sand bodies (e.g. Port Hood, Boss Point, and Scotch Village formations), as well as red and grey shale with associated coal seams are found. Palynological data suggest that the lower Cumberland Group in western Cape Breton Island is Late Namurian to earliest Westphalian A (P. Giles, pers. comm., 1998). The upper part of the Cumberland Group in western Cape Breton Island and northern Nova Scotia contains arkosic sandstone and pebble conglomerate with interstratified coal and shale. Palynological determinations had previously indicated a Westphalian C–D age for the type Inverness Formation (Hacquebard et al., 1989). Newly acquired data suggest that the formation may range in age from Westphalian B to Stephanian. The upper part of the Cumberland Group in western Cape Breton oversteps all older Carboniferous beds to lie directly on pre-Carboniferous rocks indicating the presence of a major unconformity at that level.

The Pictou Group comprises the uppermost part of the Carboniferous fill of the Maritimes Basin in central Nova Scotia and southeastern New Brunswick. Major rock types include red siltstone and mudstone, red crossbedded arkosic sandstone and pebbly sandstone, as well as intraformational mud-clast conglomerate. In western Cape Breton Island, the Broad Cove Formation is the only unit from the Pictou Group on the island, and is concordant or apparently conformable with the underlying Inverness Formation (Cumberland Group). This contrasts with north-central Nova Scotia where the Pictou Group contains five formations, two of which continue into southeastern New Brunswick (Richibucto and Tormentine formations) (Fig. 21).

Early basin subsidence

The Late Devonian to Early Carboniferous evolution of the Maritimes Basin is characterized by intense rifting, suggested by intraplate volcanism (Fisset Brook Formation and Fountain Lake Group) and coarse fault-controlled clastic sedimentation (Horton Group). This was followed by passive subsidence and regional marine flooding with accompanying evaporite and carbonate deposition within a restricted intercontinental basin (Windsor Group). The succession depicts a typical rift and sag progression (e.g. McKenzie, 1978). Late Devonian crustal-scale thinning and extension has been proposed for the early evolution of the Maritimes Basin (McCutcheon and Robinson, 1987; Marillier and Verhoef, 1989; Lynch and Tremblay, 1994). Crustal extension, accommodated in part by shallow-dipping brittle-ductile shear zones such as the Margaree shear zone described previously, in western Cape Breton Island, immediately followed the Acadian orogenic event (Lynch and Tremblay, 1994). Two periods of extension characterized by low-angle extensional faults or detachments are recognized, namely the Margaree shear zone, and the Ainslie detachment, both of which are exposed in northwestern Nova Scotia (Lynch and Tremblay, 1994; Lynch and Giles, 1996). Field evidence demonstrates that the Margaree shear zone was active at the onset of extension in the Late Devonian, and affected basement rocks and the early basin fill; however, the Ainslie detachment was only active during the Late Carboniferous (Namurian and/or Westphalian), controlling sedimentation, producing shear along the upper contact of the Macumber

Formation, and creating stratigraphic gaps within the Windsor Group (Lynch and Giles, 1996). Dextral movement in the region on steeply dipping fault systems, such as the Cobequid–Chedabucto and Hollow faults (Fig. 3), is constrained from the detailed work of Yeo and Gao (1986) and Yeo and Ruixiang (1987), to between Westphalian B and Stephanian time, apparently postdating movement on the Ainslie detachment. Structural analysis in the vicinity of the Hollow fault indicates that compression and folding were likely related to this late strike-slip event (St. Jean et al., 1993); however, the basin configuration, sediment distribution, and drape folding demonstrate an early component of normal movement along the Hollow fault, at least for the segment which bounds western Cape Breton Island and the thick accumulation of sediments along the southwestern Gulf of St. Lawrence (Fig. 22, 23).

Uplift-subsidence curves generated for extension in the continental crust (Lister et al., 1991) can be used to model the rift and sag transgression recorded in the Horton Group to Windsor Group succession. Subareal sedimentation was maintained during early rifting because the crust was buoyed by a high geothermal gradient maintained during extension. Thermal decay, following the end of rifting and crustal thinning, was succeeded by widespread passive subsidence and marine flooding. Magmatic underplating is believed to have accompanied extension in the Maritimes Basin, producing a regional gravity anomaly in the Gulf of St. Lawrence (Marillier and Verhoef, 1989) (Fig. 14), and may have resulted in widespread volcanism at surface (e.g. Fisset Brook Formation and Fountain Lake Group).

Ainslie detachment

The Ainslie detachment was first defined and described by Lynch and Tremblay (1994) and Lynch and Giles (1996), from mapping campaigns conducted in western Cape Breton Island; however, this section expands on the definition of and extent of the Ainslie detachment, by integrating evidence for the detachment in both onshore and offshore regions. On mainland Nova Scotia and Cape Breton Island, the Ainslie detachment is defined by an approximately 5–10 m thick zone of breccia, calc-mylonite, and deformed limestone and evaporite, occurring along the upper contact of the Macumber Formation, across which significant gaps in the stratigraphy of the Maritimes Basin are locally recorded (Lynch and Giles, 1996). The position of the detachment is in large part coincident with that of the Pembroke Breccia, of which some portions may be interpreted as having originated during brittle faulting (e.g. Lavoie, 1994; Lavoie and Sangster, 1995). Calc-mylonite is characterized by a strong laminated tectonic fabric, featuring thin recrystallized layers bound by very fine-grained laminae. The calc-mylonite is variably porphyroclastic, and ranges from protomylonite to mylonite according to the classification scheme of Sibson (1977). Lineations are typically absent or only poorly developed. The Ainslie detachment coincides with the stratigraphic level from which offshore salt structures have been detached, in which case the extent of the Ainslie detachment can be expanded to include the entire diapir field centered in the Gulf of St. Lawrence (Howie, 1988).

Stratigraphic omission recorded above the detachment in the onshore segments characterizes the structure as an extensional fault in these areas. Most noticeably, the Mabou Group has been downdropped, and is juxtaposed directly against the Macumber Formation across a zone of bedding-parallel calc-mylonite or fault breccia. Up to 2 km of the Windsor Group is missing in some areas.

The close stratigraphic control allows regional delineation of the Ainslie detachment. The onshore trace encompasses an area of approximately 10 000 km^2, across a distance of 150 km from western Cape Breton Island onto the mainland (Fig. 22). The detachment is variably folded by younger deformation events, and consequently outcrops well in different regions; however, the full original extent is not known because of erosion. In western Cape Breton Island the Ainslie detachment is unconformably overlain by Westphalian C clastic rocks of the Cumberland Group, which overlap basement and Windsor Group rocks (Lynch and Giles, 1996). The age of the detachment is thus constrained by field relationships to a period in the Carboniferous between the older Namurian Mabou Group which is sheared along the detachment, and the younger Westphalian C of the Cumberland Group. Evidence from seismic data in the offshore areas also points to a late Namurian to early Westphalian age for the detachment (Lynch and Keller, 1998).

Textures and fabrics of the Ainslie detachment

Structural features, textures, and fabrics of the Ainslie detachment are illustrated in Figures 26 and 27. The fault zone encompassing the Ainslie detachment varies in thickness from 3 m to 10 m, and shows a layered internal structure comprising different fault-rock types. The fault rock types consist of planar calc-mylonite, augen mylonite, and calcareous breccia and cataclasite (Fig. 26a, b). Typically, the mylonite is fine grained, and exhibits a bimodal grain-size distribution. Fine calcite grains (10–100 mm) form layers that alternate with coarser (>300 mm) recrystallized layers or boudinaged calcite veins (Fig. 27a, b).

The fault rocks within the Ainslie detachment are characterized by the presence of a strong planar fabric, in the form of a mylonitic foliation or banding (Fig. 26b, c, d). Lineations, however, are typically absent or poorly developed, but where observed consist of calcite fibres, augen ridges, and mineral trails. The orientation of the mylonitic fabric is generally parallel to bedding in the undeformed Macumber Formation limestone in the footwall.

A variety of rigid fragments are included in the mylonite, and different styles of detached folds are also recorded (Fig. 26b, c, d). The augen mylonite is usually associated with the planar calc-mylonite and contains more competent relict carbonate clasts enveloped in mylonite (Fig. 26d, 27d). Commonly, coarser calcite layers or syntectonic veins show signs of shearing and boudinage in the form of pinch-and-swell or augen structures (Fig. 27a, b, c). The mylonitic fabric wraps around elongated porphyroclasts oriented at an angle to the main flow banding suggesting rigid particle rotation during deformation (Fig. 27c). More cataclastic deformation is recognized by the development of coarse breccia toward the top

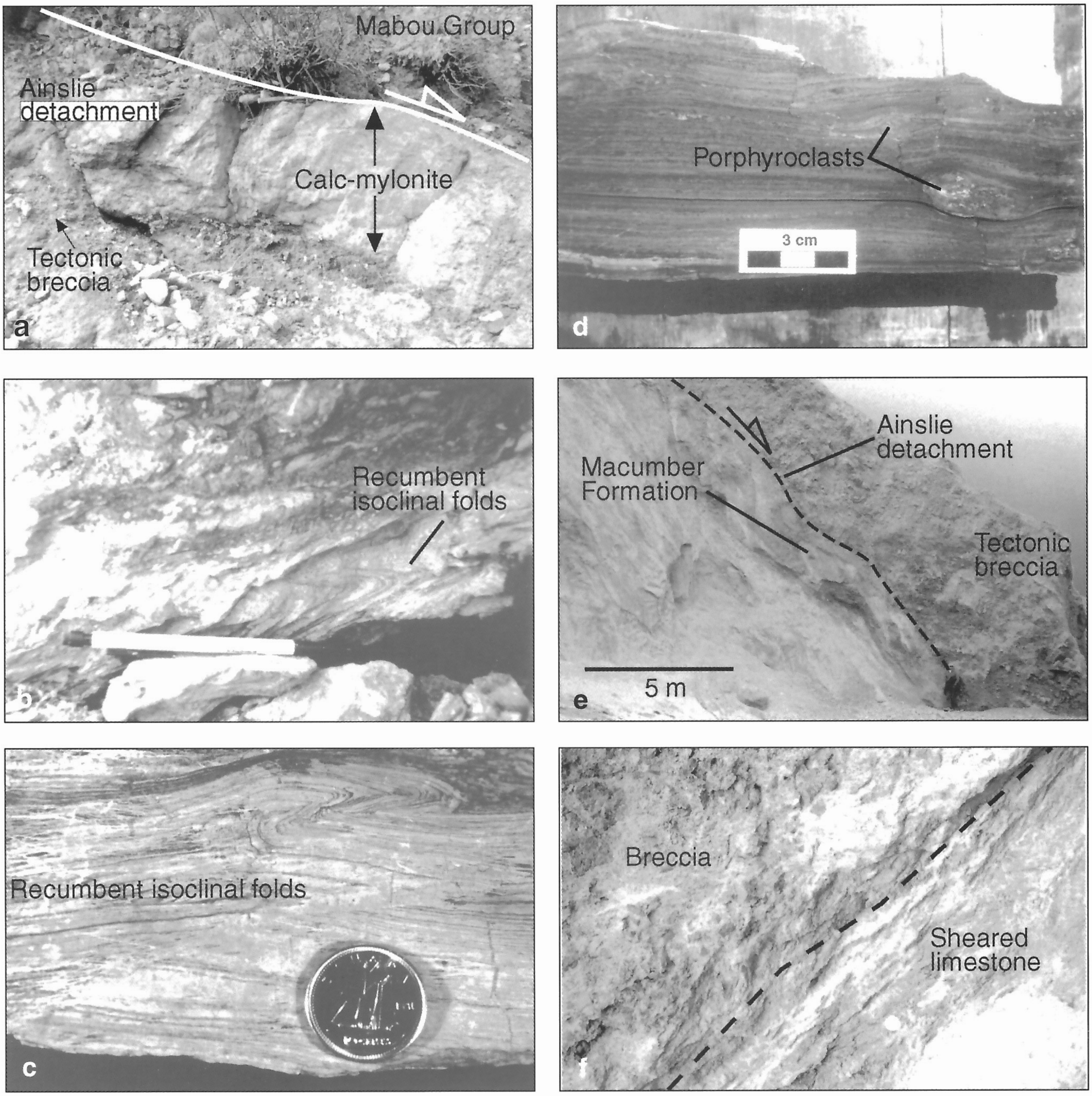


Figure 26. ***a)*** *Line shows upper limit of Ainslie detachment along top of Macumber Formation with zone of calc-mylonite (1 m thick between tips of arrows) containing intrafolial folds as well as tectonic breccia. Rocks of the overlying Mabou Group demonstrate a significant stratigraphic gap (1.5–2 km) due to downramping of units onto the detachment (photograph by G. Lynch; GSC 2000-041Z);* ***b)*** *recumbent isoclinal fold in calc-mylonite of the Ainslie detachment (photograph by G. Lynch; GSC 2000-041AA);* ***c)*** *recumbent isoclinal fold in calc-mylonite of the Ainslie detachment, with dashed line highlighting trace of axial plane (photograph by G. Lynch; GSC 2000-041BB);* ***d)*** *carbonate and anhydrite porphyroclasts enveloped in calc-mylonite of the Ainslie detachment (photograph by G. Lynch; GSC 2000-04CC);* ***e)*** *thick tectonic breccia along the Ainslie detachment at the top of the Macumber Formation, note person for scale at lower right hand of the photograph, breccia is approximately 8 m thick (photograph by G. Lynch; GSC 2000-041DD);* ***f)*** *close-up of tectonic breccia from Figure 26e; dashed line is trace of Ainslie detachment, coin for scale 2.5 cm diameter (photograph by G. Lynch; GSC 2000-041EE)*

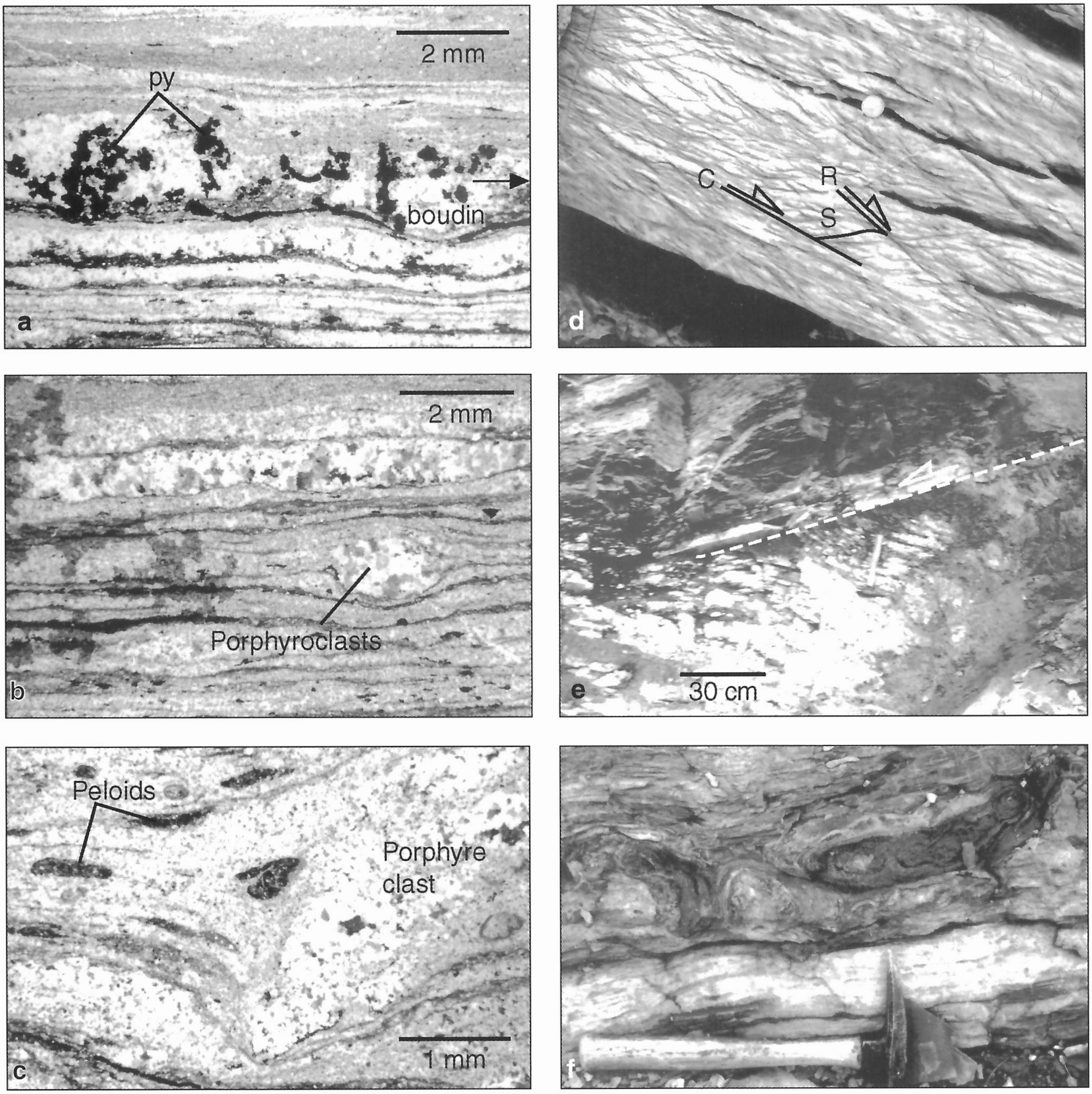


***Figure 27. a)** Photomicrograph of boudinaged recrystallized calcite-pyrite ribbon between very fine-grained horizons of mylonitized carbonate and cataclasite; **b)** photomicrograph of recrystallized calcite ribbons between very fine-grained horizons of mylonitized carbonate and cataclasite, with isolated porphyroclast resulting from extreme boudinage of recrystallized layers; **c)** flattened peloids (dark) in calc-mylonite with rotated porphyroclast (light) demonstrating noncoaxial shear along Ainslie detachment, arrow suggests sense of rotation; **d)** sheared evaporitic rocks (gypsum) directly above Ainslie detachment containing anastomosing C, S, and R shear fabrics, and asymmetrical pattern indicating westward transport, coin for scale 2.5 cm in diameter (photograph by G. Lynch; GSC 2000-041FF); **e)** listric normal fault in shale and dolostone from Mabou Group directly above Ainslie detachment, hammer for scale 30 cm (photograph by G. Lynch; GSC 2000-041GG); **f)** stromatolites in dolostone of Hastings Formation from lower Mabou Group (photograph by G. Lynch; GSC 2000-041HH).*

of the calc-mylonite zone (Fig. 26e, f). The breccia is locally cement supported with highly angular blocky or platy clasts. Voids are common and the breccia may be highly porous. The cement is mainly fibrous and nonfibrous calcite spar, with some barite (Lavoie and Sangster, 1995). Breccia bodies where appropriately exposed have been seen to be as thick as 8–10 m (Fig. 26e).

Beds of gypsum and anhydrite above the Macumber Formation are also deformed, but generally are not well exposed. The evaporite deposits are highly strained and recrystallized; complex ductile flow is indicated by the presence of small-scale sheath folds. Locally Riedel shear structures overprint the principal flattening fabric, and create an anastomosing array with the principal shear planes which may be interpreted to indicate westward transport of the hanging wall (Fig. 27d).

Microstructure

Microscopically, the mylonitic fabric is defined by thin laminations (0.1–1 mm thick) of alternating recrystallized and very fine-grained calcite (Fig. 27a, b). Coarser interstitial carbonate laminae form Monomineralic polycrystalline ribbons. Carbonate ribbons are microscopically boudinaged and segmented producing rigid inclusions or augen within the very fine-grained calc-mylonite (Fig. 27b). Grain boundary migration has locally produced ribbons with serrated grain boundaries. Carbonate veinlets both crosscut the calc-mylonite, and are truncated by it, suggesting synkinematic fluid circulation.

Intraclasts are elongate monocrystalline aggregates that are either asymmetrically or symmetrically winged. Core and mantle textures are observed within the augen; coarser grained cores are rimmed by finer grained calcite. Laminae may contain a penetrative fabric oriented at a moderate angle to the principal mylonitic fabric suggesting local noncoaxial strain. Calcite twinning is observed in coarser grained carbonate layers and in boudinaged intraclasts.

The very fine-grained laminae define the principal shear planes. The shear planes may have a sharp boundary on one side and a progressively coarsening opposite boundary. The sharp boundary contains very fine subgrains which truncate adjacent coarser carbonate laminae. Coarsening develops away from the principal shear planes, possibly as a result of progressive annealing. Recumbent isoclinal folds and intrafolial folds have affected the mylonitic fabric, not only at the outcrop scale, but in thin section as well. Axial planes parallel the mylonitic fabric, and intense layer thickening occurs in fold hinges. Strain is indicated by the occurrence of flattened intraclasts which are preserved locally within the calc-mylonite (Fig. 27c). Peloids and ooids are important components of the Macumber Formation limestone, and their spherical to subspherical shapes in the nondeformed lower portions of the Macumber Formation provide a useful reference frame for the approximation of finite strain in individual grains.

Stylolites overprint and occur approximately parallel to the main mylonitic fabric. The characteristic jagged stylolitic patterns sharply contrast the regular planar habit of the calc-mylonite, and demonstrate late dissolution and solution transfer during pure shear. Dissolution during mylonitization is indicated by the increased abundance of fine detrital quartz along flow planes, and possibly by the presence of biotite and muscovite.

Diapirs and shortening above the Ainslie detachment

Mobilized units above the Ainslie detachment form an allochthonous raft or gravity-slide, with contrasting compressional and extensional domains occurring at the front and back ends, respectively, of the allochthon. Features of the compressional domain occur mainly offshore west of the Hollow fault (Fig. 22), and are made evident by an extensive body of seismic data (e.g. Fig. 23, Marillier et al., 1989; Durling and Marillier, 1990, 1993, 1994; Grant, 1994; Langdon and Hall, 1994; Lynch and Keller, 1998). Salt structures occur along and stratigraphically above the evaporite-carbonate contact at the top of the Macumber Formation, which is the site of the bedding-parallel Ainslie detachment. The main diapir field of the Maritimes Basin is confined to the Magdalen half graben occurring offshore in the Gulf of St. Lawrence. It forms a 575 km long by 150 km wide belt of diapirs, that extends from southwest Newfoundland to the Bay of Fundy, and from the Magdalen Islands in the northwest to Cape Breton Island (Fig. 22; Howie, 1988). The Magdalen half graben is bound by the northeast-trending Hollow–Cabot fault system along the eastern margin of the Gulf of St. Lawrence, with the downdropped block to the northwest. Deep reflection seismic profiles through the Gulf of St. Lawrence (Marillier et al., 1989; Durling and Marillier, 1990) demonstrate that the basal reflector to the Windsor Group defines a gently tapering floor in the downdropped block which abuts on the Hollow–Cabot fault system. Diapirs were delineated in the region using drill hole, seismic, Bouger gravity, and magnetic anomaly data (Howie, 1988; Langdon and Hall, 1994).

Diapirs occur in a variety of forms, but typically are near-vertical straight-sided bodies that pierce all stratigraphic levels above the basal Windsor Group limestone (Howie, 1988; Langdon and Hall, 1994). In plan view the diapirs may be irregular, but elliptical shapes with northeast-trending axes are most common. Individual bodies are as large as 50 km long and 30 km wide. Salt pillows are also common, occurring as domes or doubly plunging anticlines which do not pierce overlying stratigraphy. Particularly clear images of salt pillows above a planar reflector at the base of the Windsor Group have been obtained from the offshore seismic reflection profiles (Marillier et al., 1989; Grant, 1994), with the pillows having amplitudes of approximately 3–4 km creating folds and shortening above the planar detachment (Fig. 23). Progressive dissipation of the anticlinal structures above the pillows into the overlying stratigraphy may be interpreted to indicate overlap in sedimentation and pillow formation (e.g. Jackson and Talbot, 1994); both the Windsor and Mabou groups are folded above the pillows and are considered prekinematic, whereas the Cumberland Group is more gently folded and in part synkinematic, while Pictou Group sediments are flat-lying, postkinematic deposits which postdate pillow formation. Also, long, thin, sharp-crested, salt-cored anticlines have been described by Durling et al. (1995) exhibiting hinges that are up to and greater than 30 km long. The anticlines locally pierce the overlying stratigraphy above their crests and form long, northeast-trending salt walls. Salt-cored buckle folds in the Gulf of St. Lawrence, along the western end of the cross-section (Fig. 4), indicate significant shortening relative to underlying pre-Windsor Group stratigraphy, and was likely greater in the central portion of the diapir field, where diapirs are most abundant.

The contractional domain is also characterized by the presence of salt thrusts and overhangs (Lynch and Keller, 1998). Seismic profiles (Langdon and Hall, 1994) and drill data (Howie, 1988) have outlined the presence of north- northwest-vergent salt thrusts, detached from the base of the Windsor Group at points along the western margin of the diapir field (Fig. 22). Howie (1988), for example, documented thrusting detached from the base of the Windsor Group near the Nova Scotia–New Brunswick border with folding and strongly overturned limbs in the Windsor Group demonstrating northwest transport.

At the graben margin along the Hollow–Cabot fault system, three widely spaced seismic lines (Marillier et al., 1989; Durling and Marillier, 1990) demonstrate that Windsor and Mabou group rocks in the downdropped block are upturned against the fault defining a large northeast-trending drape syncline (Fig. 23), or forced fold (e.g. Withjack et al., 1990). This drape syncline is also detached near the base of the Windsor Group, and forms a large structure with an amplitude of approximately 3 km and wavelength of approximately 5 km, although the anticlinal crest has been eroded and is not imaged. The drape fold suggests a possible kinematic link and fault displacement transfer between the Hollow fault and Ainslie detachment.

Stratigraphic gaps and extension above the Ainslie detachment

East of the Hollow fault, in the onshore areas, deformation which relates to the Ainslie detachment contrasts with that described above for the diapir field west of the Hollow fault. Instead of shortening, thickening, and thrusting of detached units, deformation is characterized by extension, listric normal faulting and the development of significant stratigraphic gaps. Some of these contrasts are illustrated in the southeast-northwest cross-section of Figure 23. In western Cape Breton Island a large klippe of the Mabou Group is isolated on a listric normal fault which downramps through the lower Windsor Group onto the top of the Macumber Formation, and merges with the detachment forming a large-scale ramp (Lynch and Giles, 1996). Exposures along a stream section through the klippe demonstrates that bedding within the lower Mabou Group is parallel to bedding in the underlying Macumber Formation across a parallel zone of calc-mylonite, cataclasite, and breccia. A large open pit mine to the east of the klippe exposes a thick evaporite sequence from the lower Windsor Group, in the footwall of the ramp. The fact that bedding at the base of the Mabou Group juxtaposed with the Ainslie detachment is parallel to the detachment indicates that an upper flat-lying detachment was active beneath the Mabou Group before it was downramped along the listric normal fault onto the detachment. The geometric relationships between bedding and the fault provide evidence for a staircase or flat and ramp geometry, and similar features are observed at the outcrop scale for the Mabou Group in the region (Fig. 27e, f). Furthermore, the shape of the ramp constrains movement of the allochthonous Mabou Group to a general westerly direction.

Along the western coastal section of Cape Breton Island, both Upper Windsor Group and Mabou Group rocks are juxtaposed with the Macumber Formation across the detachment, and bedding attitudes immediately adjacent suggest a high-angle cut-off. The hanging wall of the detachment locally features shallow-dipping overturned beds and westerly verging overturned folds with tight to close interlimb angles. Such folding is not recorded in adjacent footwall units, in which case it is proposed that the folds are related to movement on the detachment. Restoration of the Upper Windsor and Mabou groups from the coastal exposures with the same stratigraphy in the footwall of the listric fault suggests at least 10 km of displacement of hanging wall units towards the west.

At the southern end of Cape Breton Island at Aulds Cove, bedding-parallel calc-mylonite approximately 2–3 m in thickness separates the Macumber Formation from grey shale of the overlying Hastings Formation (lower Mabou Group, Giles and Lynch, 1994; Fig. 26a, b). Recumbent isoclinal folds are intensely developed in the calc-mylonite, and are overprinted by breccia along the lower portions of the mylonite. This locality provides an excellent example of the stratigraphic gap above the detachment.

Throughout southwest Cape Breton Island, the Ainslie detachment is represented by limestone breccia and calc-mylonite similar to that at Aulds Cove. Several sections feature the juxtaposition of the Mabou Group immediately with the detachment above the Macumber Formation, a relationship which is also documented in the subsurface (e.g. Mabou No. 1 well, drilled by Imperial Oil Limited) (Lynch and Giles, 1996).

The Antigonish Basin is preserved as an erosional remnant of a large synclinal structure on the mainland immediately southwest of Cape Breton Island. The Macumber and Gays River formations form the outer rim of the syncline, above basement rocks and discontinuously distributed Horton Group clastic units. The detachment is exposed along the upper contact of the Macumber Formation. Large stratigraphic gaps above the detachment occur along the southeastern edge of the syncline where the Mabou Group occurs immediately above the Macumber Formation. Relatively complete stratigraphic successions, including the Macumber Formation: lower, middle, and upper Windsor Group units; as well as the Mabou Group occur along the western and northern edges of the Antigonish Basin; however, calc-mylonite along the upper contact of the Macumber Formation in these areas suggests that the overlying Windsor and Mabou groups moved as a stratigraphically intact allochtonous package above the detachment, whereas downramping and a roll-over structure accommodated the stratigraphic gap to the southeast. Shear-sense indicators (porphyroclasts) along the northern exposures of the detachment suggest a west to northwest transport direction.

Onland, in northern Nova Scotia on the horsted block east of the Hollow fault, salt diapirs are rare. Instead, large tracts of lower Windsor Group evaporite have been stripped away creating stratigraphic gaps. It is important to note, however, that the gaps are not defined only by missing salt, but rather by the entire Windsor Group above the Macumber Formation, including thick sections of fine siliciclastic rocks and carbonate. Moreover, salt dissolution or halokenesis alone cannot account for the stratigraphic gaps, because the missing units are predominantly not evaporite deposits. Approximately 2 km of stratigraphy is missing locally, where exposures display shale of the lower Mabou Group juxtaposed against the top of the Macumber Formation across a zone of bedding-parallel shear defining the Ainslie detachment (Fig. 26a).

Discussion

Stratigraphic gaps: unconformity, onlap, or tectonic excision?

The identification of major bedding-parallel detachment faults may be hampered by the preservation of complete stratigraphic successions above large segments of the faults. If no obvious stratigraphic omissions are detected, the expression of the fault may reside entirely in its strain signature. Where significant stratigraphic gaps do occur, an unconformity might be implied if a fault is not detected. On the other hand, where stratigraphic units have been downramped, large gaps in the stratigraphic record accompany the strain features which characterize a detachment fault. In Nova Scotia, previous workers have reported that fine-grained sandstone, siltstone, and shale of the lowermost Mabou Group lie unconformably on lowermost beds of Windsor Group, although contact relations were not observed (Norman, 1935; Kelley, 1967); however, in areas where stratigraphic relations are well exposed, sections are described showing that the Mabou Group is both transitional and conformable with uppermost Windsor Group strata (Belt, 1965). The juxtaposition of the Mabou Group against uppermost and lowermost Windsor units within relatively small structural basins indicates that the degree of downcutting required by very localized unconformities would be in excess of 1500 m. Belt (1965) noted that where the basal Mabou Group occurs against the 'ribbon limestone' at the base of the Windsor Group, intense deformation is found. A further argument against a major unconformity at the base of the Mabou Group is the fact that the apparent unconformity systematically penetrates down to the top of the Macumber Formation on a regional scale, but no further; this relationship demonstrates rather that the top of the Macumber Formation is a surface which exercised a fundamental structural control on the present distribution of overlying Windsor and Mabou group rocks. The flat and ramp geometry for the Ainslie detachment bears a close resemblance to that documented for evaporite-controlled extensional detachment faulting in Tuscany (Carmignani et al., 1994).

Strain features and the development of calc-mylonite along the top of the Macumber Formation provide compelling structural evidence for a detachment. A laminated fabric, a pronounced grain size reduction of recrystallized layers, and an abundance of porphyroclasts indicate bedding- parallel shear occurred along the top of the Macumber Formation. The existence of very fine-grained calcite grains during shear may be interpreted as suggesting that grain-boundary sliding was a likely deformation mechanism (e.g. Schmid et al., 1977). The confining pressure for the Ainslie detachment can be inferred from stratigraphic reconstructions. Both the Windsor and Mabou groups were allochthonous in the hanging wall of the Ainslie detachment above the Macumber Formation, for a combined stratigraphic thickness of approximately 4–5 km. This translates to a lithostatic load during faulting of about 1.5 kbar. This estimate is considered to be a minimum since burial beneath Westphalian clastic rocks is not accounted for. A similar crustal level is inferred for nonlineated calc-mylonite developed by grain boundary sliding during superplastic flow along shallow level exposures of the Lochseiten calc-mylonite in Switzerland (Pfiffner, 1982). Furthermore, the tendency of evaporite units in localizing detachment faults is illustrated by the evaporite detachment horizons documented in the Jura Overthrust of Switzerland (Jordan and Rueesch, 1989). Pressure-solution textures and veins provide widespread evidence for fluid-assisted faulting along the Ainslie detachment, which may have resulted in seismic pumping and development of implosion breccia (e.g. Sibson, 1987). Brecciation is interpreted to have resulted in part from progressive shear and extensional unloading of upper units initiating hydrofracturing and brecciation, effectively increasing permeability of the Macumber Formation on a regional scale. Other factors which may also have initiated brecciation include an increase in the strain rate, as well as an increase in pore fluid pressure (e.g. Donath and Fruth, 1971).

Shortening balances extension

Different structural styles, associated with the Ainslie detachment and detached salt structures, are documented in the offshore area of the Gulf of St. Lawrence from that of the onshore exposures in northern Nova Scotia. In the offshore area, a contractional domain occurs, with diapirs, salt pillows, salt-cored buckle folds, and a thrust occurring above the flat-lying detachment, that record net horizontal shortening. The folds define linear sharp-crested structures, or rounded doubly plunging salt pillows which are post-kinematic with regards to early Carboniferous (Viséan–Namurian) units, and prekinematic with regards to Late Carboniferous (Westphalian C–D and younger) units; however, salt diapirs, which may have evolved from large pillows, locally pierce the younger stratigraphy (Howie, 1988). In contrast to this, from the onland region, large stratigraphic gaps above the Ainslie detachment, downramping of younger stratigraphy onto the detachment as well as salt rollers in the footwall of listric normal faults define a domain characterized by net horizontal extension. Both regimes are kinematically linked by the Ainslie detachment, and the differences in structural styles can be reconciled by material balance between the front and back ends of the same allochthonous sheet. The missing stratigraphy in the extensional regime can in part be accounted for by shortening at the front in the compressional regime. This indicates that deformation was thin-skinned and suggests that displacement of the allochthonous sheet was due to gravitational gliding. Draping onto the Hollow fault also accounts for part of the stratigraphic gap after displacement transfer between the basement normal fault and the detachment, since part of the base of the allochthonous sheet covers the new surface area created by the fault. As such the Hollow fault appears to have been largely responsible for generating the gravitational instability, causing tectonic slide along the weak layer of the carbonate-evaporite contact at the top of the Macumber Formation, and also marks the boundary between the extensional domain on the uplifted horsted block and the contractional domain within the downdropped graben.

Carbonate-hosted Pb-Zn deposits in the lower Windsor Group

Because of the important carbonate-hosted Pb-Zn deposits in the Devonian to Carboniferous stratigraphy of Ireland (e.g. McArdle, 1990), it is of metallogenic significance to point out that the Maritimes Basin approximates the rifted counterpart to the Munster Basin in Ireland (Hitzman and Large, 1986) prior to the opening of the Atlantic Ocean. Although there are significant similarities in the stratigraphy of the two regions, the Irish succession is slightly older; coarse, fluvial, clastic rocks of the Old Red Sandstone are typically Devonian, and are similar to the Late Devonian–Tournaisian Horton Group. Carbonate rocks which host the orebodies in Ireland are Tournaisian to Viséan and are locally associated with Late Viséan evaporite deposits, whereas the Windsor Group is Viséan and includes thick evaporite deposits. Both regions also contain extensive Namurian clastic deposits (Rider, 1978). The apparent ages of mineralization in both regions is also similar, within the limits of error, and is approximately constrained to Namurian–Westphalian in the Maritimes Basin (Ravenhurst and Zentilli, 1987; Ravenhurst et al., 1989; Pan et al., 1993; Kontak et al., 1994), whereas it is likely Viséan (Chadian–Arundian; McArdle, 1990) to Westphalian–Stephanian in Ireland. Tectonic parallels can also be drawn; Devonian to Tournaisian normal faulting, locally influenced by inherited basement structures and extensional detachment faults have affected both regions in relation to postorogenic extensional collapse of the Caledonian Orogen in Europe (e.g. McClay et al., 1986) and extensional collapse of the Acadian Orogen in Canada (Lynch and Tremblay, 1994). Younger, flat-lying extensional detachment faults beneath Namurian units are also recorded in both areas (Rider, 1978; Lynch and Giles, 1996). Furthermore, Late- to post-Carboniferous compression and reverse faulting has affected the Munster Basin (Hitzman and Large, 1986) during the Hercynian orogeny (Coller et al., 1986), which is analogous to Alleghenian compression in the Maritimes Basin (Fig. 24). Due to these stratigraphic and structural similarities, common metallogenic features recognized between Irish (Munster) and the Maritimes basins are not surprising.

Recognition of the Ainslie detachment in northern Nova Scotia has important implications for the understanding of some Pb-Zn mineralization in the Windsor Group. The Jubilee Pb-Zn deposit is located in central Cape Breton Island, and is hosted by stratabound to brecciated Macumber Formation limestone. Galena, sphalerite, pyrite, marcasite, chalcopyrite, and calcite occur as void fillings, as veins, as disseminations, or as limestone replacements (Paradis et al., 1993; Fallara et al., 1994). Although the Ainslie detachment occurs to the northwest of the Gays River Pb-Zn deposit which is situated in central Nova Scotia, the age of the Gays River deposit has been estimated by fission track, K-Ar, Ar-Ar dating methods, and paleomagnetic methods to be 330–300 Ma (Ravenhurst et al., 1989; Pan et al., 1993; Kontak et al., 1994). This time interval corresponds approximately with the available age constraints for movement on the detachment (late-Namurian to early Westphalian; Lynch and Giles, 1996). Furthermore, boudinage of calcite-pyrite ribbons within the calc-mylonite, and the occurrence of synkinematic calcite-dolomite-fluorite veins along the detachment provide clear evidence of hydrothermal activity overlapping faulting. Because of the spatial relationship between the detachment and mineralization at the Jubilee deposit, and due to the apparent overlap in timing between the regional mineralizing event and movement on the detachment, it appears likely that the detachment influenced mineralization.

It has been proposed that coarse clastic rocks of the Horton Group acted as the principal lower basin aquifer, and that the thick evaporite succession in the lower Windsor Group provided an effective seal over this aquifer (Ravenhurst and Zentilli, 1987). Abrupt changes in the thickness of the Horton Group and pinch out due to horst-and-graben structures were also significant in controlling fluid flow; however, the Ainslie detachment likely affected the basin hydrology in at least three ways: 1) brecciation along the detachment effectively increased permeability in the Macumber Formation; 2) movement on the detachment may have resulted in hydraulic pumping (Sibson, 1987), creating gradients in fluid pressure resulting in fluid migration, and; 3) large segments of the evaporite cap were removed by the detachment, breaching the seal and opening the deep basin fluid system to shallow circulating waters and upper stratigraphic units. This last point may be of particular significance since fluid mixing has been shown to be an important mineralizing process at the Jubilee deposit (Fallara et al., 1994).

Analogues of the Ainslie detachment

Comparisons with other detached salt basins around the world helps place the structural style of the Maritimes Basin into a global context. Recent compilations of detached diapir basins have been published by authors such as Jackson and Vendeville (1994). Geographically, salt basins which have developed significant diapiric structures are principally concentrated along the Atlantic and Mediterranean–Red Sea rift systems, occurring as Mesozoic to Tertiary rifted margin gravity slides or failed rift graben slumps. Although thick, late Paleozoic salt deposits are widespread, sufficient documentation has demonstrated that diapirs are surprisingly rare or altogether absent from these basins (Jackson and Vendeville, 1994). In this respect the example presented here from the Maritimes Basin is unique, as a late Paleozoic detached salt basin with extensive diapir structures. Key factors allowing for the development of diapirs include sufficiently thick salt, deep burial, sedimentary differential loading, and in particular an extensional tectonic event superposed onto the salt basin (Jackson and Vendeville, 1994).

It has been suggested that diapirs in some cases can be related to the detachment of large contiguous sheets, or rafts, along weak salt layers, with contrasting compressional and extensional domains at the front and back ends, respectively, of the allochthonous sheets (e.g. Kehle, 1970). The most extensive diapir development is predominantly, but not exclusively, restricted to the compressional front where tectonic thickening builds critical mass allowing for diapirs to evolve. A spectacular and well studied example of this type of salt buildup is from the northern Gulf of Mexico (Worrall and Snelson, 1989), which has contractional structures at the salt front that are similar to these imaged from the Maritimes Basin. The presence of a system of extensional faults onshore in Texas and Louisiana also suggests the salt is allochthous and has undergone downdip gravitation glide. It should be noted that in a similar tectonic setting in the Gulf Coast of the southern United States, heat is lost from depth by active diapiric intrusion of salt above regional detachment faults, and by the migration of hot formation waters from zones of high fluid pressure upwards along regional growth faults which are rooted in the detachments (Jones and Wallace, 1974). One can also speculate that this model for carbonate-hosted Pb-Zn mineralization in an extensional environment related to detachment faulting may have widespread significance; the similarities recorded between the Irish-type deposits and those of the Maritimes Basin, as well as the documentation of Late Carboniferous detachment faulting in both regions (Rider, 1978; Lynch and Giles, 1996), suggest that a similar mechanism may have been responsible for mineralization.

A further example of what could correspond to a regionally extensive detached salt sheet or slide occurs in the Brazilian Atlantic continental margin. Demercian et al. (1993) described regions of thin-skinned extensional and compressional deformation updip and downdip of the continental margin, respectively, which suggests the salt sheet is allochthonous.

One of the most remarkable features of detached salt basins in general, including the Maritimes Basin, is the size of the mobilized rafts. In the study region the total area of mobile salt encompassing the offshore diapir field and mapped segments of the Ainslie detachment onshore, covers approximately 67 000 km^2. This estimate represents a minimum since erosion has affected Carboniferous units and the breakaway to surface of the Ainslie detachment in the onshore areas. The allochthonous domain is comparable in size to those of the Mediterranean and the North Sea, but is an order of magnitude smaller than the zones of mobile salt from the Gulf of Mexico or the Brazilian continental margin.

Laboratory simulations are also useful in elucidating processes of salt tectonics, including the relationships between basement normal faults and detachments (Withjack et al., 1990; Jackson and Vendeville, 1994). Basement normal faults beneath salt basins do not directly truncate all of the basin stratigraphy, but rather abut against the weak salt horizon where motion is transferred creating detachment and forced folding where the raft passes over the faulted block. The apparent relationship between the Hollow fault and Ainslie detachment provides a natural example of fault displacement transfer and forced or drape folding above a single, large, master basement fault (e.g. Withjack et al., 1990). Scale models simulate detachment as thick zones of predominantly pervasive simple shear and complex flow through the salt layers, which are not modelled as being truly detached; however, the field example presented here demonstrates that much of the glide occurred along a discrete detached surface at the base of the evaporite deposits, defined in this case as the Ainslie detachment. This does not preclude that a significant component of the shear was distributed in the evaporite deposits of the Maritimes Basin.

Summary, Ainslie detachment

Carbonate units from the Windsor Group form thin, laterally persistent marker units, of which the most conspicuous is the Macumber Formation at the base of the Windsor Group. The Ainslie detachment is a regional bedding-parallel detachment fault occurring along the top of the Macumber Formation in the basal Windsor Group. The pronounced planar anisotropy within the Windsor Group exercised a fundamental control on the propagation of the fault, and the regional persistence of the Macumber Formation allowed for the lateral continuity of the detachment across a wide region. Crosscutting relationships and basin fill constrain the age of movement on the detachment to between Namurian and Westphalian C–D. The detachment apparently controlled early Westphalian clastic sedimentation of the Port Hood Formation (Cumberland Group), which is restricted to the hanging wall of the detachment. The transport direction is demonstrated from the geometry of listric normal faults above the detachment that cut downsection towards the west, and by various kinematic indicators along the detachment. Greater than 10 km of movement is estimated to have occurred along the Ainslie detachment, during westward transport.

In the eastern Gulf of St. Lawrence and in northern Nova Scotia the Windsor Group, as well as overlying Namurian to early Westphalian units, were mobilized above the Ainslie detachment, forming an allochthonous raft measuring at least 67 000 km^2. Contrasting contractional and extensional structural styles are recorded at the front and back ends of the allochthon. The front slid into a large half-graben, over the upper tip of the basement-normal Hollow fault, developing seismically imaged buckle folds, thrusts, and a broad diapir field that

characterize the contractional domain. A drape syncline at the graben margin suggests fault displacement transfer between the Hollow fault and Ainslie detachment, and that the two faults were kinematically linked. The trailing edge of the raft is exposed onshore in the uplifted regions, where listric normal faults and stratigraphic gaps above the detachment characterize the extensional domain. Downramping of the Mabou Group onto the Ainslie detachment has produced broad areas of allochthonous Mabou Group rocks flat-lying on the basal detachment, effectively excising the entire Windsor Group above the Macumber Formation. Here evaporite deposits are isolated in roll-over anticlines or as salt rollers. Fieldwork in the extensional domain demonstrates that the Ainslie detachment is stratigraphically controlled by the Macumber Formation limestone, along its weak upper contact with overlying evaporite deposits, and that deformation is thin-skinned.

Globally, detached evaporite basins and diapiric provinces are principally concentrated along the Atlantic and Mediterranean–Red Sea systems, as large Mesozoic–Tertiary continental margin gravity slides and failed rift slumps; however, the late Paleozoic diapiric Maritimes Basin in the Gulf of St. Lawrence is an exception to this and was detached in the wake of orogenic buildup, extensional collapse, and core complex formation, prior to Mesozoic Atlantic rifting.

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APPENDIX 1

List of reactions represented in Figure 11 used for pressure temperature determinations

Sample HP280
Number of independent reactions = 3

1) 6cZo + 15aQz + 5Py + 4Gr = 18An + 3Tr
2) 2cZo + aQz + Py = 2An + Tsc
3) 2cZo + aQz + Alm = 2An + FeTs
4) 24cZo + 3Tr + 10Py = 12An + 4Gr + 15Tsc
5) 3Tsc + 12aQz + 2Py + 4Gr = 12An + 3Tr
6) 4Py + 3Tr + 12cZo = 9Tsc + 6aQz + 4Gr
7) 4Gr + 10aQz + 5Tsc = 4cZo + 3Tr + 8An
8) 24cZo + 3Tr + 15Alm = 12An + 4Gr + 5Py + 15FeTs
9) 3FeTs + 12aQz + 5Py + 4Gr = 3Alm + 12An + 3Tr
10) 9Alm + 3Tr + 12cZo = 9FeTs + 6aQz + 5Py + 4Gr
11) Alm + Tsc = FeTs + Py
12) 10Alm + 3Tr + 24cZo = 10FeTs + 5Tsc + 4Gr + 12An
13) 2Alm + 4Gr + 12aQz + 5Tsc = 2 FeTs + 3Tr + 12An
14) 12cZo + 3Tr + 4Alm = 4Gr + 6aQz + 5Tsc + 4FeTs

P = 8405 ± 303 bar, T = 649 ± 23 °C

Sample KC-54
Number of independent reactions = 3

1) 3Tsc + 12aQz + 2Phl + 4Gr + 2Alm = 2Ann + 12An + 3Tr
2) 4Gr + 2Py + 12aQz + 3Tsc = 3Tr + 12An
3) FeTs + Phl = Ann + Tsc
4) Phl + Alm = Ann + Py
5) 2Alm + 4Gr + 12aQz + 5Tsc = 2FeTs + 3 Tr + 12 An
6) 3FeTs + 12aQz + 5Phl + 4Gr + 2Alm = 5Ann + 12An + 3Tr
7) 3FeTs + 12aQz + 2Py + 3Phl + 4Gr = 3Ann + 12An + 3Tr
8) Alm + Tsc = FeTs + Py
9) 3FeTs + 12aQz + 5Py + 4Gr = 3Alm + 12An + 3Tr

P = 8195 ± 215 bar, T = 699 ± 14°C

Sample GL567
Number of independent reactions = 3

1) Alm + Phl = Py + Ann
2) Gr + Ms + Py = Phl + 3 An
3) 31Py + 23Ms + 8Ann + 12W = 31Phl + 48aQz + 6St
4) Ms + Gr + Alm = Ann + 3An
5) 23Ms + 31Alm + 12W = 23Ann + 48aQz + 6St
6) 23Py + 23Ms + 8Alm + 12W = 23Phl + 48aQz + 6St
7) 6St + 48aQz + 8Ms + 31Gr = 8Ann + 93An + 12W
8) 23Gr + 8Phl + 48aQz + 6St = 8Py + 69An + 8Ann + 12W
9) 6St + 48aQz + 23Gr = 8Alm + 69An + 12W

P = 4917 ± 445 bar, T = 598 ± 21°C

Sample GL476
Number of independent reactions = 3

1) 2Ms + 4Gr + 3Cd = 12An + 2Phl + 3aQz
2) 6An + 2Py + 3aQz = 2Gr + 3Cd
3) 31Gr + 8Ms + 48aQz + 6St = 93An + 8Ann + 12W
4) Gr + Ms + Py = Phl + 3An
5) 3aQz + 4Py + 2Ms = 3Cd + 2Phl
6) 48Cd + 95Gr + 40Ms + 6St = 32Phl + 285An + 8Ann + 12W
7) 62Phl + 285aQz + 24St = 30Ms + 93Cd + 32Ann + 48W
8) 15Gr + 8Phl + 60aQz + 6St = 12Cd + 45An + 8Ann + 12W
9) 48Cd + 63Gr + 8Ms + 6St = 32Py + 189An + 8Ann + 12W
10) 16Ms + 62Py + 189aQz + 12St = 93Cd + 16Ann + 24W
11) 95Py + 55Ms + 8Ann + 12W = 48Cd + 63Phl + 6St
12) 48Cd + 55Gr + 8Phl + 6St = 40Py + 165An + 8Ann + 12W
13) 31Py + 23Ms + 8Ann + 12W = 31Phl + 48aQz + 6St
14) 16Phl + 30Py + 165aQz + 12St = 69Cd + 16Ann + 24W
15) 23Gr + 8Phl + 48aQz + 6St = 8Py + 69An + 8Ann + 12W

P = 4170 ± 51 bar, T = 476 ± 2°C

Sample KC-10
Number of independent reactions = 3

1) 7bQz + 4Ms + 2Chl + 4Alm = 4Ann + 5Cd + 8W
2) 3bQz + 2Phl + 2Ms + 4Alm = 4Ann + 3Cd
3) 31Alm + 23Ms + 12W = 6St + 48bQz + 23Ann
4) 7Phl + Ms + 8Alm + 12W = 8Ann + 3Chl + 3Cd
5) 4Alm + 5Phl + 12W = 3bQz + Ms + 3Chl + 4Ann
6) Chl + Ms + 2bQz = Phl + Cd + 4W
7) 4Alm + 4Phl + 8W = bQz + Cd + 2Chl + 4Ann
8) 74Alm + 6Chl + 58Ms = 12St + 75bQz + 15Cd + 58Ann
9) 409Alm + 96Chl + 353Ms = 42St + 240Cd + 353Ann + 300W
10) 24St + 409bQz + 32Ms + 62Chl = 32Ann + 155Cd + 296W
11) 24St + 353bQz + 46Chl = 32Alm + 115Cd + 232W
12) 95Alm + 55Ms + 32Phl + 12W = 6St + 48Cd + 87Ann
13) 24St + 285bQz + 62Phl = 32Ann + 93Cd + 30Ms + 48W
14) 24St + 261bQz + 46Phl = 32Alm + 69Cd + 46Ms + 48W
15) 12St + 165bQz + 46Phl + 30Alm = 46Ann + 69Cd + 24W
16) 25Phl + 54Ms + 3Chl + 87Alm = 79Ann + 45Cd + 6St
17) 27Alm + 3Chl + 24Ms = 6St + 45bQz + 5Phl + 19Ann
18) 24St + 261bQz + 74Phl = 32Ann + 12Chl + 81Cd + 42Ms
19) 24St + 237bQz + 58Phl = 32Alm + 12Chl + 57Cd + 58Ms
20) 6St + 81bQz + 29Phl + 21Alm = 29Ann + 3Chl + 36Cd
21) 6St + 80Phl + 33Alm + 180W = 41Ann + 48Chl + 39Ms
22) 64Ann + 285Chl + 345Ms = 48St + 409Phl + 99Cd + 1044W
23) 64Alm + 261Chl + 353Ms = 48St + 353Phl + 123Cd + 948W
24) 6St + 353Phl + 345Alm + 648W = 353Ann + 165Chl + 117Cd
25) 32Ann + 93Chl + 123Ms = 24St + 99bQz + 155Phl + 324W
26) 32Alm + 69Chl + 115Ms = 24St + 123bQz + 115Phl + 228W
27) 123Alm + 115Phl + 288W = 6St + 117bQz + 69Chl + 115Ann
28) 30Chl + 32Phl + 345bQz + 24St = 123Cd + 32Ann + 168W

Alm, almandine
An, anorthosite
Ann, annite
Cd, cordierite
Chl, chlorite
FeTs, iron techermakite
Gr, garnet
Ms, muscovite
aQz, alpha-quartz
bQz, beta-quartz
Phl, phlogopite
Py, pyrite
St, staurolite
Tr, tremolite
Tsc tschermakite
W, water
cZo, clino-zoisite