UNIVERSITÉ DU QUÉBEC À MONTRÉAL

ÉVOLUTION TECTONOSTRATIGRAPHIQUE DU DOMAINE OCÉANIQUE DES APPALACHES DU SUD DU QUÉBEC DANS SON CONTEXTE PÉRI-LAURENTIEN

THÈSE PRÉSENTÉE

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RÉSUMÉ

L'évolution paléozoïque des Appalaches a été ponctuée par l'obduction d'ophiolites sur la paléo-marge laurentienne durant l'Ordovicien moyen à supérieur. Les travaux présentés dans cette thèse visent à déterminer la nature et la chronologie des phénomènes tectono-sédimentaires associés à cette obduction, tels que préservée au sein de la chaîne taconienne du sud du Québec. Les roches cambriennes-ordoviciennes des Appalaches du sud du Québec sont divisées en deux zones lithotectoniques : les zones de Humber et de Dunnage, qui représentent, respectivement, les vestiges de la paléo-marge laurentienne et du domaine océanique adjacent. Une courverture sédimentaire silurienne-dévonienne, la Ceinture de Gaspé, masque toutefois une partie importante des terrains cambriens-ordoviciens. La zone de Dunnage inclut, d'ouest en est, une ceinture ophiolitique, le Mélange de Saint-Daniel, le Groupe de Magog et l'arc volcanique du Complexe d'Ascot. De nouvelles données stratigraphiques et géochronologiques $^{40}Ar/^{39}Ar$ et U-Pb sont présentées ici selon le contexte des Appalaches du sud du Québec et du Vermont afin de reconstituer et synthétiser l'évolution lithotectonique de la chaîne taconienne.

Les travaux ont été concentrés sur le Mélange de Saint-Daniel et certaines unités sédimentaires et ophiolitiques qui lui sont associées dans les régions de la Beauce et du lac Memphrémagog : le Complexe ultramafique de la Rivière-des-Plante et l'ophiolite du Lac-Brompton, la Volcanite de Ware, la séquence de la Colline Bunker et le Groupe de Bolton. Le Complexe ultramafique de la Rivière-des-Plante, tout comme l'ophiolite du Lac-Brompton, sont composés de péridotite mantellique harzburgitique recoupée par des intrusions syn-tectoniques de granitoïde à muscovite et biotite. Ces granitoïdes sont interprétés comme résultant de la fusion partielle des roches sédimentaires de la marge continentale laurentienne durant l'obduction des ophiolites. Le Complexe ultramafique de la Rivière-des-Plante est corrélé stratigraphiquement avec les ophiolites de Thetford-Mines, d'Asbestos et de Lac-Brompton, qui sont collectivement interprétés comme faisant partie d'un même segment de lithosphère océanique obductée. Le Mélange de Saint-Daniel repose en discordance sur l'ensemble de ces massifs ophiolitiques et provient en partie de l'érosion des roches ophiolitiques et celles de la marge laurentienne. Cette discordance entaille profondément les ophiolites et atteint localement les unités continentales sous-jacentes.

La séquence de la colline Bunker est divisée en un membre sédimentaire et un membre volcanique et, contrairement aux études précédentes, elle est excluse du Mélange de Saint-Daniel. Le membre sédimentaire peut être corrélé avec les roches métasédimentaires de la zone de Humber, tandis que le membre volcanique possède des affinités avec les unités inférieures du Groupe de Magog, ce qui est concordant avec sa position stratigraphique et un âge U-Pb de 455 ± 6 Ma qui a été obtenu pour ses roches volcanoclastiques. Le Groupe de Bolton est constitué de coulées basaltiques interstratifiées dans le Mélange de Saint-Daniel et d'intrusions mafiques qui le recoupent. La Volcanite de Ware est également interstratifiée dans le Mélange de Saint-Daniel, mais celle-ci est principalement de composition felsique.

Nos travaux de géochronologie $^{40}Ar/^{39}Ar$ sur le Complexe ultramafique de la Rivière-des-Plante et l'ophiolite du Lac-Brompton suggèrent qu'ils ont subi une histoire thermochronologique semblable à celle de l'ophiolite de Thetford-Mines. Suite à la cristallisation de la lithosphère océanique, l'obduction des ophiolites a été marquée par la formation des roches métamorphiques infraophiolitiques entre ca. 479 et 472 Ma, et s'est terminée vers 460-457 Ma, un âge qui correspond aux derniers stades de recristallisation et de déformation ductile dans la nappe ophiolitique et à la sédimentation du Mélange de Saint-Daniel. Nos données soulignent également un diachronisme dans l'obduction des ophiolites entre les régions de la Gaspésie et du sud du Québec, un phénomène qui est possiblement relié à la géométrie irrégulière de la zone de collision. La présence du Groupe de Bolton et de la Volcanite de Ware au sein du Mélange de Saint-Daniel peut être expliquée par un événement de délamination de la plaque subductée durant les stades tardifs de mise en place des ophiolites.

Le Groupe de Magog repose sur le Mélange de Saint-Daniel à la faveur d'une discordance et ensemble ils forment un bassin avant-arc syn-collisionnel. Les provenances suggérées par la datation de zircons et muscovites détritiques provenant du Groupe de Magog et de la Formation de Compton (Ceinture de Gaspé) peuvent être entièrement réconciliées avec le contexte péri-laurentien de ces formations sédimentaires.

Mots clés : Appalaches du Québec, ophiolite, tectonique, obduction, mélange, Dunnage, stratigraphie, géochronologie, $^{40}Ar/^{39}Ar$, U-Pb

CHAPITRE I

INTRODUCTION

1.1 Organisation de la thèse

Les résultats présentés dans cette thèse sont le produit de trois campagnes de terrain d'environ trois mois chacune dans la région de l'Estrie-Beauce du sud du Québec au cours des étés 2007, 2008 et 2009. Des travaux de reconnaissance et d'échantillonnage ont aussi été réalisés aux États-Unis en 2009 dans les états du Vermont et du Maine, et deux semaines de terrain ont été requises à l'été 2010 afin de compléter certaines observations. Un stage de cartographie a également été réalisé en Albanie au printemps 2007 afin de comparer les roches ophiolitiques des Dinarides avec celles des Appalaches du Québec.

La thèse est divisée en quatre chapitres. Le premier est constitué d'une définition de la problématique de recherche abordée dans ce doctorat. Les chapitres II, III et IV sont rédigés sous la forme d'articles scientifiques dans lesquels sont présentés et discutés l'essentiel des résultats. Le premier article est déjà publié, le second est sous presse et le troisième a été préparé pour une soumission ultérieure. Une liste exhaustive des publications, résumés de conférences et documents produits dans le cadre de ce doctorat est donnée à la fin du premier chapitre (cf. 1.3 Liste des contributions). En plus de travaux originaux effectués pendant cette thèse, le Chapitre III inclut des travaux de cartographie et d'échantillonnage effectués par N. Pinet et A. Tremblay dans la péninsule gaspésienne en 2007. Dans les annexes A, B et C sont présentées des cartes géologiques qui synthétisent les observations et interprétations cartographiques réalisées pendant les séjours de terrain. Les annexes D et E contiennent, respectivement, des analyses géochimiques multiélémentaires publiées (c.f. Chapitre II) et non-publiées, tandis que l'Annexe F contient une série d'analyses U-Pb qui sont présentés et discutés dans le Chapitre IV.

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1.2 Problématique et objectifs de la recherche

La chaîne orogénique des Appalaches résulte de la fermeture des océans Iapétus (Cambrien-Dévonien) et Rhéic (fig. 1.1; Dévonien-Carbonifère). Elle sillonne la marge est de l'Amérique du Nord en s'étendant du sud-est des États-Unis (Alabama-Arkansas) jusqu'à Terre-Neuve et peut être corrélée avec la chaîne des Calédonides du Groenland et du nord-est de l'Europe (van Staal et al., 1998; van Staal, 2005 ; Dewey et Strachan, 2005; Gee, 2005). Le dessin irrégulier des terrains appalachiens définit une série de promontoires et réentrants (Thomas, 1977). Le promontoire de New York sert ainsi de division géographique arbitraire, entre les Appalaches du Nord et les Appalaches du Sud (fig. 1.2 ; Hibbard, van Staal et Rankin, 2007, 2010). La zone d'étude est située dans les Appalaches du Nord, au niveau du réentrant de Québec, lequel comporte cinq grandes zones tectono-stratigraphiques néoprotérozoïques à ordoviciennes, qui sont, d'ouest en est : les zones de Humber, Dunnage, Gander, Avalon et Meguma (fig. 1.2; Williams, 1979; Williams, Colman-Sadd et Swinden, 1988) :

1) *La zone de Humber* correspond à l'ancienne marge continentale du paléo-continent Laurentia, à l'ouest de l'océan Iapétus;
2) *La zone de Dunnage* représente les vestiges de l'océan Iapétus, incluant des séquences d'arcs insulaires, des ophiolites, ainsi que divers types de mélanges et formations sédimentaires qui ont évolué en périphérie des paléo-continents Laurentia (sous-zone de Notre-Dame) et Gondwana (sous-zone d'Exploits);
3) *La zone de Gander* est considérée comme l'alter ego de la zone de Humber ayant été formée sur la marge est de l'océan Iapétus, en périphérie de Gondwana;
4) *La zone d'Avalon* inclut les vestiges d'arcs volcaniques et de bassins volcano-sédimentaires néoprotérozoïques dont l'évolution en bordure de Gondwana a précédé l'ouverture de l'océan Iapétus;
5) *La zone de Meguma* est interprétée comme un fragment de Gondwana qui est resté accolé à Laurentia suite à l'ouverture de l'océan Atlantique.

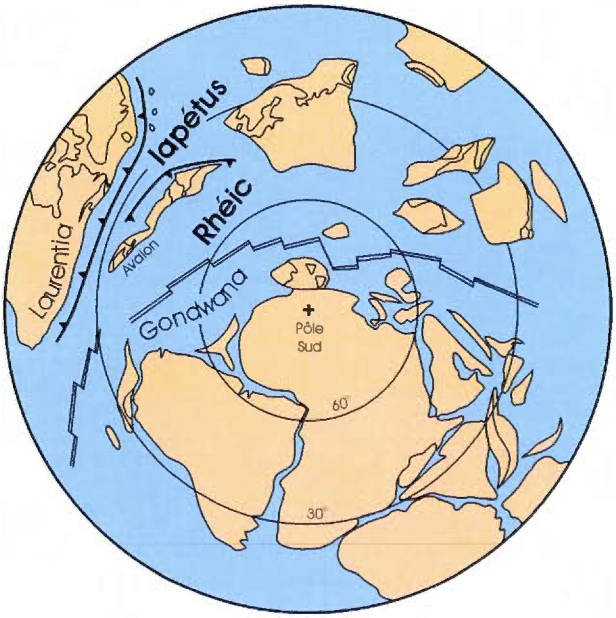

Figure 1.1 : Reconstitution paléo-géographique montrant la position des principales masses continentales, ainsi que l'étendue des océans Iapétus et Rhéic à 460 Ma. Modifié de Cocks et Torsvik (2002).

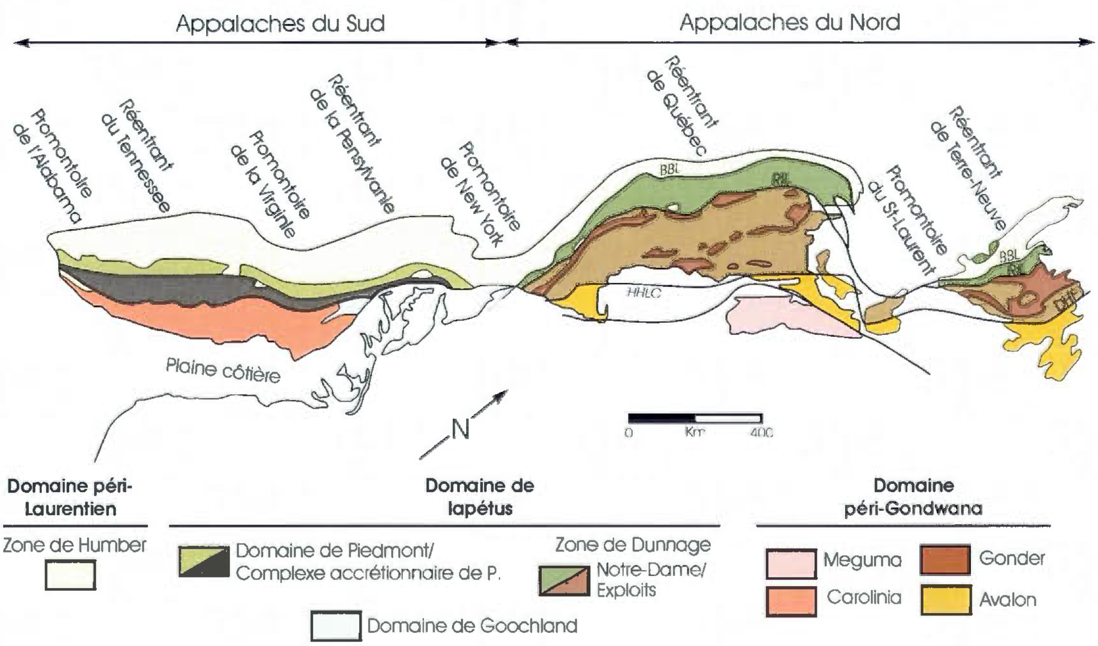

Figure 1.2 : Principales subdivisions litho-tectoniques des terrains appalachiens pré-siluriens. (Modifié de Hibbard, van Staal et Rankin, 2007). BBL – ligne Baie Verte-Brompton; DHF – faille de Dover-Hermitage Bay; HHLC – zone de faille de Honey Hill-Lake Char; RIL – ligne Red Indian.

Les zones de Dunnage, Gander, Avalon et Meguma ont été successivement accrétées à la marge laurentienne suite à quatre phases orogéniques principales, qui sont, respectivement, l'orogénie taconienne (Cambrien-Ordovicien), salinienne (Silurien), acadienne (Silurien-Dévonien) et Alléghanienne (Carbonifère-Permien) (Osberg et al., 1989; Robinson et al., 1998; van Staal et al., 1998; van Staal, 2005, 2007; Malo et al., 2008). Seules les zones de Humber et Dunnage (sous-zone de Notre-Dame), ainsi que les déformations et épisodes métamorphiques d'âge taconien, salinien et acadien sont préservés dans les Appalaches du Québec (St-Julien et Hubert, 1975; Castonguay et al., 2001, 2007; Tremblay et Castonguay, 2002). La Ceinture de Gaspé, une couverture sédimentaire silurienne-dévonienne s'étendant de l'état du Connecticut jusqu'à la péninsule gaspésienne, masque une partie importante des terrains cambriens-ordoviciens (Bourque, Malo et Kirkwood, 2000; Lavoie et Asselin, 2004 ; Tremblay et Pinet, 2005). Les limites géographiques des zones de Humber et Dunnage sont bien documentées et correspondent à des failles ou zones de failles dont la cinématique et la géométrie présumée sont illustrées sur la figure 1.3. La limite nord-ouest des terrains allochtones de la zone de Humber est représentée par la ligne de Logan, tandis que la ligne Baie Verte – Brompton et la faille de La Guadeloupe correspondent, respectivement, aux limites nord-ouest et sud-est de la zone de Dunnage dans le sud du Québec.

Le sud du Québec présente un maximum d'unités de la zone de Dunnage, soit, d'ouest en est : une ceinture ophiolitique qui inclut les complexes ophiolitiques de Thetford-Mines, d'Asbestos, du Lac-Brompton et du Mont-Orford, le Mélange de Saint-Daniel, le bassin avant-arc du Groupe de Magog et l'arc volcanique du Complexe d'Ascot (fig. 1.3; Tremblay, Malo et St-Julien, 1995). Cette région demeure donc un secteur-clé pour l'étude et l'interprétation du schéma évolutif tectono-stratigraphique associé à la fermeture de l'océan Iapétus et à la construction de la chaîne taconienne.

Plusieurs études comparant la géologie du secteur de l'Estrie-Beauce avec des environnements géotectoniques plus jeunes et mieux préservés, tels que la chaîne d'obduction omanaise (Pinet et Tremblay, 1995a, 1995b) ou la zone de collision arc-continent de Banda (fig. 1.4 Tremblay, 1992a), ont été proposées afin d'expliquer la nature et la géométrie finie des zones de Humber et Dunnage. Dans ces modèles, le Mélange de Saint-Daniel représente les vestiges d'un complexe de subduction taconien dans lequel auraient été incorporées diverses unités sédimentaires, métasédimentaires, volcaniques et ophiolitiques (Cousineau, 1990; Cousineau et St-Julien, 1992).

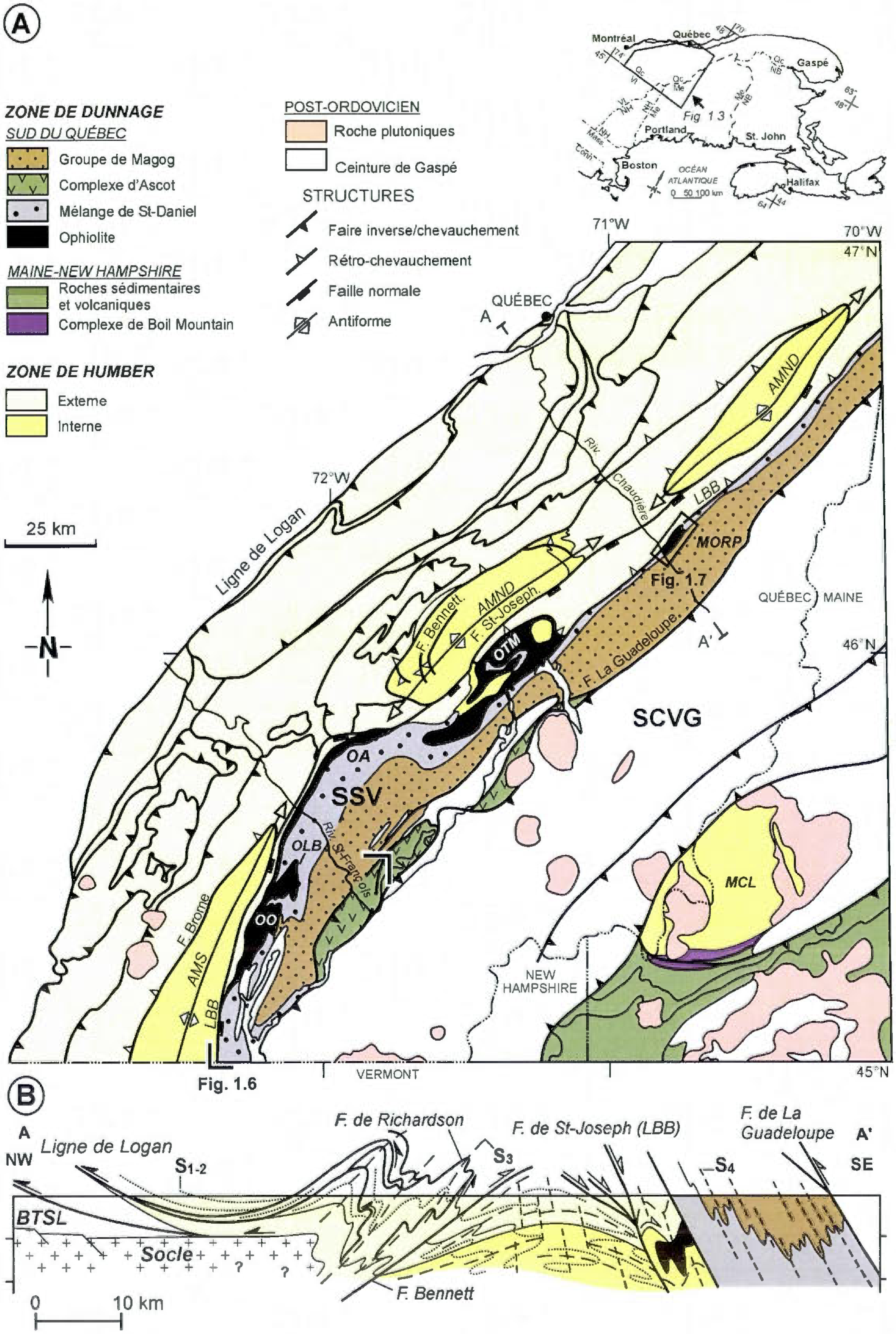
A
ZONE DE DUNNAGE
SUD DU QUÉBEC
Groupe de Magog
Complexe d'Ascot
Mélange de St-Daniel
Ophiolite
MAINE-NEW HAMPSHIRE
Roches sédimentaires et volcaniques
Complexe de Boil Mountain
ZONE DE HUMBER
Externe
Interne
POST-ORDOVICIEN
Roche plutoniques
Ceinture de Gaspé
STRUCTURES
Faire inverse/chevauchement
Rétro-chevauchement
Faille normale
Antiforme
Montréal
Québec
Gaspé
Portland
St. John
Boston
Halifax
OCÉAN ATLANTIQUE
Fig 1.3
25 km
N
Ligne de Logan
QUÉBEC
71°W
70°W
47°N
72°W
46°N
45°N
AMND
LBB
MORP
Fig. 1.7
Riv. Chaudière
F. Bennett
F. St-Joseph
OTM
F. La Guadeloupe
QUÉBEC MAINE
SCVG
OA
SSV
OLB
Riv. St-François
F. Brome
OO
AMS
MCL
NEW HAMPSHIRE
VERMONT
Fig. 1.6
B
A
NW
Ligne de Logan
F. de Richardson
F. de St-Joseph (LBB)
F. de La Guadeloupe
A'
SE
BTSL
Socle
F. Bennett
0
10 km

Figure 1.3 : a) Carte géologique de la région de l'Estrie-Beauce des Appalaches du sud du Québec et de la Nouvelle-Angleterre. Tirée de De Souza et Tremblay (2010). AMND – anticlinorium des Monts-Notre-Dame; AMS – anticlinorium des Monts-Suttons; LBB – ligne Baie Verte-Brompton ; MCL – Massif de Chain Lakes; MORP – Mélange ophiolitique de la Rivière-des-Plante; OA – ophiolite d'Asbestos; OO – ophiolite du Mont-Orford; OTM – ophiolite de Thetford-Mines; Synclinorium de Conencticut Valley-Gaspé; SSV – synclinorium de Saint-Victor. b) Coupe structurale illustrant la géométrie et la cinématique des déformations régionales des Appalaches du sud du Québec. Tirée de Tremblay et Pinet (2005).

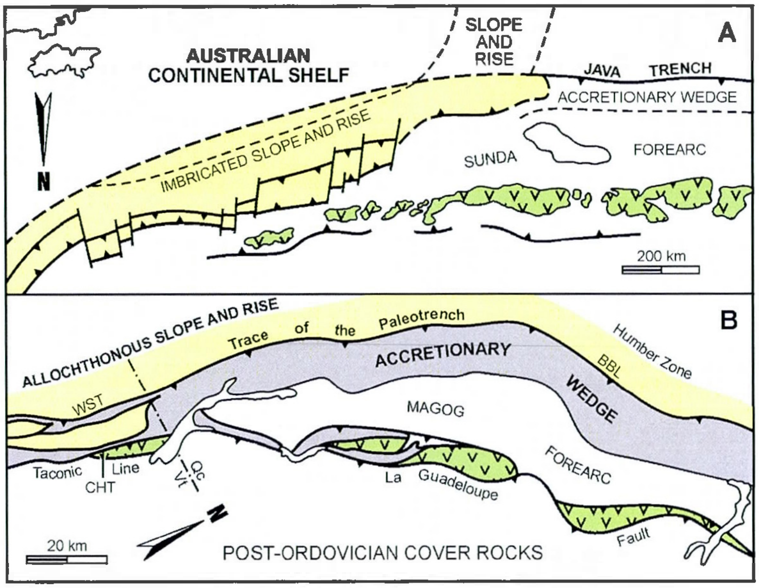

Figure 1.4 : Analogies structurales et tectoniques entre la zone de collision l'arc de Banda du nord de l'Australie et le domaine océanique du sud du Québéc et du nord du Vermont; a) Carte schématique de la zone de collision de l'arc Banda et de la fosse de Java; b) Carte schématique du domaine océanique des Appalaches du sud du Québec et du nord du Vermont. WST – Whitcomb Summit thrust; CHT – Coburn Hill thrust. Modifié de Tremblay (1992a).

Plus récemment, Schroetter (2004), dans une étude détaillée portant sur l'évolution stratigraphique et tectonique du complexe ophiolitique de Thetford-Mines, est plutôt arrivé aux conclusion suivantes :

1) la structure de l'ophiolite peut être expliquée par la superposition de déformations pré- (syn-océaniques), syn- et post-obduction ;

2) le Mélange de Saint-Daniel est d'origine sédimentaire. Il repose en discordance sur différents niveaux stratigraphiques de l'ophiolite, soit de la croûte supérieure-inférieure et possiblement du manteau ;

3) Le Mélange de Saint-Daniel constitue la base d'un bassin sédimentaire syn-obduction principalement représenté par le Groupe de Magog, et provient de l'érosion de reliefs composés d'ophiolites et de roches métamorphiques continentales.

Ces dernières conclusions ont également pu être démontrées dans les régions du Lac-Brompton (Daoust, 2007 ; De Souza et al., 2008) et d'Orford (Schroetter, Tremblay et Bédard, 2005 ; Schroetter et al., 2006), et suggèrent possiblement une évolution tectonique commune pour l'ensemble de ces régions. La problématique principale de cette thèse est donc de revoir, à l'échelle de la région de l'Estrie-Beauce, l'évolution stratigraphique, tectonique et temporelle de la zone de Dunnage dans le contexte du réentrant de Québec. Plus précisément, il est proposé d'étudier les relations tectono-stratigraphiques à l'intérieur même du Mélange de Saint-Daniel et avec les unités adjacentes des ophiolites, du Groupe de Magog et du Complexe d'Ascot dans le but de comprendre les processus qui ont mené à la construction de la chaîne taconienne.

1.2.1 Les Appalaches du sud du Québec

Dans le sud du Québec, la zone de Humber forme une ceinture de plis-et-chevauchements (fig. 1.3; St-Julien et Hubert, 1975; Tremblay et Castonguay, 2002; Sasseville et al., 2008). Elle est principalement constituée de roches sédimentaires silicoclastiques et de roches volcaniques d'âge Néoprotérozoïque à Ordovicien inférieur représentant des faciès de rift et de marge passive (St-Julien et Hubert, 1975; Lavoie et al., 2003). La zone de Humber est subdivisée en une zone externe composée de roches faiblement à non-métamorphisées, et une zone interne caractérisée par un métamorphisme variant du faciès des schistes verts à celui des amphibolites (fig. 1.3; Tremblay et Castonguay, 2002). La zone interne est limitée au sud-est par la faille de Saint-Joseph, une importante faille normale qui forme localement une structure composite avec la ligne Baie Verte-Brompton (Pinet, Tremblay et Sosson, 1996; Tremblay et Castonguay, 2002, 2003). La zone de Dunnage affleure dans le toit de la faille Saint-Joseph et au sud-est de la ligne Baie Verte – Brompton, au sein du synclinorium de Saint-Victor (fig. 1.3; Tremblay 1992a ; Tremblay, Malo et St-Julien, 1995). La Ceinture de Gaspé est constituée de deux ensembles stratigraphiques, l'un est autochtone et repose en discordance sur les zones de Humber et Dunnage, tandis que l'autre affleure au sud-est de la faille de La Guadeloupe et forme les assises du synclinorium de Connecticut Valley – Gaspé (fig. 1.3; Lavoie et Asselin, 2004; Tremblay et Pinet, 2005).

La déformation régionale dans les Appalaches du sud du Québec peut être expliquée par la superposition de trois épisodes de déformation et métamorphisme dont l'évolution est contrainte par diverses méthodes de datations isotopiques et biostratigraphiques (fig. 1.5; Tremblay et

Castonguay, 2002; Castonguay et al., 2001, 2012; Castonguay, Ruffet et Tremblay, 2007; Sassseville et al. 2008; Tremblay, Ruffet et Bédard, 2011). Le premier épisode est associé au chevauchement vers le nord-ouest des nappes de la zone de Humber et à l'obduction des ophiolites durant l'orogénie taconienne à l'Ordovicien moyen à supérieur (471-445 Ma). Le second correspond à un épisode de rétro-chevauchement possiblement associé à l'orogénie salinienne ou à l'effondrement gravitaire de la chaîne taconienne au Silurien-Dévonien inférieur (435-405 Ma). Enfin, le métamorphisme régional du Dévonien moyen est attribué à une compression acadienne (390-376 Ma).

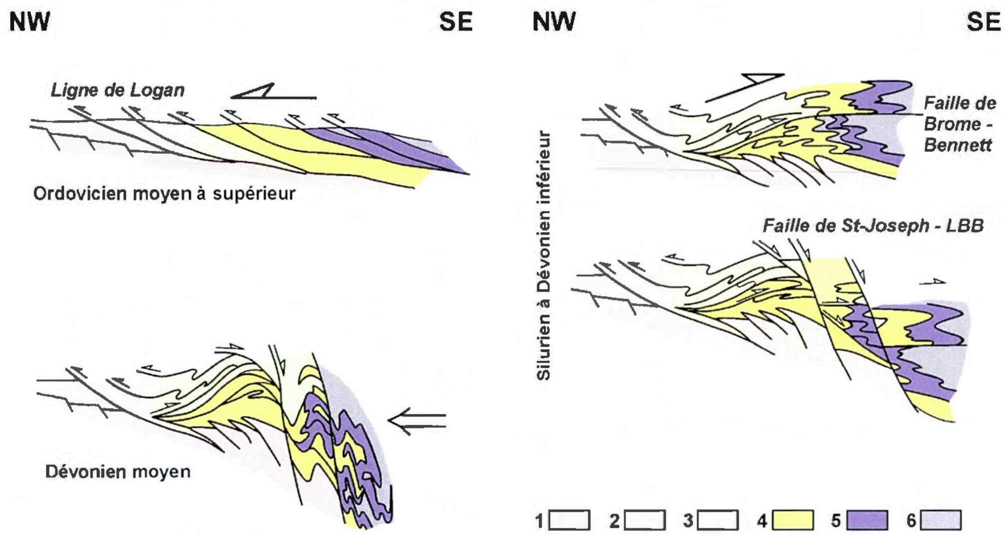

Figure 1.5 : Modèle schématique pour l'évolution structurale de la marge laurentienne dans le sud du Québec. 1 – socle grenvillien; 2 – Basses-terres-du-St-Laurent; 3 – zone de Humber externe; 4 – zone de Humber interne; 5 – ophiolites; 6 – roches sédimentaires de la zone de Dunnage. Tiré de Tremblay et Castonguay (2002).

1.2.1.1 Les ophiolites

La ceinture ophiolitique du sud du Québec est constituée de quatre principaux massifs, ceux de Thetford-Mines, d'Asbestos, du Lac-Brompton et du Mont-Orford (fig. 1.3a). Les complexes du Lac-Brompton, d'Asbestos et de Thetford-Mines sont principalement composés de lithosphère océanique (croûte et manteau) d'affinité boninitique et sont interprétées comme des segments d'une même nappe ophiolitique (Schroetter et al., 2003; Schroetter, Tremblay et Bédard, 2005; De Souza et al., 2008; Tremblay, Meshi et Bédard, 2009).

Deux âge U-Pb de 478 +3/-2 et 480 ± 2 Ma ont été obtenus sur les plagiogranites du complexe de Thetford-Mines et sont interprétés comme représentant l'âge de formation de l'ophiolite

(Whitehead, Dunning et Spray, 2000), tandis que les âges $^{40}Ar/^{39}Ar$ sur sa semelle métamorphique varient de 477 ± 5 et 471 Ma sur amphibole, à 466 Ma et 460-457 Ma sur muscovite (Whitehead, Reynolds et Spray, 1995; Tremblay, Ruffet et Bédard, 2011). Des roches granitiques à biotite et muscovite sont également présentes dans les faciès mantelliques des ophiolites; ceux-ci ont donné des âges U-Pb de 470 +5/-3 Ma et 469 ± 4 Ma (Whitehead, Dunning et Spray, 2000) et $^{40}Ar/^{39}Ar$ sur muscovite de ca. 466 Ma à 460 Ma dans la région de Thetford-Mines (Tremblay, Ruffet et Bédard, 2011). Ces granitoïdes sont interprétés, sur les bases de leur composition géochimique et minéralogique, ainsi que de modélisations thermodynamiques, comme les produits de la fusion partielle de la marge laurentienne durant l'obduction des ophiolites (Whitehead, Dunning et Spray, 2000; Tremblay, Ruffet et Bédard, 2011). Le complexe du Mont-Orford est exclusivement constitué de faciès crustaux, incluant des roches volcaniques et des filons d'affinité boninitique, tholéiitique et transitionnelle-alcaline (Harnois et Morency, 1989; Laurent et Hébert, 1989; Huot, Hébert et Turcotte, 2002). Un âge U-Pb sur zircon de 504 ± 3 Ma a été obtenu sur une trondhjémite associée aux gabbros et diabases de ce complexe (David et Marquis, 1994). Le contexte tectono-stratigraphique des roches ophiolitiques est donc particulièrement bien contraint dans la région de Thetford-Mines par rapport aux régions adjacentes, notamment celle du Lac-Brompton, où il n'existe aucune donnée géochronologique pour les roches ophiolitiques ou celles de la semelle métamorphique. Le lien possible entre les ophiolites et certaines unités du Mélange de Saint-Daniel ayant une origine présumément ophiolitique reste spéculatif. Un des objectifs des travaux a donc été de compléter l'échantillonnage géochronologique des roches ophiolitiques et métamorphiques associées afin d'infirmer, de confirmer ou de proposer des corrélations tectono-stratigraphiques le long de la ceinture ophiolitique. Un lien génétique possible avec certains fragments ophiolitiques du Mélange de Saint-Daniel est également à vérifier.

1.2.1.2 Le Mélange de Saint-Daniel

Le Mélange de Saint-Daniel affleure principalement sur le flanc nord-est du synclinorium de St-Victor ainsi que dans la région du lac Memphrémagog, entre la baie Fitch et le lac Massawippi (figs. 1.3a, 1.6). Ce mélange est caractérisé par d'importantes variations lithologiques latérales et verticales, et son épaisseur demeure imprécise. Il est principalement composé de roches sédimentaires gréso-pélitiques à conglomératiques et bréchiques dont les faciès les plus typiques sont une argilite à cailloux et une interstratification de grès, siltstone et argilite laminée (St-

Julien, 1987; Lavoie, 1989; Cousineau 1990; Schroetter et al., 2006). Toutefois, il contient également des lithologies ophiolitiques, métamorphiques, volcaniques et sédimentaires qui sont considérées comme des écailles tectoniques au sein du mélange (Lamothe, 1981a; Cousineau 1990; Tremblay, 1990, 1992a). Cousineau (1990) et Tremblay (1992a) l'interprètent comme un complexe accrétionnaire composé d'une matrice sédimentaire contenant une série d'écailles tectoniques et d'olistolithes kilométriques à pluri-kilométriques d'origines variées.

Les principales unités qui sont considérées comme des écailles tectoniques sont : 1) le Mélange ophiolitiques de la Rivière-des-Plante, 2) la Volcanite de Ware, 3) le Groupe de Bolton, et 4) la séquence de la colline Bunker (Cousineau, 1990; Tremblay, 1992a; Tremblay, Malo et St-Julien, 1995). Bien qu'elles recèlent une importance de premier plan dans les reconstructions paléotectoniques, l'origine et l'âge de ces unités, ainsi que leurs relations tectoniques et stratigraphiques avec les faciès sédimentaires « typiques » du Mélange de Saint-Daniel, sont jusqu'alors peu ou pas documentés de façon détaillée. Un des buts de cette étude est donc de déterminer ces caractéristiques pour les unités susmentionnées et détaillées ci-bas.

1) Le *Mélange ophiolitique de la Rivière-des-Plante* (fig. 1.7; Cousineau 1990, 1991) est situé le long de la ligne Baie Verte – Brompton au niveau de la rivière Chaudière. Il est essentiellement composé de divers types de péridotite serpentinisée, de granitoïde, ainsi que d'une unité de brèche felsique localement mylonitique et contenant une paragénèse minérale suggérant un métamorphisme au faciès des amphibolites. Le protolithe de cette brèche est interprété comme un grès feldspathique ou un diamicton (Cousineau, 1990, 1991; Trzcienski, Rodgers et Guidotti, 1992). Ce type d'association lithologique a également été reconnu dans la zone de Dunnage de la péninsule gaspésienne et en Nouvelle-Angleterre (Williams et St-Julien, 1982; De Brouker, 1986; Trzcienski, Rodgers et Guidotti, 1992). Dans cette dernière région les roches métamorphiques du massif de Chain Lakes (gneiss, brèche et migmatites) possèdent des similarités pétrographiques avec celles du Mélange ophiolitique de la Rivière-des-Plante et est intimement relié à des roches ophiolitiques (Boone et Boudette, 1989; Cousineau, 1991; Trzcienski, Rodgers et Guidotti, 1992; Gerbi et al., 2006; Gerbi, Johnson et Aleinikoff, 2006). Bien que plusieurs interprétations ont été proposées pour l'origine du Mélange ophiolitique de la Rivière-des-Plante et un possible lien génétique avec le massif de Chain Lakes, celles-ci demeurent incertaines (Pinet et Tremblay, 1995b; Kusky, Chow et Bowring, 1997; Gerbi et al., 2006; Gerbi, Johnson et Aleinikoff, 2006).

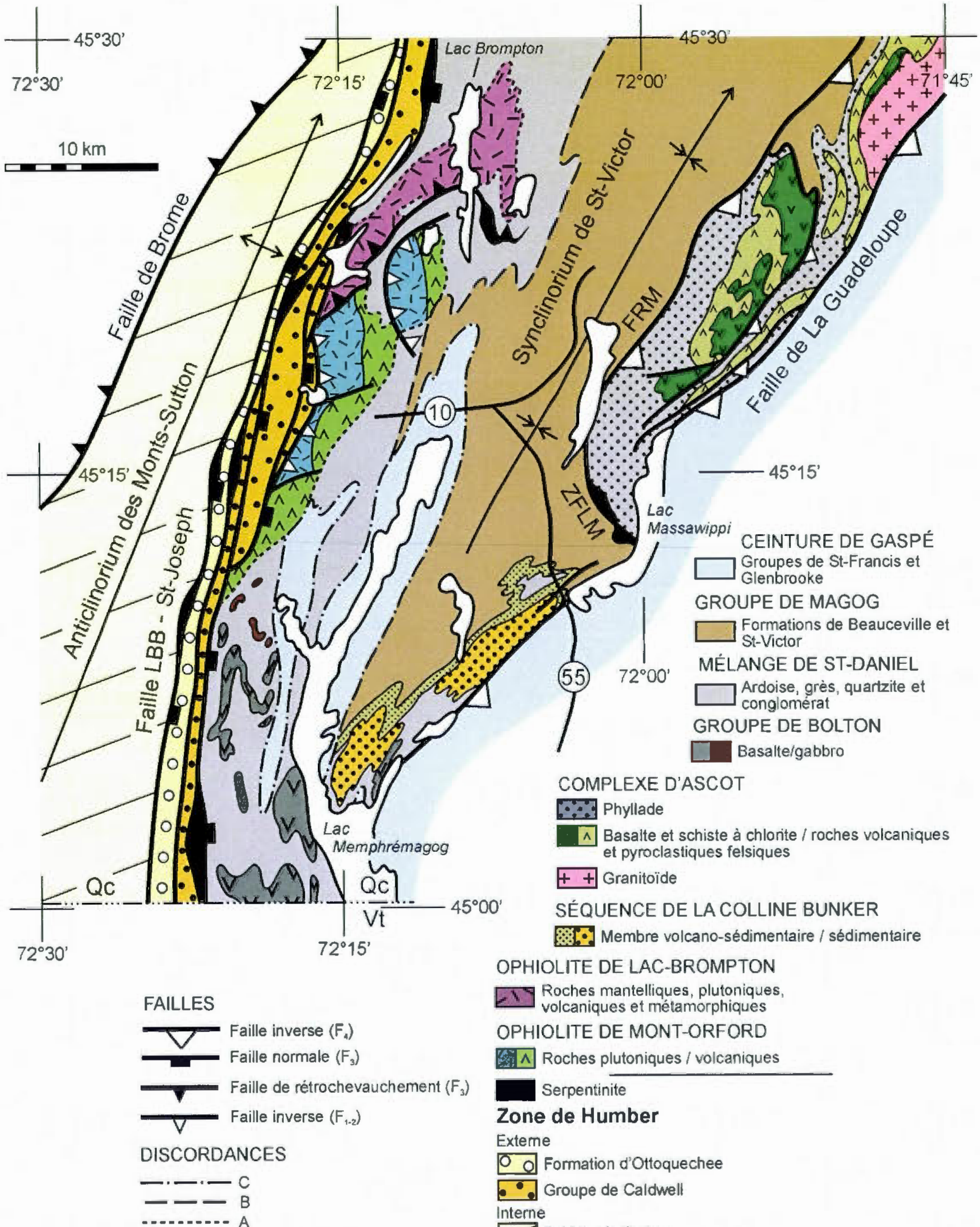

Figure 1.6 : Carte géologique de la région du lac Memphrémagog. LBB – Ligne Baie Verte-Brompton; FRM – Faille de la rivière Magog; ZFLM – Zone de faille du lac Massawippi. Lamothe, 1979, 1981a, 1981b; Modifié de Doolan et al., (1982), Slivitzky et St-Julien (1987); Tremblay (1990, 1992b).

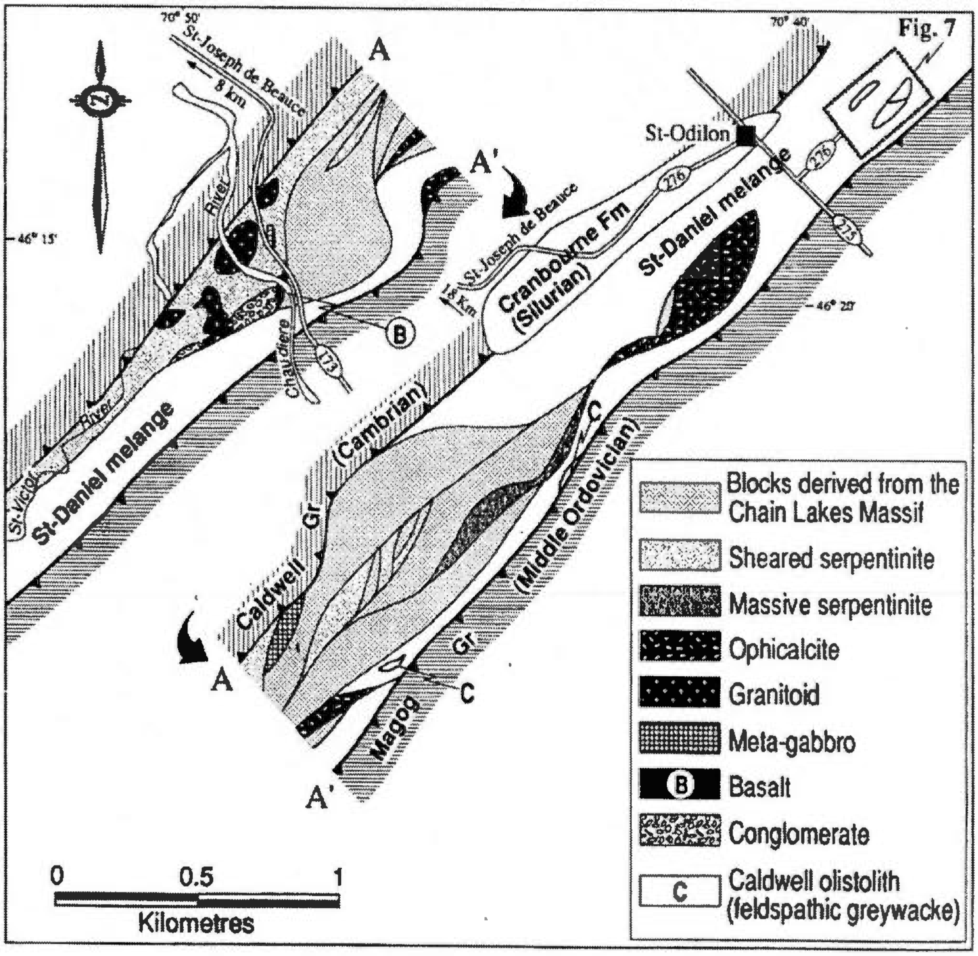

Figure 1.7 : Carte géologique du Mélange ophiolitique de la Rivière-des-Plante. Tirée de Cousineau (1991).

2) Le terme « *Volcanite de Ware* » a été introduit par Cousineau (1990) pour désigner des roches volcaniques et pyroclastiques de composition felsique qui affleurent dans le Mélange de Saint-Daniel à 70 km au sud-est de la ville de Québec. Puisque Cousineau (1990) interprète la Volcanite de Ware comme une écaille tectonique au sein du Mélange de Saint-Daniel, il lui confère un âge Ordovicien inférieur. D'après Béland (1957), les roches de la Volcanite de Ware reposent plutôt en discordance sur le Mélange de Saint-Daniel (alors inclus dans le Beauceville de McKay, 1921). Celui-ci les considère alors comme étant postérieures à l'Ordovicien moyen et leur assigne, sous toute réserve, un âge Silurien ou Dévonien.

3) Le *Groupe de Bolton* est formé de roches volcaniques et intrusives mafiques qui affleurent au sein des ardoises du Mélange de Saint-Daniel et constituent des reliefs ponctuels ou en forme de

crête allongée entre le lac Memphrémagog et la ligne Baie Verte - Brompton (fig. 1.6; Clark et Fairbairn, 1936; Cooke, 1950; Lamothe, 1981b; Rickard, 1991). Ces basaltes et gabbros-diabases sont corrélés, au Vermont, avec les Volcanites de Coburn Hill (Doolan et al., 1982) et la Suite Intrusive du Mont Norris (Kim et al., 2003). Selon Lamothe (1981b) et Tremblay (1992a), les roches du Groupe de Bolton représentent des lambeaux ophiolitiques imbriqués dans le Mélange de Saint-Daniel. Doolan et al. (1982), Rickard (1991) et Trottier, Brown et Gauthier (1991) décrivent plutôt ces mêmes roches comme des horizons volcaniques interstratifiés dans les ardoises du Saint-Daniel.

4) La *séquence de la colline Bunker* affleure entre les lacs Massawippi et Memphrémagog. Elle peut être divisée en un membre sédimentaire et un membre volcano-sédimentaire (fig. 1.6; Tremblay, 1990; Blais, 1991). Le membre sédimentaire est représenté par des greywackes feldspathiques souvent comparés aux grès du Groupe de Caldwell de la zone de Humber (De Romer, 1980; Tremblay, 1990; Blais, 1991). Le membre volcano-sédimentaire est constitué de tuf felsique avec des interlits de grès voclanoclastique et de conglomérat volcanique caractérisé par la présence de fuchsite et de serpentine détritique (Tremblay, 1990; Blais, 1991). Ce dernier membre est recoupé par des intrusions mafiques carbonatisées (Tremblay, 1990). D'après Tremblay (1990), le membre volcano-sédimentaire représente la base de la séquence et le membre sédimentaire le sommet. Toujours d'après Tremblay (1990), les roches de la séquence de la colline Bunker affleurent à la faveur d'un synclinal antiforme qui passe à un antiforme anticlinal dans le Groupe de Magog, suggérant ainsi la présence d'une discontinuité majeure entre ces deux unités. Tremblay (1992a) propose que la séquence de la colline Bunker représente une fenêtre structurale exposant des roches cambriennes de type Humber au sein de la zone de Dunnage. À l'instar de De Romer (1980), Tremblay (1990) inclut la séquence de la colline Bunker dans le Mélange de Saint-Daniel, tandis que Slivitsky et St-Julien (1987) attribuent le membre volcano-sédimentaire au Groupe de Magog. D'autres unités de type Humber et de type Magog sont également présentes dans la région de la Beauce et ont été considérées dans l'étude (Cousineau, 1990).

Il ressort de ces travaux antérieurs que le Mélange de Saint-Daniel est une unité particulièrement importante de part son étendue géographique, mais surtout problématique en raison de sa complexité, de la variété des types lithologiques qui y sont rencontrés et de la nature des contacts qui y ont été décrits. Les travaux de Schroetter (2004) dans la région de Thetford-Mines sont, en ce sens, très pertinents puisqu'ils ont permis d'éclaircir les relations tectono-stratigraphiques et temporelles entre les ophiolites et le Mélange de Saint-Daniel. Schroetter et al. (2006) décrivent une unité de brèche polygénique marquant localement une discordance angulaire entre les roches ophiolitiques et les faciès gréso-pélitiques du Mélange de Saint-Daniel. Les brèches qui caractérisent cette unité sont constituées de fragments de roches ophiolitiques (croûte et manteau), sédimentaires et métamorphiques (micaschiste et quartzite). Des datations $^{40}Ar/^{39}Ar$ sur des muscovites de certains fragments métamorphiques ont livré des âges entre ca. 467 et 463 Ma, qui correspondent à ceux obtenus sur la faciès métasédimentaires de la semelle métamorphique infraophiolitique et de la zone de Humber interne dans la région de Thetford-Mines (469-457 Ma; Castonguay et al., 2001; Tremblay, Ruffet et Bédard, 2011). Ceci indique que le Mélange de Saint-Daniel est une unité syn-orogénique qui résulte de l'érosion des roches ophiolites et infraophiolitiques, et qui a été déposée sur l'ophiolite, elle-même alors en obduction (Schroetter et al., 2006; Tremblay, Ruffet et Bédard, 2011). Cette conclusion a donc des répercussions sur l'interprétation de l'évolution tectonique et stratigraphique de la zone de Dunnage et sert de prémice aux travaux présentés dans cette thèse.

1.2.1.3 Le Groupe de Magog

Le Groupe de Magog est une séquence sédimentaire flyschique d'une épaisseur d'environ 10 km exposée au cœur du synclinorium de Saint-Victor. La Formation de Saint-Victor constitue l'unité sommitale du Groupe de Magog et se présente sous la forme d'une épaisse (~ 7 km) séquence de turbidites gréso-pélitiques (St-Julien, 1987; Cousineau et St-Julien, 1994). La partie inférieure du Groupe de Magog est composée de roches volcanoclastiques, de tuff et d'ardoise non-différenciés appartenant à la Formation de Beauceville (Slivitzky et St-Julien, 1987; St-Julien, 1987) mais qui sont subdivisés en trois formations distinctes dans la région de la Beauce, les formations de Frontière, Etchemin et Beauceville (Cousineau et St-Julien, 1994). Les formations de Frontière et d'Etchemin sont principalement constituées de grès volcanoclastique et de tuf felsique, respectivement, tandis que la Formation de Beauceville est formée d'ardoise graphitique, de roches volcanoclastiques et de tuf. Bien que le contact inférieur du Groupe de Magog soit une discordance entre les régions d'Orford et de Thetford-Mines (St-Julien et Hubert, 1975; Doolan et al., 1982; Schroetter et al., 2006), celui-ci est considéré comme une faille majeure dans la région de la Beauce (Cousineau et St-Julien, 1992, 1994). La présence de graptolites dans les formations de Beauceville et de Saint-Victor lui confère un âge Llanvirnien à

Caradocien (Riva 1974; Cousineau et St-Julien, 1994). Les relations stratigraphiques entre chacune des formations du Groupe de Magog sont relativement bien contraintes, mais un lien stratigraphique ou structural avec le Mélange de Saint-Daniel et les unités dites de type Magog qui affleurent au sein de ce dernier sont à préciser.

1.2.1.4 Le Complexe d'Ascot

Le Complexe d'Ascot est constitué de roches volcaniques mafiques et felsiques, de granitoïde syn-volcanique, de tuf felsique et de schiste argileux, qui sont collectivement interprétés comme les vestiges d'un arc volcanique. (figs. 1.3a, 1.7; Tremblay, Hébert et Bergeron, 1989; Tremblay, 1992a). Les schistes argileux ont généralement une structure bréchique à laminée et sont corrélés avec celles du Mélange de Saint-Daniel (Tremblay et St-Julien, 1990). Les roches volcaniques felsiques du Complexe d'Ascot ont donné des âges U-Pb de 441 +7/-12 Ma et 460 ± 3 Ma (David et Marquis, 1994), tandis qu'un granitoïde a donné un âge $^{40}Ar/^{39}Ar$ sur muscovite de 462 ± 2 Ma (Tremblay, Ruffet et Castonguay, 2000). La composition géochimique des laves suggère un environnement de supra-subduction (Tremblay, Hébert et Bergeron, 1989), alors que la signature isotopique et l'abondance de zircons hérités dans les unités felsiques indiquent qu'il a été formé à proximité de Laurentia (David et Marquis, 1994; Tremblay et al., 1994).

1.3 Méthodologie

Du point de vue méthodologique, la problématique de cette thèse a été abordée en combinant des travaux de cartographie géologique aux méthodes de datation U-Pb et $^{40}Ar/^{39}Ar$.

1.3.1 Cartographie géologique

La plupart des feuillets SNRC de la région de l'Estrie-Beauce ont été cartographiés au 1 : 20 000 ou 1 : 50 000, une cartographie systématique n'est donc évidemment pas nécessaire. Les travaux de terrain ont plutôt été effectués de façon ciblée dans le but de résoudre les problématiques propres à chacunes des unités et ensembles étudiés. Comme la plupart des secteurs d'intérêt présentent des évidences de déformation polyphasée, un relevé systématique des éléments structuraux a été effectué. Certaines zones présentant une stratigraphie et/ou des éléments structuraux complexes ont été cartographiées au 1 : 20 000 en incluant les données acquises lors de relevés géologiques antérieurs. Les sites où des échantillons ont été prélevés pour des fins de datation isotopique, ainsi que certaines zones d'intérêt particulier ont été cartographiées de façon détaillée à une échelle variant de 1 : 500 à 1 : 2 000. L'essentiel des données cartographiques sont synthétisées dans les cartes présentées en annexe, ainsi que dans les manuscrits des chapitres II, III et IV.

1.3.2 Géochronologie

Les méthodes de datation isotopique utilisées, ont été choisies en fonction des matériaux et des phénomènes géologiques présents sur les différents terrains étudiés. La méthode $^{40}Ar/^{39}Ar$, avec des températures de fermeture de 450-500°C pour la muscovite et de 550-600°C pour l'amphibole (cf. Tremblay, Ruffet et Bédard, 2011), est idéale pour dater le refroidissement de roches ignées et métamorphiques, ainsi que des phénomènes métamorphiques de moyenne à basse température. Cette technique d'analyse, dont les résultats sont présentés dans le manuscrit du Chapitre III, a été appliquée aux roches ophiolitiques, métamorphiques et granitiques de la région du Lac-Brompton et de la Rivière-des-Plante. La méthode U-Pb permet plutôt d'obtenir l'âge de cristallisation du zircon dans une roche ignée ou dans un environnement métamorphique à des températures élevées, celui-ci ayant une température de fermeture pour le Pb de l'ordre de >900°C (Cherniak et Watson, 2003). La technique d'analyse U-Pb par spectrométrie de masse à source plasma couplée à une source laser (LA-ICP-MS – *laser ablation-inductively coupled plasma-mass spectroscopy*), permet par ailleurs d'effectuer un grand nombre d'analyses dans un délai et à des frais raisonnables (Kosler et Sylvester, 2003). Cette technique se prête donc bien à la datation de zircons détritiques, afin de contraindre la provenance et possiblement l'âge d'unités sédimentaires (Gehrels, 2011; Fedo, Sircombe et Rainbird, 2003), comme de zircons extraits de roches ignées ou métamorphiques (Kosler et Sylvester, 2003). Il en résulte que cet outil est particulièrement utile dans les études paléo-tectoniques, telle que celle présentée dans cette thèse. Des échantillons de roches sédimentaires et volcaniques ont donc été prélevés dans le Groupe de Magog, la séquence de la Colline Bunker et la Ceinture de Gaspé (Groupe de Saint-Francis) afin de compléter la synthèse stratigraphique présentée dans le Chapitre IV. Certains de ces échantillons sélectionnés pour les datations U-Pb sur zircons détritiques se sont révélés riches en muscovites détritiques. Celles-ci ont été analysées par la méthode $^{40}Ar/^{39}Ar$ afin d'en caractériser la source et/ou de documenter un possible événement métamorphique de remise à zéro.

1.4 Liste des contributions

1.4.1 Articles publiés et sous presse

De Souza, S. et A. Tremblay. 2010. « The Rivière-des-Plante ultramafic Complex, southern Québec : Stratigraphy, structure and implications for the Chain Lakes massif ». Dans *From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region*, R. Tollo, J. Bartholomew, J. Hibbard et P. Karabinos, p. 123-140. Geological Society of America, Memoir 206.

De Souza, S., A. Tremblay, G. Ruffet et N. Pinet. 2012. « Ophiolite obduction in the Quebec Appalachians, Canada – 40Ar/39Ar age constraints and evidence for syn-tectonic erosion and sedimentation ». Canadian Journal of Earth Sciences (sous presse).

1.4.2 Résumés de conférences

Tremblay, A., N. Pinet, S. De Souza, G. Ruffet, et S. Castonguay. 2008. « The suture between Laurentia and peri-Laurentian oceanic terranes of the Québec Appalachians – New insights from ophiolitic mélanges in Gaspé Peninsula and southern Québec ». Association géologique du Canada- Association minéralogique du Canada, réunion annuelle, Québec (QC). Résumés, vol. 33, p. 171-172.

De Souza, S., A. Tremblay, et S. Castonguay. 2008. « The Rivière-des-Plante ophiolitic Mélange, southern Québec Appalachians - A tectonic mélange or a dismembered and eroded ophiolite? ». Geological Society of America, northeastern section meeting, Buffalo (NY). Abstracts with programs, vol. 40, no. 2, p. 24.

De Souza, S., et A. Tremblay. 2009. « The Rivière-des-Plante ophiolitic Mélange, southern Québec : Origin and implications for the Chain Lakes massif ». Geological Society of America, northeastern section meeting, Portland (ME). Abstracts with programs, vol. 41, no. 3, p. 4.

De Souza, S., et Tremblay, A. 2010. « Le Complexe ultramafique de la Rivière-des-Plante, Appalaches du sud du Québec : Implications pour l'origine du massif de Chain Lakes ». Congrès des étudiants du GEOTOP, Université McGill, Montréal, Québec.

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De Souza, S., A. Tremblay, A. Meshi, et G. Ruffet. 2010. « The Mélange – Ophiolite conundrum, insights from Iapetan and Tethyan ophiolite belts, Canada and Albania ». Tectonic crossroads: Evolving orogens of Eurasia-Africa-Arabia. Geological Society of America, specialty meeting, Ankara, Turquie. Abstracts with programs, p. 45.

De Souza, S., A. Tremblay, A. Meshi. 2011. « The Mélange – Ophiolite conundrum, insights from Iapetan and Tethyan ophiolite belts, Canada and Albania ». Congrès des étudiants du GEOTOP, Université McGill, Montréal, Québec.

1.4.3 Autres documents

De Souza, S. et Tremblay, S. (soumis). « Compilation géologique, région des Appalaches. Ministère des ressources naturelles du Québec, 21 cartes géologiques 1 :50000, feuillets SNRC 21E14-15-16, 21K13, 21L01-02-03-06-07-08-09-10-11-14-15-16, 21M01-02-08, 21N04-05 ».

CHAPITRE II

THE RIVIÈRE-DES-PLANTE ULTRAMAFIC COMPLEX, SOUTHERN QUEBEC : STRATIGRAPHY, STRUCTURE, AND IMPLICATIONS FOR THE CHAIN LAKES MASSIF

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De Souza, S., et A. Tremblay. 2010. « The Rivière-des-Plante ultramafic Complex, southern Quebec : Stratigraphy, structure and implications for the Chain Lakes massif ». Dans *From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region*, R.P. Tollo, J. Bartholomew, J. Hibbard et P. Karabinos, p. 123-140. Geological Society of America, Memoir 206.

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ABSTRACT

The Rivière-des-Plante ultramafic Complex (RPUC) lies along the Baie Verte-Brompton line in southern Quebec and was previously interpreted as an ophiolitic mélange. It is bounded on the northwest by a northwest-dipping thrust fault and unconformably overlain by conglomerates belonging to the Saint-Daniel Mélange to the southeast. It comprises harzburgite, serpentinite, ophicalcite, gabbro, granite and granofelsic to mylonitic fragmental rocks. The latter have been interpreted as "exotic" metasedimentary rocks correlative with those of the Chain Lakes massif of western Maine. Our mapping suggests that the RPUC is not a mélange, but rather a deeply-eroded ophiolitic remnant mostly represented by mantle peridotites that correlate with those of the Thetford-Mines ophiolite. The granofelsic to mylonitic rocks represent xenolith-bearing granitoids crosscutting the peridotites, rather than "exotic" blocks derived from the Chain Lakes massif. These granites are similar to ca. 470 Ma peridotite-hosted granitoids of the Thetford-Mines ophiolite, which were generated by anatexis of the Laurentian margin during ophiolite obduction. The comparison of metasedimentary rocks of the Chain Lakes massif with those of the southern Quebec Laurentian margin, as well as stratigraphic and geochronological data for both the southern Quebec and western Maine Appalachians, suggests that the Chain Lakes likely represents more or less in-situ Laurentian margin, and that metamorphism and anatexis dated at 469 Ma, may have been caused by the obduction of the southern Quebec ophiolites.

2.1 Introduction

The Chain Lakes massif straddles the Quebec – Maine border and is distinguished from surrounding Ordovician mafic-ultramafic rocks, mélange and flysch of the Boundary Mountains anticlinorium (fig. 2.1) by its high-grade metamorphism and characteristic fragmental texture. Although many interpretations have been proposed regarding its origin, the Chain Lakes massif still remains problematic in most tectonic models of the New England and southern Quebec Appalachians. In many paleogeographic reconstructions and lithotectonic maps, the Chain Lakes massif is shown as part of an exotic Precambrian terrane (Williams and Hatcher, 1983; Boone and Boudette, 1989) or a microcontinent rifted away from Laurentia (Waldron and van Staal, 2001; Gerbi et al. 2006, Gerbi, Johnson and Aleinikoff, 2006; van Staal et al., 2007) or part of Gondwana (Zen, 1983; Kusky, Chow and Bowring, 1997) in Lower Paleozoic times. Pinet and Tremblay (1995b) suggested, on the basis of regional correlations of Chain Lakes-type terranes and adjacent Ordovician ophiolites, mélanges and flysch of western Maine with those of southern Quebec, that the Chain Lakes massif represents metasedimentary rocks deposited on Laurentia and overthrust by ophiolitic nappes during the Taconic orogeny. In southern Quebec, Chain Lakes-type rocks occur in the Rivière-des-Plante ultramafic Complex (RPUC; previously known as the Rivière-des-Plante ophiolitic Mélange; fig. 2.2), and consist of granofelsic to mylonitic fragmental rocks which have been interpreted as metasedimentary rocks correlative with those of the Chain Lakes massif (Cousineau, 1991; Trzcienski, Rodgers and Guidotti, 1992).

This contribution presents new field data on the RPUC and better documents its structural and stratigraphic characteristics, as well as its relationships with surrounding rock units. It is shown that the RPUC represents a deeply-eroded ophiolitic basement over which was deposited the Saint-Daniel Mélange. The Chain Lakes-type granofelsic to mylonitic rocks of the RPUC are xenolith-bearing granitoids intruding ophiolitic mantle peridotites considered to be correlative with the mantle sequence of the Thetford-Mines ophiolite (fig. 2.1). As for the peridotite-hosted granites of the Thetford-Mines ophiolite (Whitehead, Dunning and Spray, 2000), the RPUC granitoids are attributed to the partial melting of the overthrusted Laurentian continental margin rocks during ophiolite obduction. On a larger scale, the Chain Lakes massif is interpreted as a structural inlier exposing the vestiges of the Laurentian margin that was overridden by the southern Quebec and western Maine ophiolites and migmatized during obduction. It is suggested that the Chain Lakes massif does not correlate with the RPUC but more likely represents the crustal melt source from which were formed peridotite-hosted granitoids crosscutting both the RPUC and the Thetford-Mines ophiolite.

2.1.1 The southern Quebec Appalachians

The southern Quebec Appalachians consist of two lithotectonic assemblages of Cambrian-Ordovician rocks, the Humber and Dunnage zones (fig. 2.1; Williams, 1979). The Humber zone corresponds to the remnants of the Laurentian continental margin, whereas the Dunnage zone forms the vestiges of a peri-Laurentian oceanic domain (Notre Dame subzone; Williams, 1979; Williams, Colman-Sadd and Swinden, 1988). Both the Humber and Dunnage zones correspond, respectively, to the Laurentian and Iapetan realms of Hibbard, van Staal and Rankin (2007).

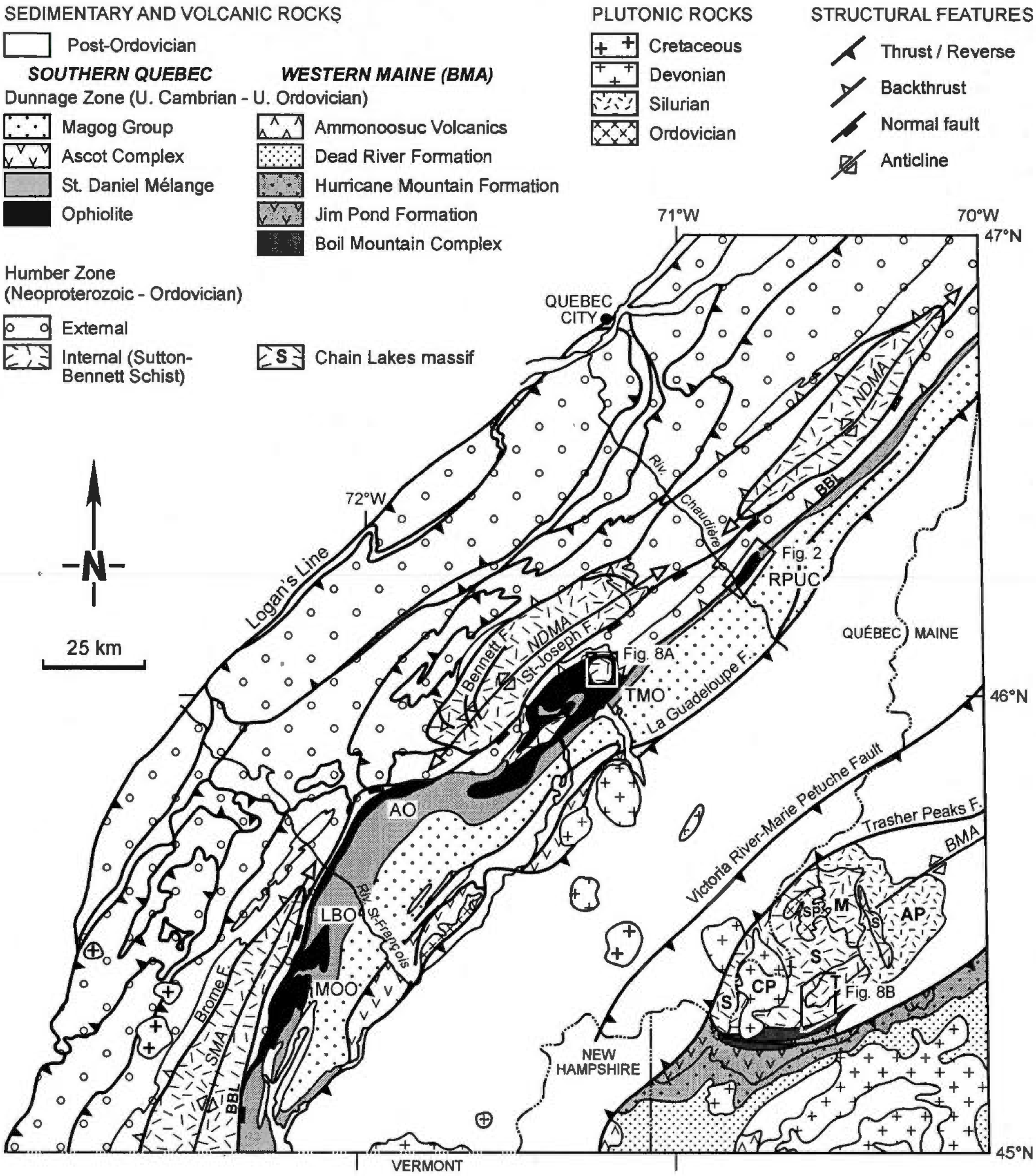

Figure 2.1 : Geological map of the southern Quebec and western Maine Appalachians (after Tremblay and Castonguay, 2002; Moench and Aleinikoff, 2002; Hibbard et al., 2006; Gerbi, 2005). BBL – Baie Verte-Brompton line; BMA – Boundary Mountains anticlinorium; SMA – Sutton Mountains anticlinorium; NDMA – Notre-Dame Mountains anticlinorium; MOO – Mont-Orford ophiolite; LBO – Lac-Brompton ophiolite; AO – Asbestos ophiolite; TMO – Thetford-Mines ophiolite; RPUC – Rivière-des-Plante ultramafic Complex; SP – Skinner pluton; AP – Attean pluton; CP – Chain of Ponds pluton; Chain Lakes massif : S – Sarampus Falls facies; T – Twin Bridges facies; M – McKenney Stream facies.

The boundary between the Humber and Dunnage zones is the Baie Verte – Brompton line (fig. 2.1), which is a fault zone delineated by dismembered ophiolites, mélanges and serpentinite slices (Williams and St-Julien, 1982; Tremblay and Castonguay, 2002).

The Humber zone is divided into an external and internal zone consisting of very low-grade sedimentary and volcanic rocks, and greenschist- to amphibolite-facies metamorphic rocks (the Sutton-Bennett Schist on fig. 2.1; Tremblay and Pinet, 1994; Tremblay and Castonguay, 2002), respectively. Metasedimentary rocks belonging to the internal Humber zone are locally exposed southeast of the Baie Verte – Brompton line, as structural inliers beneath the Thetford-Mines ophiolite (fig. 2.1). In the study area, the external Humber zone is represented by the Caldwell Group, which is mostly made up of feldspathic sandstone and slate, and tholeiitic basalt (St-Julien, 1987; Cousineau, 1990). Sandstone petrography and basalt geochemistry suggest that the Caldwell Group formed during the final stages of Iapetus rifting and that it represents Laurentian continental slope and/or rise facies deposits (Tawadros, 1977; Cousineau, 1990; Bédard and Stevenson, 1999).

In southern Quebec, the Dunnage zone comprises: 1) the Southern Quebec Ophiolite Belt, 2) the Saint-Daniel Mélange, 3) the Magog Group, and 4) the Ascot Complex (fig. 2.1).

The Southern Quebec Ophiolite Belt is represented by four ophiolitic massifs, the Thetford-Mines, Asbestos, Lac-Brompton and Mont-Orford ophiolites (fig. 2.1). U-Pb zircon geochronology on plagiogranites from the Thetford-Mines and Mont-Orford ophiolites yielded ages of 478 $^{+3}/_{-2}$-480 ± 2 Ma (Whitehead, Dunning and Spray, 2000) and 504 ± 3 Ma (David and Marquis, 1994), respectively. The onset of ophiolite obduction onto Laurentia is locally constrained by $^{40}Ar/^{39}Ar$ dating of amphiboles from the dynamothermal metamorphic sole of the Thetford-Mines ophiolite, which yielded an age of 477 ± 5 Ma (Whitehead, Reynolds and Spray, 1995). The mantle peridotites of the Thetford-Mines, Asbestos and Lac-Brompton ophiolites are crosscut by anatectic granitoids dated at 470 $^{+5}/_{-3}$ and 469 ± 4 Ma in the Thetford-Mines area (Whitehead, Dunning and Spray, 2000). These are referred to as peridotite-hosted granitoids and have been attributed to the partial melting of continental margin sedimentary rocks during obduction, as indicated by their peraluminous composition, the occurrence of Precambrian inherited zircons and the low $^{208}Pb/^{206}Pb$ ratios of the dated zircons (Whitehead, Dunning and Spray, 2000). The Mont-Orford ophiolite possibly represents a volcanic arc that rifted to form the younger Thetford-Mines, Asbestos and Lac-Brompton ophiolites (De Souza et al., 2008), which are interpreted as different pieces of the same ophiolitic nappe (Schroetter, Tremblay and Bédard, 2005; De Souza et al., 2008). In addition to ophiolitic massifs, ultramafic rocks and serpentinites occur disparately along the Baie Verte – Brompton line, some of which bearing more or less "exotic" rock-types. In southern Quebec, the Rivière-des-Plante ultramafic Complex, which was previously interpreted as an ophiolitic mélange (Cousineau, 1991), is by far the largest of these ultramafic bodies. It is located 40 km to the NE of the Thetford-Mines ophiolite (fig. 2.1) and consists of peridotites, serpentinites, ophicalcite, gabbro, granitic rocks, and a fragmental granofelsic to mylonitic rock that has been interpreted as correlative with the Chain Lakes massif (Cousineau, 1991; Trzcienski, Rodgers and Guidotti, 1992).

The Saint-Daniel Mélange is a Llanvirn sedimentary unit that forms the base of a syncollisional sedimentary basin unconformably overlying the southern Quebec ophiolites (Schroetter et al., 2006). Debris flows and conglomerates characterizing the base of the Saint-Daniel were formed during synorogenic uplift and erosion of the ophiolite basement and underlying metamorphic rocks (see Schroetter et al., 2006 and De Souza et al., 2008 for details). The Magog Group (Cousineau and St-Julien, 1994) is 10 km-thick and represents an onlapping Caradoc flysch sequence overlying the Saint-Daniel Mélange. The Ascot Complex is interpreted as the remnant of a 460 ± 3 Ma (David and Marquis, 1994) volcanic arc sequence (Tremblay, Hébert and Bergeron, 1989; Tremblay, Ruffet and Castonguay, 2000) made up of a bimodal metavolcanic rock series, overlain by and in fault contact with laminated and pebbly phyllites that have been correlated with the Saint-Daniel Mélange (Tremblay and St-Julien, 1990).

Schroetter, Tremblay and Bédard (2005) have shown that the Humber zone, the southern Quebec ophiolites and the overlying Saint-Daniel Mélange share a similar structural evolution. An Early to Late Ordovician (480-445 Ma; Castonguay et al., 2001; Tremblay and Castonguay, 2002; Sasseville et al., 2008) S_{1-2} schistosity is associated with ophiolite obduction during the taconic orogeny (Tremblay and Pinet, 1994; Tremblay and Castonguay, 2002; Schroetter, Tremblay and Bédard, 2005). Two generations of post-obduction structures are recognized: 1) D_3 SE-verging Silurian – Early Devonian (430-410 Ma; Castonguay et al., 2001; Sasseville et al., 2008) backthrusts and backfolds that culminated with the formation of steep southeast-dipping normal faults such as the composite St-Joseph – Baie Verte-Brompton line fault (fig. 2.1; Pinet, Castonguay and Tremblay, 1996; Pinet, Tremblay and Sosson, 1996. Tremblay and Castonguay, 2002; Castonguay and Tremblay, 2003) and 2) D_4 Late Devonian NW-verging folds and reverse faults related to the Acadian orogeny (Tremblay and Pinet, 1994; Tremblay and Castonguay, 2002; Schroetter, Tremblay and Bédard, 2005; Sasseville et al., 2008).

2.2 The RPUC and adjacent rock units

The RPUC consists of serpentinized peridotite, ophicalcite, gabbro and granitic rocks (fig. 2.2). The latter rock type is divided into two distinct facies: 1) Type 1 granites, previously described by Cousineau (1991) as "continental" K-rich granitoids, are petrographically and geochemically similar to peridotite-hosted granitoids of the Thetford-Mines, Asbestos and Lac-Brompton ophiolites (Whitehead, Dunning and Spray, 2000; De Souza et al., 2008); 2) Type 2 granites are represented by granofelsic to mylonitic fragmental rocks that were previously mapped as coarse-grained rift-related metasedimentary rocks (Cousineau, 1991; Trzcienski, Rodgers and Guidotti, 1992), but are here re-interpreted as xenolith-bearing granitic intrusions (see below).

To the northwest, the contact between the RPUC and the Caldwell Group is a northwest-dipping thrust fault, whereas the contact with the overlying Saint-Daniel Mélange, to the southeast, is a major unconformity (fig. 2.2). Field data and interpretations presented below are based on geological mapping, geochemical characterization of granitic rocks and detailed observations of three outcrops where critical elements of the RPUC and its relationships with adjacent rock units are exposed, the Rivière-des-Plante, Bras Saint-Victor, and Highway 73 outcrops (see figure 2.2 for location).

2.2.1 Ophiolitic rocks

Serpentinized peridotite, ophicalcite, gabbro and scaly serpentinite are the dominant ophiolitic lithologies of the RPUC. Serpentinized harzburgite is the most common type of peridotite but dunite is locally found. The harzburgite is usually massive but shows slickenlined fractures and shear zones. A tectonic foliation, defined by orthopyroxene porphyroclasts, is locally present.

The overall mineralogy, texture and structure of the harzburgite are typical of mantle peridotites found in ophiolites (Juteau and Maury, 1999). Along the northwestern contact of the RPUC with the Caldwell Group, the peridotites are strongly sheared and locally grade into talc-carbonate schist, whereas to the southeast, at the contact with the Saint-Daniel Mélange, ophicalcite breccias are locally present. These breccias have been described in detail by Lavoie and Cousineau (1995) and attributed to hydraulic fracturing and carbonate cementation proximal to a seafloor hydrothermal vent.

The gabbros are massive, medium-grained and mesocratic, or appear as a mesocratic heterogenous gabbroic breccia showing medium-grained to pegmatitic fragments dispersed in a finer-grained gabbroic matrix. Chloritization and albite-quartz or epidote veins are common. Fine-grained massive diabase also occurs; it shows a subophitic texture and a greenschist-facies mineral assemblage is developed (chlorite, actinolite, albite, quartz).

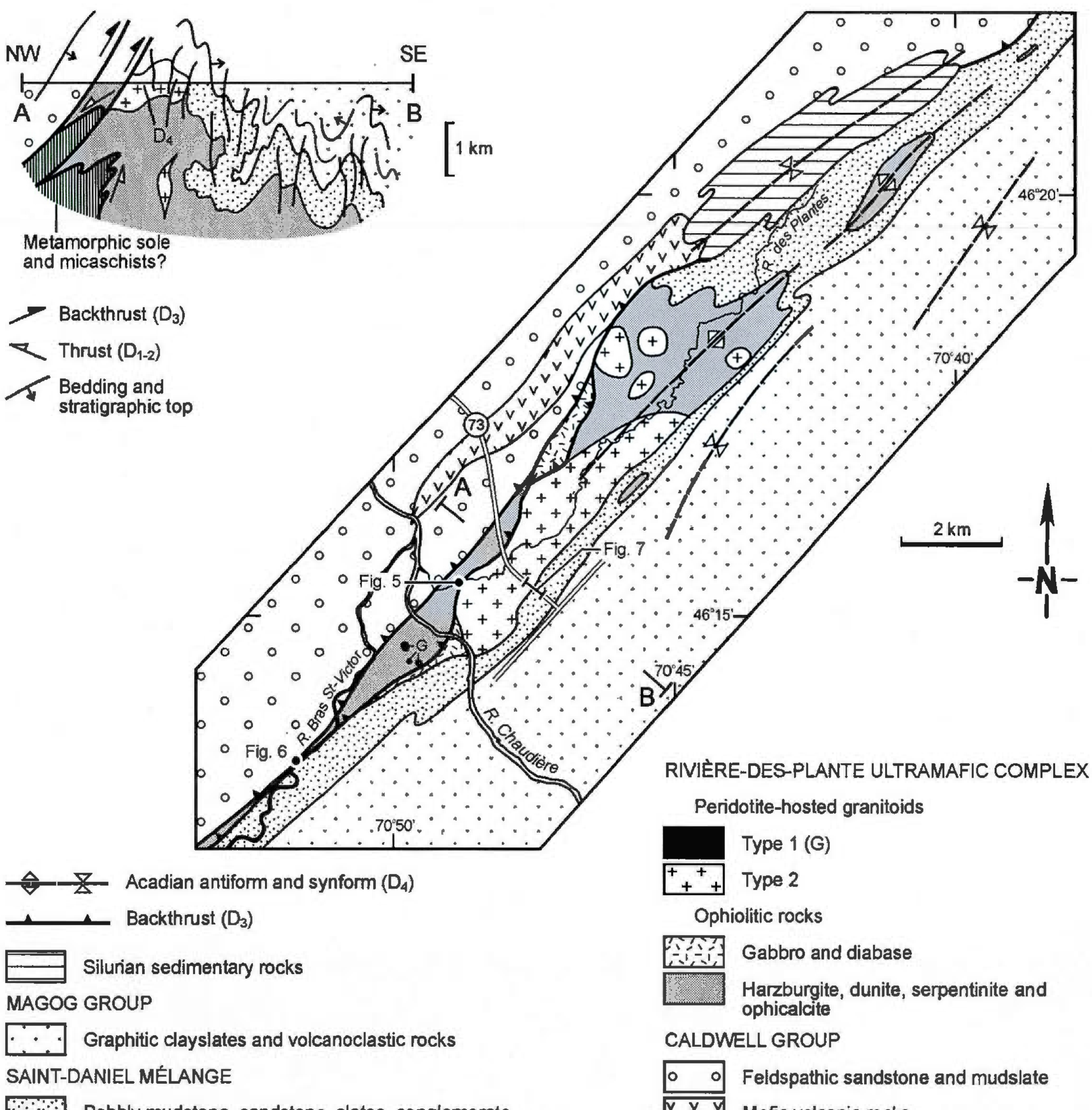

Figure 2.2 : Geological map of the Rivière-des-Plante ultramafic Complex. Geological mapping during this study combined with field data from previous work by St-Julien (1987) and Cousineau (1990). Vertical exaggeration in cross-section is ~ 1.5. See figure 2.1 for location.

2.2.2 Peridotite-hosted granitic rocks

Type 1 granite

Type 1 granites occur as small intrusions (< 100 m) in the southwestern part of the RPUC (fig. 2.2), or as dykes and fault-bounded bodies (< 10 m). They are equigranular to porphyritic, medium to coarse-grained, locally foliated and often rodingitized along their margins. In thin section, these rocks are made up of euhedral to sub-euhedral plagioclase, quartz, K-feldspar, biotite, muscovite and minor zircon, apatite and oxides. Hornblende can be locally abundant. Type 1 granites are peraluminous, corundum-normative (1.45-1.69 vol. %) and have a normative composition of granite *sensu stricto*, as well as high SiO_2 (71 wt. %) and K_2O (3.4-6.4 wt. %) values similar to continental granophyres (Appendix D). They are enriched in Rb, Ba, Ce and depleted in Yb, Y when compared to ocean ridge granites, and have trace element profiles similar to those of the Thetford-Mines ophiolite peridotite-hosted granitoids (fig. 2.3).

Type 2 granite

Xenolith-bearing Type 2 granites were previously interpreted as metasedimentary rocks by Cousineau (1991) and Trzcienski, Rodgers and Guidotti (1992), who also recognized a gneissic mylonitic facies developed at the expense of these rocks. The fragmental texture of Type 2 granite is given by abundant randomly-oriented xenoliths dispersed in a medium-grained felsic matrix (fig. 2.4a). The matrix consists of anhedral K-feldspar, anhedral to euhedral plagioclase, muscovite, reddish-brown biotite, zircon and apatite. Chlorite and sericite aggregates are found as pseudomorphs after cordierite (Cousineau, 1991; Trzcienski, Rodgers and Guidotti, 1992) and garnet has been locally observed. Quartz grains show undulose extinction and are recrystallized in the foliated facies. K-feldspar commonly shows concentric to oscillatory zoning and forms an interlocking texture with quartz and plagioclase (fig. 2.4b). Zircon grains are commonly rounded and interpreted as xenocrysts; some of them display euhedral prismatic overgrowths. U-Pb zircon dating of Type 2 granite yielded ages from 2708 to 571 Ma, with most ages varying between 1640 and 1005 Ma (Dunning and Cousineau, 1990), suggesting that most xenocrysts were derived from a Laurentian source. The matrix penetrates into gneissic xenoliths along foliation planes and 5 to 20 cm wide granitic dykes are common. The oscillatory zoning in K-feldspar, the occurrence of interlocking textures between various mineral species, and the overall mineralogical composition and field relations indicate that this matrix is of igneous origin and crystallized from a granitic melt. As for Type 1 granites, this matrix is peraluminous, corundum-normative (2.5-5.1 vol. %) and has a normative composition of granite *sensu stricto*, as well as high SiO_2 (66-75 wt. %) and K_2O (3.9-5.6 wt %) values (Appendix D). It is enriched in Rb, Ba, Ce and depleted in Yb, Y when compared to ocean ridge granites and shows higher Zr and Hf values relative to peridotite-hosted granitoids from the Thetford-Mines ophiolite (fig. 2.3). The abundance of xenoliths in Type 2 granites, and the higher Zr and Hf values relative to Type 1 granites and peridotite-hosted granites of the Thetford-Mines ophiolite, can be attributed to source heterogeneities, variations in petrogenetic processes during partial melting of the sedimentary protolith, and/or differentiation/assimilation of the granitic magmas.

The proportion of xenoliths commonly ranges between 5-15% and does not exceed 20%. These are subrounded to angular, and generally show a well-developed internal foliation or mylonitic fabric. Their size ranges from < 1 cm to approximately 1 meter-long. They consist of schistose to gneissic amphibolite and abundant metasedimentary rocks. The latter rock type is frequently hornfelsed, in which case the xenoliths show a white to light gray metasomatic aureole (fig. 2.4a). Some of the metasedimentary xenoliths are made up of biotite, plagioclase, quartz, garnet and muscovite with or without sillimanite and/or andalusite. Nodules of polycrystalline quartz are present and range in size from 1 to 3 cm; they can be interpreted as inherited quartz vein fragments as typically found in S-type granites (White, Chappell and Wyborn, 1991). Mica-rich xenoliths (< 1 cm), consisting of biotite and sillimanite, have been observed in thin section. These are similar to surmicaceous enclaves found in anatectic granites and are commonly interpreted as small residues of partial melting of a sedimentary protolith (Didier and Barbarin, 1991; Montel et al., 1991).

A mylonitic facies of the xenolith-bearing Type 2 granite is exposed at the Rivière-des-Plante outcrop (fig. 2.5), where it forms the northwestern margin of a large Type 2 granitic body. From east to west, the Type 2 granite progressively shows grain size reduction and acquires a well-developed gneissic to mylonitic layering hosting southwest-plunging mineral and stretching lineations marked by stretched xenoliths and quartz rods (fig. 2.5b). The xenoliths are progressively transposed into the mylonitic layering.

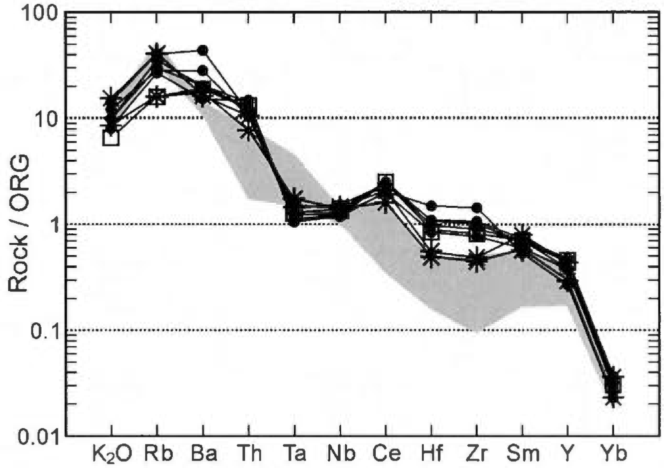

Figure 2.3 : Ocean ridge granite (ORG)-normalized trace element diagram for Type 1 (asterisks), Type 2 (black dots) and mylonitic Type 2 (open squares) granites of the RPC, and peridotite-hosted granitoids of the Thetford-Mines ophiolite (grey compositional field; this study and Whitehead, Dunning and Spray (2000)). ORG values are from Pearce, Harris and Tindle (1984).

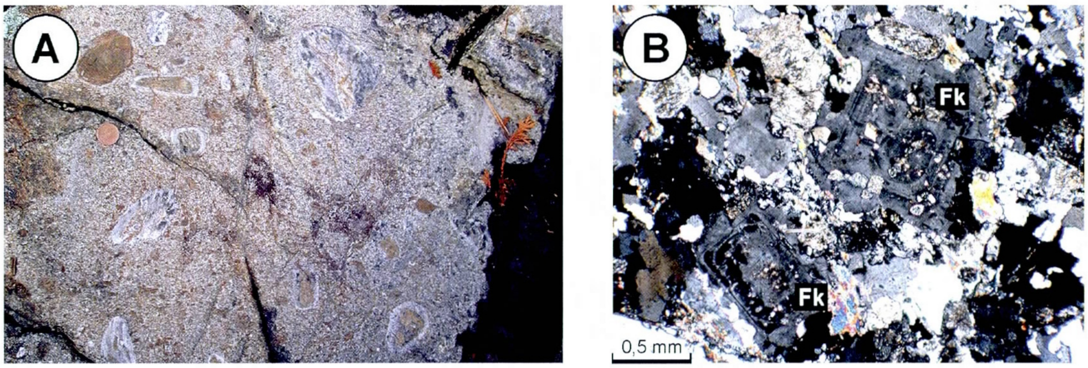

Figure 2.4 : a) Hornfelsed xenoliths in Type 2 granite showing light gray metasomatic aureoles; b) K-feldspar (Fk) showing concentric oscillatory zoning and forming an interlocking texture with quartz and plagioclase.

The granitic mylonite shows a porphyroclastic texture and consists of quartz, plagioclase, K-feldspar, muscovite, biotite, pinitized cordierite, garnet and zircon. Shear bands and asymmetric pressure shadows indicate a right-lateral sense of shear (fig. 2.5c). Within 10-15 meters from the contact with serpentinite to the west, the granitic mylonite is rodingitized, its "primary" mineralogy is progressively modified into an assemblage of prehnite, chlorite, grossular, zoisite and actinolite. The rodingitized mylonite is folded and overprinted by a northwest-dipping cleavage. The serpentinite adjacent to the mylonite consists of lens-shaped to sub-rounded harzburgite blocks (1m to < 10 cm) with relict primary structures dispersed in a strongly lineated scaly serpentinite matrix. S-C fabrics and slickenlines indicate a reverse, top-to-the-southeast sense of shear (fig. 2.5d).

The origin of the granitic mylonitic rocks has always been problematic in the structural interpretation of the RPUC. According to Trzcienski, Rodgers and Guidotti (1992), mylonitization was possibly associated with major right-lateral shear zones (perhaps with several hundred kilometers of displacement), whereas Cousineau (1991) attributed that mylonitic deformation to a metamorphic event pre-dating the tectonic emplacement into the peridotites. Considering that the RPUC Type 2 granites correlate with ophiolitic peridotite-hosted granitoids, an alternative interpretation would be that mylonitization was more or less synchronous with obduction and cooling of the ophiolitic rocks and granites, as suggested for the Thetford-Mines ophiolite (Whitehead, Dunning and Spray, 2000; Tremblay and Ruffet, 2008). The lack of penetrative deformation and mylonite development in most areas surrounding the contact between the peridotites and xenolith-bearing granites suggests that intrusive contacts are however preserved.

2.2.3 Summary

Field observations and petrography of the granofelsic to mylonitic fragmental rocks belonging to the RPUC, together with their geochemical characteristics, indicate that these are xenolith-bearing granites occurring as small intrusions and dykes crosscutting the peridotites. We suggest that Type 1 and Type 2 granites are correlative with the ca. 470 Ma peridotite-hosted granitoids of the Thetford-Mines ophiolite, which are interpreted as having formed by partial melting of Laurentian sedimentary rocks during ophiolite obduction (Whitehead, Dunning and Spray, 2000), a widely-accepted mechanism for the formation of peraluminous granitic rocks intruding mantle and lower crustal rocks in ophiolites (Pearce, 1989; Cox, Searle and Pedersen, 1999; Searle and Cox, 1999; Li et al., 2008; among others). Such granitoids commonly form by low-pressure anatexis (3-5.5 GPa) of continental margin sedimentary rocks during the obduction of a young (< 10 Ma) oceanic lithosphere (Pearce, 1989; Cox et al., 1999; Whitehead, Dunning and Spray, 2000). Shear heating along the sole thrust and residual heat from the obducting ophiolitic rocks are the inferred heat sources for partial melting of the sedimentary protolith (Pearce, 1989; Whitehead, Dunning and Spray, 2000). Such granitic rocks occur as undeformed to gneissic/mylonitic dykes (1-100 m) and bodies (500 m to > 5 km) and have been reported from many ophiolites around the world (Ledru, 1980; Pearce, 1989; Cox, Searle and Pedersen, 1999; Skjerlie et al., 2000; Whitehead, Dunning and Spray, 2000). Xenoliths are present in some of these granitoids; they commonly derive from the ophiolitic wall-rock, but metamorphic fragments and biotite-rich restitic material (surmicaceous enclaves) are also found (Ledru, 1980; Cox, Searle and Pedersen, 1999). The widespread occurence of inherited zircon xenocrysts in these rocks (Skjerlie et al., 2000; Li et al., 2008), their peraluminous composition and the presence of restitic material and/or migmatitic xenoliths (Ledru, 1980, Cox, Searle and Pedersen, 1999) demonstrate that they represent anatectic melts generated from a sedimentary protolith.

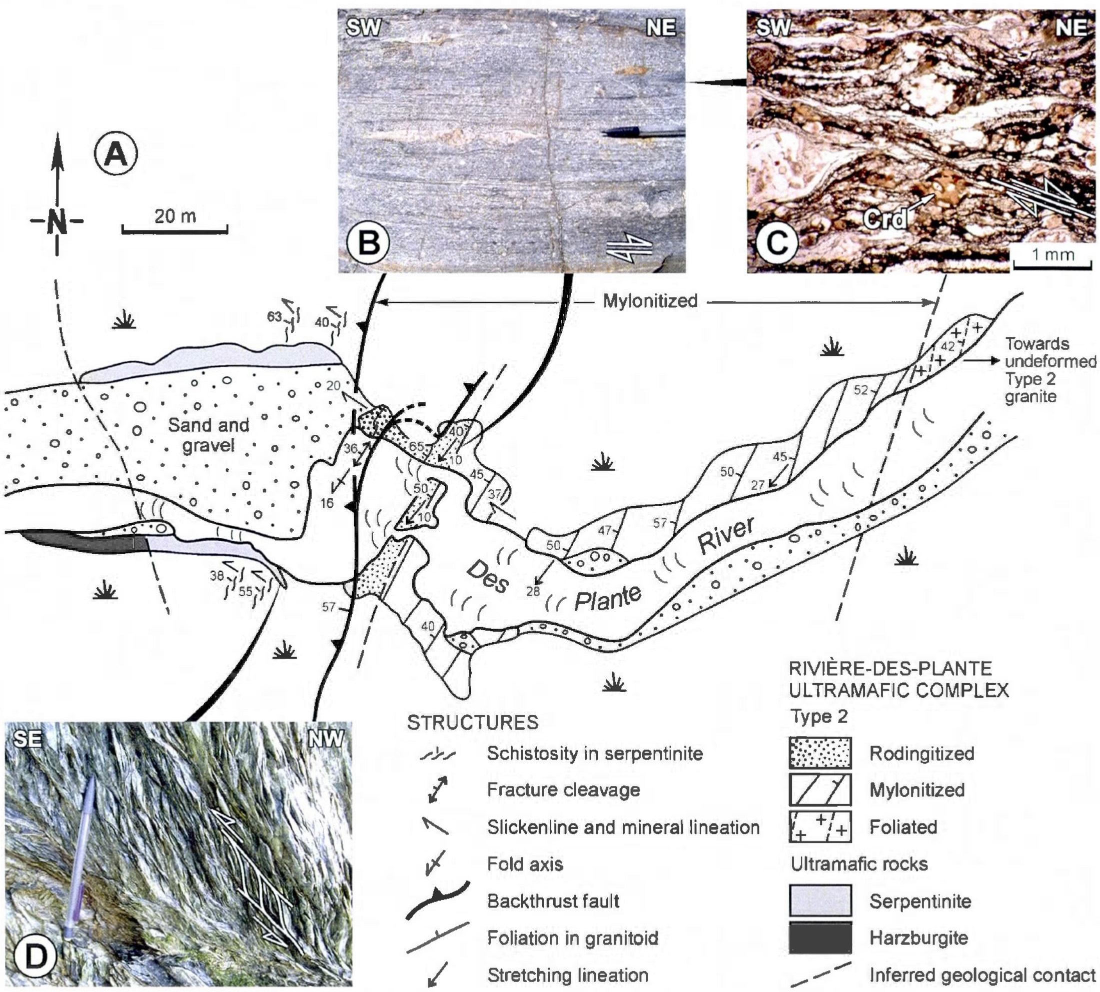

Figure 2.5 : a) Detailed map of the RPUC at the Rivière-des-Plante outcrop; b) Finely laminated Type 2 mylonitic granite. Note the presence of stretched quartz nodules; c) Porphyroclastic texture of the granitic mylonite and shear band indicating a right-lateral sense of shear. Crd – pinitized cordierite; d) S-C fabrics in the serpentinite indicating the top-to-the-southeast thrusting. See text for details and figure 2.2 for location.

2.2.4 The contact with the Caldwell Group

The contact between the RPUC and the Caldwell Group is well-exposed at the Bras Saint-Victor outcrop, where the Caldwell Group is thrust over both the RPUC and the Saint-Daniel Mélange (see detail map; fig. 2.6). At this site, the ultramafic rocks are similar to the blocky serpentinite facies described above for the Rivière-des-Plante outcrop, and grade into a talc-carbonate schist over approximately 5 m towards the contact with a pebbly mudstone of the Saint-Daniel Mélange.

To the northwest, the Caldwell Group crops out in the hanging wall of a southeast-directed thrust fault and consists of green and black sandstone, slate and quartzite (St-Julien, 1987). The bedding is overturned and dips moderately towards the northwest. Slickensided fault surfaces with down-dip lineations are developed in pelitic beds. Hydrothermal breccia, crush breccia, cataclastic shearing and local boudinage structures in the sandstone occur along the contact with the serpentinites, indicating an overall brittle fault deformation. Kinematic indicators such as slickensided surfaces, foliation dragging in the Caldwell Group, and C-S fabrics in the serpentinites suggest a reverse, top-to-the-southeast sense of shear. This fault zone corresponds to the same southeast-directed fault described at the Rivière-des-Plante outcrop (see above) and marks the northwestern limit of the RPUC.

At the Bras Saint-Victor outcrop, the contact between the serpentinites and the Saint-Daniel is reworked by a late, subvertical to steeply northwest-dipping normal fault marked by discontinuous quartz veins. The northwest-dipping foliation found in the pebbly mudstone, the serpentinites and the Caldwell Group is attributed to a Silurian D_3 backthrusting deformational event documented by Tremblay and Pinet (1994). The normal faulting event probably correlates with the late D_3 extensional deformation recognized along the Baie Verte – Brompton line and the Saint-Joseph fault in southern Quebec (Pinet, Tremblay and Sosson. 1996; Tremblay and Castonguay, 2002; Castonguay, Ruffet and Tremblay, 2007).

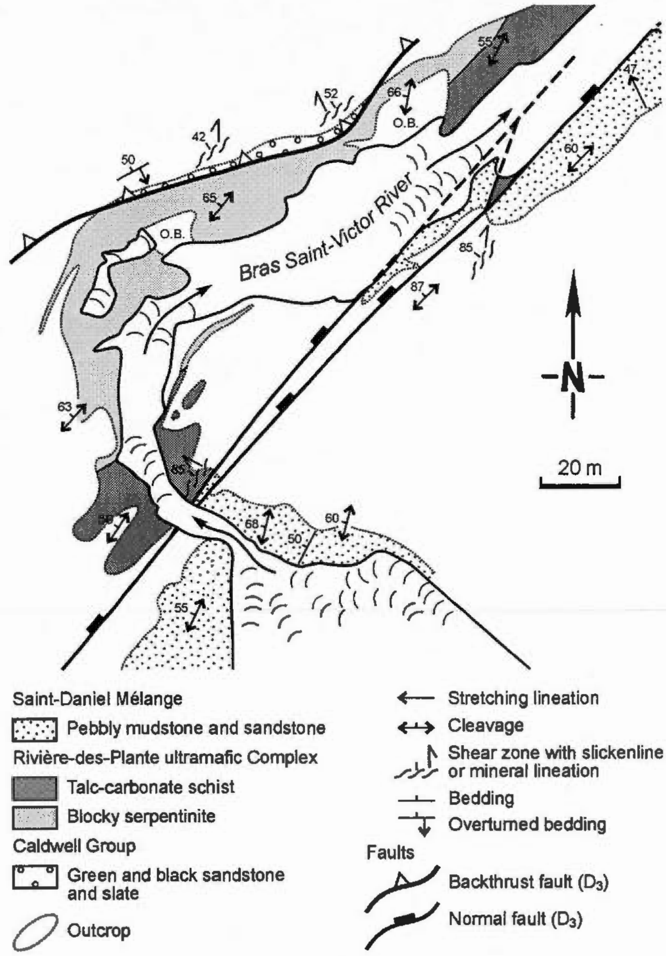

Figure 2.6 : Detailed map of the RPUC at the Bras Saint-Victor outcrop located along the main backthrust fault separating the RPUC – Saint-Daniel Mélange in the footwall, and the Caldwell Group in the hanging wall. See figure 2.2 for location.

2.2.5 The Saint-Daniel Mélange

The Highway 73 outcrop is a ca. 400 meters-long roadcut where the contact between the RPUC and the Saint-Daniel Mélange is exposed (fig. 2.7). Along this roadcut, debris flows belonging to the Saint-Daniel unconformably overlie serpentinite and xenolith-bearing Type 2 granite of the RPUC. The basal debris flows are successively overlain, towards the southeast, by 1) a well-stratified sequence of southeast-dipping turbiditic, green and black, chromite-bearing sandstone and mudstone interbedded with graphitic black slate, 2) a pebbly mudstone unit, 3) serpentinite sandstone and conglomerate, 4) green sandstone and conglomerate with metasedimentary clasts and 5) purple, green and black mudstone interbedded with green sandstone (see fig. 2.7). The

RPUC granite is not foliated but shows a set of regularly-spaced, southeast-dipping slickenlined fractures. The ultramafic rocks are represented by the block-in-matrix and talcose serpentinite facies, but are devoid of major shear structures such as those characterizing the serpentinites of the Rivière-des-Plante and Bras Saint-Victor outcrops. Northwest-directed reverse shear zones are developed along the section and a vertical to steep southeast-dipping cleavage cuts across the bedding. Regional folding is marked by a well-developed southwest-plunging upright asymmetric dextral fold (fig. 2.7a). These structures, which correspond to regional deformation in the Magog Group, were formed during the Acadian orogeny (Cousineau and Tremblay, 1993) and correlate with the D_4 phase of Tremblay and Pinet (1994).

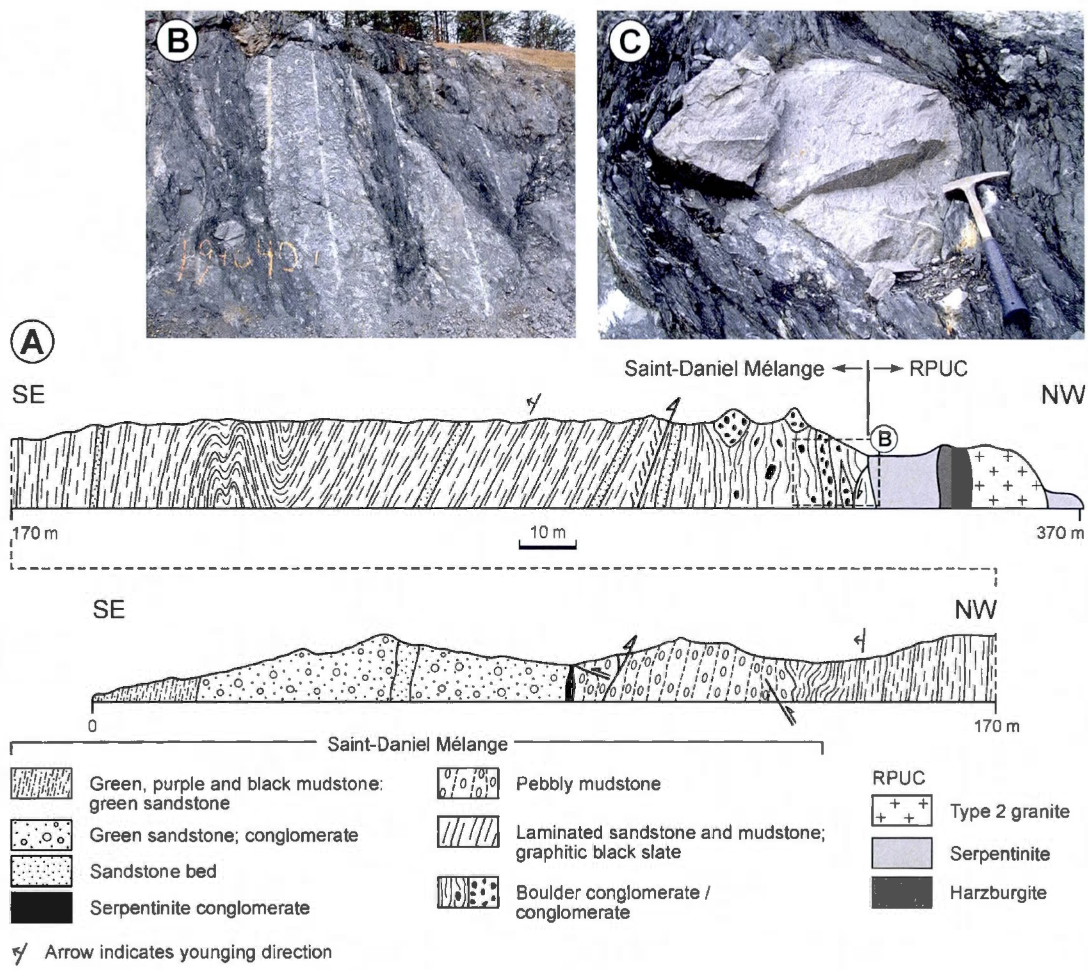

Figure 2.7 : a) Detailed sketch of the Highway 73 roadcut, which exposes the contact between the RPUC and the Saint-Daniel Mélange; b) Basal conglomeratic debris flows of the Saint-Daniel Mélange; c) Boulder of foliated Type 2 granite of the RPUC. Vertical exaggeration in section is ~ 1.5. See text for details and figure 2.2 for location.

The base of the stratigraphic sequence consists of a matrix-supported boulder conglomerate, which mainly consists of sub-rounded to angular blocks up to ca. 2 meters across of foliated and unfoliated Type 2 granite of the RPUC (figs. 2.7b, 2.7c). Detrital chromite grains and clasts of foliated green sandstone, slate and quartzite are found in minor amount. The matrix of these conglomerates shows a penetrative cleavage and the beds are locally boudinaged (fig. 2.7). Similar conglomerates and breccias are found discontinuously in the surrounding area along the contact between the Saint-Daniel and the RPUC, although that grain size is commonly finer (< 25 cm) and metamorphic rocks locally make up to 50% of the fragments (micaschist, quartzite and vein quartz). Such lithological variations have also been documented for correlative conglomeratic and breccia units overlying the Thetford-Mines ophiolite (Schroetter et al., 2006), and interpreted as directly reflecting the different rock-types exposed in source areas. The predominance of xenolith-bearing granites and the occurrence of detrital chromite grains in the Saint-Daniel basal debris flows along the highway 73 roadcut indicate that they were essentially derived from the RPUC. This, combined with the lack of major shear structures at the RPUC – Saint-Daniel contact, clearly indicates that the base of the Saint-Daniel Mélange is a major unconformity as documented in the Thetford-Mines and Lac-Brompton ophiolites (Schroetter et al., 2006; De Souza et al., 2008).

2.2.6 Stratigraphic and structural synthesis

The RPUC is characterized by serpentinized peridotites, ophiolitic gabbros, and by the occurrence of xenolith-bearing peridotite-hosted granitoids similar to those of the Thetford-Mines ophiolite. The ultramafic rocks vary from massive harzburgite and dunite to strongly sheared serpentinite along the contact with the Caldwell Group. The massive peridotite facies occurs throughout the RPUC and locally shows a typical seafloor hydrothermal alteration marked by the occurrence of ophicalcites towards the contact with the Saint-Daniel Mélange. The RPUC – Saint-Daniel contact is a major unconformity overlain by discontinuous horizons of debris flows mostly derived from the recycling of the RPUC lithologies.

The RPUC shows a polyphase structural history, which partially obscures its original ophiolitic stratigraphy and relationships with adjacent rock units. A comparison with the well-established structural history of the southern Quebec Appalachians (Tremblay and Castonguay, 2002), suggests that three deformational episodes have contributed to the structural framework of the

RPUC: 1) the obduction of oceanic lithosphere during the taconic orogeny (D_{1-2}); 2) Silurian – Early Devonian backthrusting and normal faulting (D_3); and 3) reverse faulting and folding during the Acadian orogeny (D_4).

2.3 The Chain Lakes massif and adjacent rock units

In the light of this new stratigraphic framework and interpretations regarding the origin of the RPUC and of its Chain Lakes-type fragmental granofelsic rocks, it is appropriate to highlight the stratigraphic and stuctural features of the Chain Lakes massif and adjacent rock units at the type-locality, and discuss an alternative hypothesis regarding its possible correlation in the southern Quebec Appalachians.

In western Maine, the Chain Lakes massif comprises three mappable units of more or less migmatized and fragmental quartzofeldspathic metasedimentary rock facies (see fig. 2.1). Fragments in the Chain Lakes massif mostly consist of amphibolite, felsic metavolcanic rocks, calcsilicate, metapelite and quartz nodules (Boudette et al., 1989; Gerbi, Johnson and Aleinikoff (2006). On the basis of detrital U-Pb zircon ages, Cousineau (1991) and Gerbi, Johnson and Aleinikoff, 2006) concluded that the Chain Lakes massif derived from the erosion of Laurentian continental crust. High-grade metamorphism and associated migmatization in the Chain Lakes massif are of Middle Ordovician age (469 ± 4 Ma), and peak metamorphic conditions have been estimated to be of upper amphibolite facies (750°C – 800°C and 4-5 GPa; Gerbi, Johnson and Aleinikoff, 2006). Early Ordovician U-Pb zircon ages were also obtained on a felsic metavolcanic rock fragment (479 ± 6 Ma), on zircon overgrowths from an amphibolite lens (486 ± 11 Ma) and metasedimentary rocks (483 ± 8 Ma; 477 ± 6 Ma) of the Chain Lakes massif (Gerbi et al., 2006; Gerbi, Johnson and Aleinikoff, 2006). Based on zircon morphology and Th/U ratios, the ages yielded by the amphibolite and the felsic metavolcanic rock were interpreted as crystallization ages, whereas those obtained on the metasedimentary rocks were interpreted as metamorphic ages (Gerbi et al., 2006; Gerbi, Johnson and Aleinikoff, 2006). Since these ages correspond more or less to crystallization ages yielded by the Boil Mountain Complex and Jim Pond Formation (see below), subduction beneath the Chain Lakes massif and arc volcanism were invoked to account for the origin of the amphibolite and the felsic metavolcanic rock, as well as metamorphic zircon growth in the metasedimentary rocks (Gerbi et al., 2006; Gerbi, Johnson and Aleinikoff, 2006). However, the significance of these Early Ordovician ages is debatable; they may also be the result of high-grade metamorphism and/or migmatization of the Chain Lakes massif, rather than arc magmatism.

The Boil Mountain Complex lies south of the Chain Lakes massif; it is a thin (<3 km) unit of ultramafic and mafic plutonic rocks and plagiogranite that is overlain by mafic and felsic volcanic rocks of the Jim Pond Formation (Boudette, 1982; Coish and Rogers, 1987; Kusky, Chow and Bowring, 1997). U-Pb zircon geochronology on the Boil Mountain Complex and Jim Pond Formation yield Early Ordovician ages of 477 $^{+7}/_{-5}$ Ma (Kusky, Chow and Bowring, 1997; Gerbi et al., 2006) and 484 ± 5 Ma (Moench, Aleinikoff and Boudette, 2000), respectively. Gerbi et al. (2006) argued that the Boil Mountain Complex is not an ophiolite as previously suggested (e.g. Boudette, 1982; Coish and Rogers, 1987; Kusky, Chow and Bowring, 1997), but rather represents an intrusion crosscutting both the Jim Pond Formation and the Chain Lakes massif, an equivocal interpretation with which we disagree. Although the Boil Mountain Complex and the Jim Pond Formation may not be entirely co-genetic (Gerbi et al., 2006), we agree with the previous interpretations that the Boil Mountain Complex forms part of an ophiolite, and that both units were formed in the same suprasubduction zone oceanic basin.

Different interpretations relating the Chain Lakes massif with Laurentia or Gondwana have also been proposed over the last few decades (Marleau, 1968; Boudette, 1982; Zen, 1983; Boone and Boudette, 1989; Moench and Aleinikoff, 2002; Gerbi et al., 2006, Gerbi, Johnson and Aleinikoff, 2006; among others). Trzcienski, Rodgers and Guidotti (1992) and Pinet and Tremblay (1995b) suggested that it represents an inlier of the Laurentian continental margin over which were obducted the southern Quebec and western Maine ophiolites. We would like here to emphasize such an interpretation by briefly comparing the Bécancour dome of southern Quebec, an infra-ophiolitic metasedimentary sequence of the Thetford-Mines area (fig. 2.8a), with the Chain Lakes massif of western Maine. The Bécancour dome is a structural inlier through the Thetford-Mines ophiolite where greenschist-grade gneisses and quartzitic schists (Birkett, 1981) of Laurentian affinity are exposed. Metamorphism and deformation in the Bécancour dome are basically related to ophiolite obduction during the taconic orogeny at ca. 460-470 Ma (Castonguay et al., 2001; Tremblay and Castonguay, 2002). These metasedimentary rocks are similar to the Sarampus Falls facies of the Chain Lakes massif (Gerbi, personal communication, 2008). As for the Chain Lakes massif, the penetrative foliation marking the compositional layering is affected by a late sub-vertical cleavage, and defines a dome-like structure with down-dip stretching lineations (figs. 2.8a, 2.8b; Tremblay and Pinet, 1994; Gerbi, Johnson and Aleinikoff, 2006). These structural and lithological similarities suggest that both rock units may have undergone a similar geodynamic evolution and that, as suggested by Trzcienski, Rodgers and Guidotti (1992) and Pinet and Tremblay (1995b), the Chain Lakes massif likely represents an inlier of the Laurentian margin.

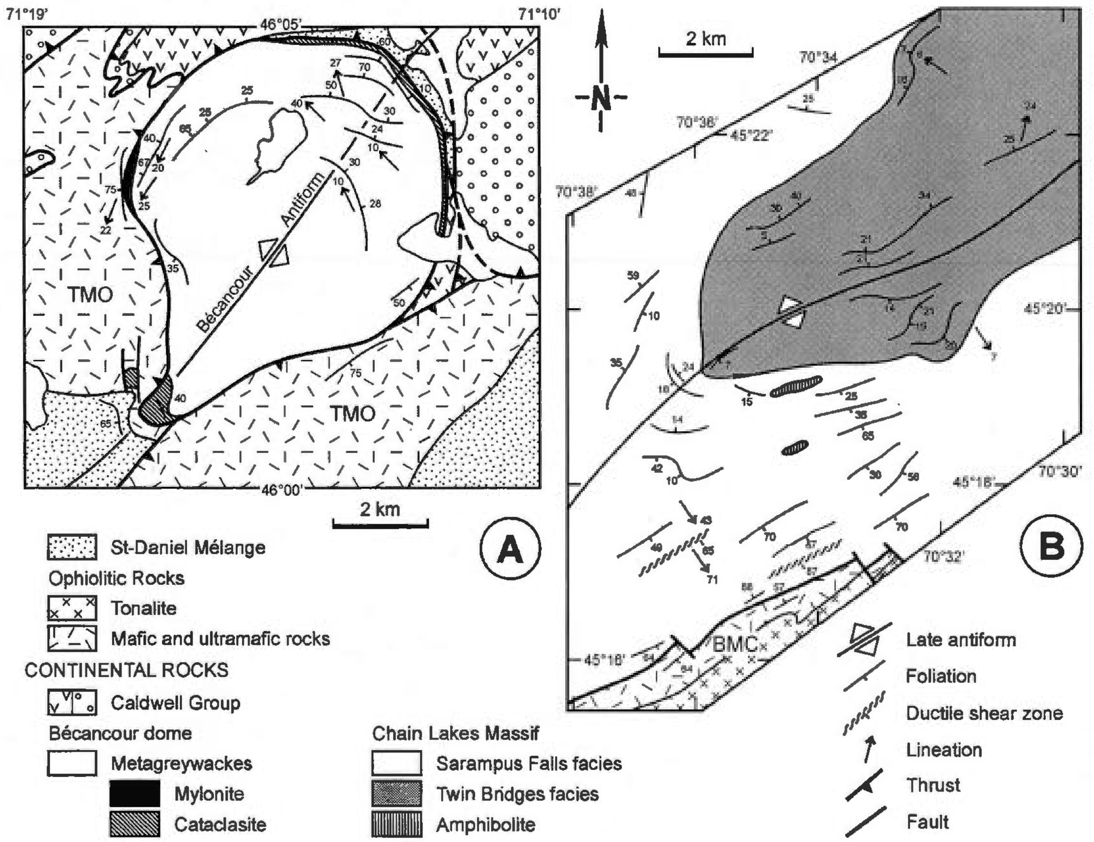

Figure 2.8 : a) Detailed geological map of the Bécancour dome; and b) the Chain Lakes massif. Both maps show foliation trends and stretching lineations. Modified from Tremblay and Pinet (1994) and Gerbi (2005), respectively. See text for details and figure 2.1 for location; common legend for a) and b). BMC – Boil Mountain Complex; TMO – Thetford-Mines ophiolite.

2.4 Discussion – The RPUC and Chain Lakes massif revisited

2.4.1 The RPUC as a deeply-eroded ophiolitic basement

Cousineau (1991) previously interpreted the RPUC as a diapiric mélange into which were incorporated "exotic" blocks detached from the Chain Lakes massif, prior to its tectonic emplacement into the Saint-Daniel Mélange. Our mapping of the RPUC indicates however that scaly, mélange-type serpentinites are restricted to backthrusting shear zones along its northwestern contact with the Caldwell Group, and that the xenolith-bearing Type 2 granitoids occur as lenses, dykes or small plutons that intruded the peridotites prior to any deformational event that affected these rocks and cannot be considered as exotic blocks. Moreover, the contact with the overlying Saint-Daniel Mélange is an unconformity, clearly indicating that the RPUC was not tectonically emplaced into the Saint-Daniel. We thus believe that the term "mélange" is inappropriate to describe these rocks and propose to refer to this unit as the Rivière-des-Plante ultramafic Complex. The lack of mafic volcanic rocks, plagiogranite and other oceanic crustal rocks in the RPUC prevents a precise determination of its chronology and tectonic environment of formation. However, the common occurence of mantle peridotites, peridotite-hosted granitoids and ophicalcites, in addition to the presence of ophiolite-derived debris flows in the overlying Saint-Daniel Mélange, suggest that the RPUC likely represents a piece of the Thetford-Mines ophiolite. The abundance of metasedimentary xenoliths, restitic material and Laurentian zircon xenocrysts in Type 2 granite of the RPUC, as well as its inferred age of crystallization and peraluminous composition, are all consistent with derivation by the anatexis of Laurentian sedimentary rocks. Following intrusion into the mantle peridotites, these granites were moderately to strongly deformed and locally mylonitized as obduction proceeded.

Schroetter et al. (2003) suggested that the Thetford-Mines ophiolite formed along a slow-spreading ridge and that synoceanic normal faults contributed to the exhumation of its lower crustal rocks. Similar normal faults were also invoked for the exhumation of upper mantle rocks and ophicalcite formation in the Lac-Brompton ophiolite (De Souza et al., 2008). Modern ophicalcite occurences are documented in oceanic core complexes where they form by circulation of low-T (< 40-90 °C) Ca-enriched fluids in ultramafic rocks causing the precipitation of carbonates as a result of mixing with seawater (Ludwig et al., 2006). Such ophicalcites are common in Alpine ophiolites and ultramafic massifs, and they are commonly overlain by basalt and/or pelagic sedimentary rocks, clearly indicating the early exhumation of oceanic mantle and/or lower crustal rocks to the seafloor (Bernoulli, 1985; Lagabrielle and Lemoine, 1997). Modern oceanic core complexes commonly form at, or close to ridge-transform intersections (Tucholke, Lin and Kleinrock, 1998; Blackman et al., 2002; Baines et al., 2003, among others), suggesting that the RPUC may have formed close to a fracture zone, as proposed by Cousineau (1991). If this is the case, the mantle peridotites were then uplifted in the footwall of normal faults (Schroetter et al., 2006; Tremblay, Meshi and Bédard, 2009) rather than by diapiric processes. Synobduction erosion and uplifting also contributed to the thinning of the ophiolite nappe and the exhumation of ultramafic rocks, as indicated by the occurence of ophiolitic and metasedimentary detritus in the Saint-Daniel Mélange.

2.4.2 Implications for the Chain lakes massif

In the light of our new data and re-interpretation of the RPUC, the inferred correlation with the Chain Lakes massif and adjacent units has to be discussed. We have shown that the fragmental granofelsic rocks of the RPUC are xenolith-bearing granitoids similar to those found in the Thetford-Mines ophiolitic mantle sequence, rather than coarse-grained rift-related metasedimentary rocks. Such peridotite-hosted granitoids are dated at ca. 470 Ma in the Thetford-Mines ophiolite, suggesting that anatexis of Laurentian sedimentary rocks was coeval with peak metamorphism and migmatization in the Chain Lakes massif at ca. 469 Ma. We think that granitic melts extracted from the Chain Lakes massif probably contributed to the formation of peridotite-hosted granitoids of the RPUC and Thetford-Mines ophiolite. Gerbi, Johnson and Aleinikoff (2006) argued that the stratigraphical disaggregation of the protolith during anatexis is the most probable mechanism to explain the widespread occurrence (900 km^2) of the fragmental migmatitic rocks of the Chain Lakes massif and their homogenous clast population. Such anatexis-related fragmentation and migmatite formation are also consistent with the occurrence of restitic material (surmicaceous enclaves) and unmelted wall-rock (xenoliths of amphibolite and metasedimentary rocks) in the RPUC Type 2 granitoids. However, a direct correlation of the RPUC xenolith-bearing granitoids with migmatitic metasedimentary rocks of the Chain Lakes massif still remains ambiguous. Although both rock units may have had a similar source, they clearly underwent a different petrogenetic evolution; the Chain Lake massif rocks representing more or less *in situ* migmatized Laurentian margin sedimentary rocks, whereas the RPUC granitoids were intruded as granitic melts into mantle peridotites during obduction, and transported with the ophiolite nappe to their current location (fig. 2.9). Metasedimentary rocks similar to the Sarampus Falls facies of the Chain Lakes massif (Gerbi, personnal communication, 2008), but of lower metamorphic grade, do occur in the Bécancour dome of southern Quebec. As for the Chain Lakes massif, the protolith of the Bécancour dome was derived from the erosion of Laurentia and was mainly deformed and metamorphosed during the taconic orogeny (Tremblay and Castonguay, 2002). If the correlation between the Chain Lakes massif and the Bécancour dome is right, it implies that both underwent a similar geodynamic evolution, therefore suggesting that deformation, metamorphism and partial melting of metasedimentary rocks in the Chain Lakes massif are likely related to ophiolite obduction (fig. 2.9). In such a model, the Chain Lakes massif would represent the most likely melt source for peridotite-hosted granitoids of the RPUC and Thetford-Mines ophiolite. Considering the available geochronological constraints for the southern Quebec Appalachians (fig. 2.10), and the inferred thickness of the Thetford-Mines ophiolite (ca. '7-8 km; Laurent, 1975, 1977), this interpretation is consistent with peak metamorphic conditions (750- 800°C – 4-5 GPa) and age estimates for high-grade metamorphism recorded by the Chain Lakes massif. The inferred sequence of events that affected the Laurentian margin along the Quebec-Maine transect is synthesized on figure 2.10 and can be summarized as follows: 1) formation and cooling of the metamorphic sole underlying the Thetford-Mines ophiolite (ca. 480-475 Ma), shortly after, or perhaps during the final increments of suprasubduction zone seafloor spreading; 2) greenschist to amphibolite-facies metamorphism of the Laurentian margin during obduction of the ophiolites, (ca. 480-470 Ma); 3) local anatexis of the Laurentian margin as preserved in the Chain Lakes massif, and formation of the peridotite-hosted granitoids (ca. 470 Ma); and 4) greenschist-facies metamorphism in the internal Humber zone of southern Quebec, during final emplacement of the ophiolites and progressive thickening of the Laurentian margin (ca. 470-460 Ma; Castonguay et al., 2001; Tremblay and Castonguay, 2002; Castonguay, Ruffet and Tremblay, 2007).

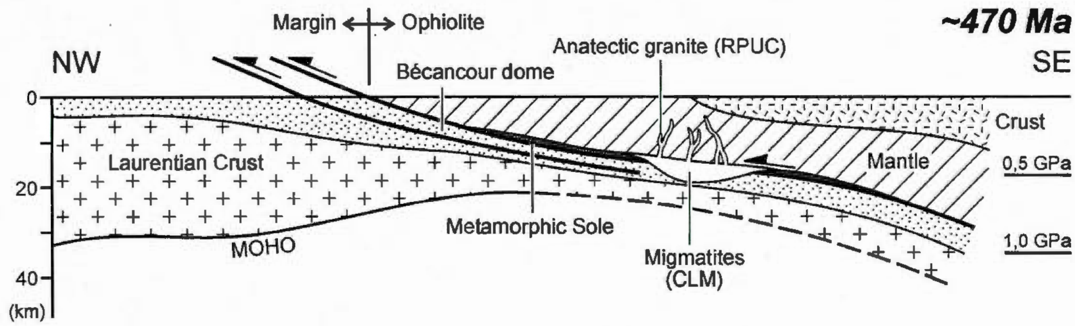

Figure 2.9 : Tectonic model for the formation of peridotite-hosted granitoids of the RPUC and the anatexis of the Chain Lakes massif during the taconic orogeny. Modified from Searle and Cox (1999). See text for discussion. CLM – Chain Lakes massif.

2.5 Conclusion

Our observations on the RPUC provide geological evidence for its origin as a deeply-eroded ophiolite remnant correlative with the mantle sequence of the composite Thetford-Mines, Asbestos and Lac-Brompton ophiolites, and unconformably overlain by chaotic to well-stratified sedimentary rocks belonging to the Saint-Daniel Mélange. In addition to typical ophiolitic serpentinized peridotites, ophicalcite and gabbro, the RPUC comprises anatectic granites which contain inherited Laurentian material and show a fragmental texture marked by abundant xenoliths of metamorphic rocks dispersed in a granitic matrix. These granites are interpreted as xenolith-bearing facies of peridotite-hosted granitoids found in the Thetford-Mines ophiolite, and which can be attributed to the anatexis of the Laurentian margin during ophiolite obduction (Whitehead, Dunning and Spray, 2000). The correlation of these fragmental igneous rocks with high-grade metasedimentary rocks of the Chain Lakes massif, as previously suggested, is not verified. In fact, the Chain Lakes massif better compares (and likely correlates) with infra-ophiolitic Laurentian metasedimentary rocks of southern Quebec preserved, for instance, in the Bécancour dome of the Thetford-Mines area. This suggests that the Chain Lakes massif, as the Bécancour dome, likely represents a structural inlier of continental metasedimentary rocks and that its protolith was deposited on, or close to Laurentian continental crust. We suggest that high-grade metamorphism and anatexis of the massif was caused by the obduction of a large ophiolitic nappe that is currently represented by the Thetford-Mines – Asbestos – Lac-Brompton ophiolites in southern Quebec, and part of which is possibly preserved as the Boil Mountain Complex and Jim Pond Formation in western Maine. Migmatized metasedimentary rocks of the Chain Lakes massif appear as a likely melt source for the peridotite-hosted granitoids occurring in the RPUC and Thetford-Mines ophiolite. The leading edge of Laurentia and the root zone of the Southern Quebec ophiolite Belt are consequently probably located east of the Chain Lakes massif, as originally suggested by Zen (1983) and Pinet and Tremblay (1995b).

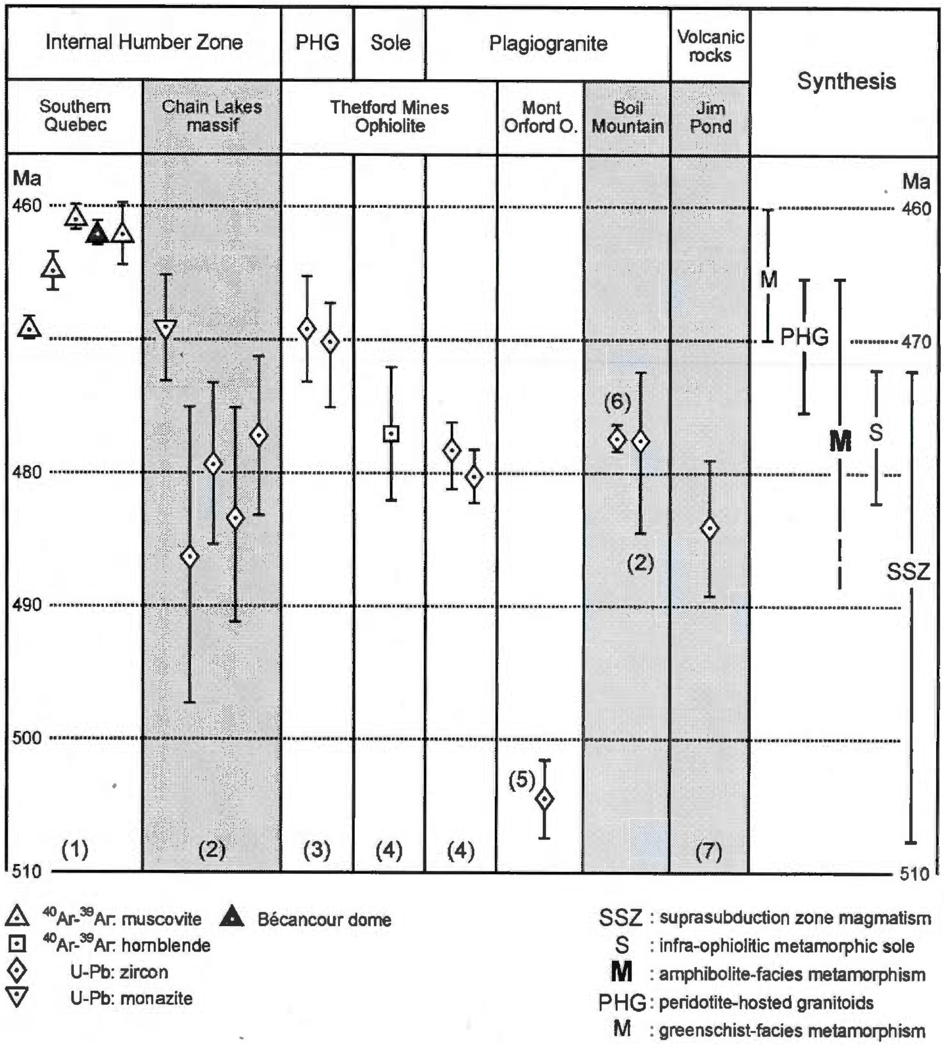

Figure 2.10 : U-Pb and ^{40}Ar-^{39}Ar age data for the internal Humber zone, the Thetford-Mines and Mont-Orford ophiolites, the Chain Lakes massif, the Boil Mountain Complex and the Jim Pond Formation. Data sources: (1) – Castonguay et al. (2001); (2) – Gerbi et al. (2006) and Gerbi, Johnson and Aleinikoff (2006); (3) – Whitehead, Reynolds and Spray (1995); (4) – Whitehead, Dunning and Spray (2000); (5) – David and Marquis (1994); (6) – Kusky, Chow and Bowring (1997); (7) – Moench, Aleinikoff and Boudette (2000).

2.6 Acknowledgements

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CHAPITRE III

OPHIOLITE OBDUCTION IN THE QUEBEC APPALACHIANS, CANADA – 40Ar/39Ar AGE CONSTRAINTS AND EVIDENCE FOR SYN-TECTONIC EROSION AND SEDIMENTATION

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ABSTRACT

Detailed field work conducted in the Dunnage zone of the Quebec Appalachians, is here combined with $^{40}Ar/^{39}Ar$ dating on a series of ophiolitic massifs, crosscutting granites and associated metamorphic rocks occurring along the Baie Verte-Brompton line, the Taconian suture between Laurentia and Lower Paleozoic peri-Laurentian oceanic terranes. Studied massifs are the Lac-Brompton ophiolite and the Rivière-des-Plante ultramafic Complex in southern Quebec, and the Nadeau Ophiolitic Mélange in the Gaspé Peninsula. Our work suggests that these massifs form remnants of eroded ophiolitic nappes, which are unconformably overlain by the Saint-Daniel and Rivière-Port-Daniel mélanges, and correlate with the Thetford-Mines and Mont-Albert ophiolitic complexes. Our $^{40}Ar/^{39}Ar$ data and compiled regional age constraints indicate that ophiolite obduction was diachronous along the strike of the orogen. The timing of obduction and mélange formation varies accordingly to the irregular geometry of the Early Paleozoic Laurentian margin, with earlier collision occurring along, or at the margins of promontories. Obduction was initiated with the formation of infraophiolitic metamorphic soles between ca. 479 and 472 Ma in southern Quebec and the Nadeau Ophiolitic Mélange, and possibly as late as ca. 470-466 Ma for the Mont-Albert Complex. These sole rocks were later exhumed and translated onto the Laurentian margin with the overlying ophiolites between 475 and 460-457 Ma. The uplifting and erosion of the orogenic wedge during the waning stages of obduction, has resulted in the sedimentation of olistostromal mélanges and onlapping flysch units above the ophiolitic nappes, as well as foredeep flysch successions during the latest Arenig(?) to earliest Caradoc.

RÉSUMÉ

Des études détaillées effectuées dans la zone de Dunnage des Appalaches du Québec sont combinées à de la géochronologie $^{40}Ar/^{39}Ar$ sur des roches granitiques et métamorphiques associées à des massifs ophiolitiques situés le long de la ligne Baie Verte-Brompton, la suture taconienne entre Laurentia et les terrains océaniques péri-laurentiens accrétés du Paléozoïque inférieur. Les massifs étudiés sont l'ophiolite du Lac-Brompton et le Complexe ultramafique de la Rivière-des-Plante dans le sud du Québec, ainsi que le Mélange ophiolitique de Nadeau dans la péninsule gaspésienne. Les données acquises suggèrent que ces massifs représentent des fragments de nappes ophiolitiques corrélables sur lesquelles reposent en discordance les

mélanges de Saint-Daniel et de Rivière-Port-Daniel. Nos données $^{40}Ar/^{39}Ar$ et la compilation de données géochronologiques existantes, indiquent que l'obduction des ophiolites a été diachrone le long de l'orogène. La chronologie de l'obduction varie selon la géométrie de la marge laurentienne du Paléozoïque inférieur, avec une collision précoce au niveau des promontoires. L'obduction s'est initiée avec la formation des semelles métamorphiques infraophiolitiques entre ca. 479 et 472 Ma dans le sud du Québec et le Mélange ophiolitique de Nadeau, et probablement entre ca. 470-466 Ma pour le Complexe du Mont-Albert. Celles-ci ont ensuite été exhumées et transportées sur la marge laurentienne avec les roches ophiolitiques sus-jacentes entre 475 et 460-457 Ma. La surrection et l'érosion du prisme orogénique pendant les stades finaux d'obduction, ont engendré la sédimentation de mélanges olistostromaux et de flyschs sur les nappes ophiolitiques, ainsi que de successions flyschiques dans l'avant-pays, de l'Arenig tardif(?) jusqu'au Caradoc.

3.1 Introduction

Ophiolitic complexes and mélanges occur discontinuously along the Baie Verte-Brompton line (Williams and St-Julien, 1982), a narrow linear belt extending for 1500 km in the Canadian Appalachians, from southern Quebec to Newfoundland (fig. 3.1a). Williams and St-Julien (1982) originally interpreted the Baie Verte-Brompton line as a steeply-dipping polyphase tectonic zone of ophiolite emplacement representing the surface expression of the Taconian suture zone between Laurentia and the bordering Iapetus Ocean. In Quebec, the Taconian orogeny resulted from an Early to Middle Ordovician, Tethyan-type obduction event involving the emplacement of large nappes of suprasubduction zone oceanic lithosphere onto Laurentia (Pinet and Tremblay, 1995b; Tremblay and Castonguay, 2002; Schroetter, Tremblay and Bédard, 2005; Malo et al., 2008; Tremblay, Meshi and Bédard, 2009; Tremblay, Ruffet and Bédard, 2011). The term "obduction", which generally refers to any type of ophiolite emplacement mechanism (e.g., Coleman, 1971; Dewey, 1976; Moores, 1982; Searle and Stevens, 1984; Searle et al., 1994; Gregory, Gray and Miller, 1998; Gray and Gregory, 2003; Whattam, 2009), is here used to describe the combination of processes by which an ophiolite is emplaced onto a continental margin, from the inception of subduction to its subaerial exposure as part of an orogenic belt (Gray, Gregory and Miller, 2000; Wakabayashi and Dilek, 2003; Tremblay, Ruffet and Bédard, 2011).

At the orogen scale, Cambrian-Ordovician lithotectonic features of the Appalachians define embayments and promontories that are believed to be inherited from the cratonic break-up of Rodinia (fig. 3.1a; Thomas, 1977; Allen, Thomas and Lavoie, 2010), and to have played an important role in the sedimentological and tectonic evolution of the Appalachians (Stockmal et al., 1987; van Staal et al., 1998; Malo et al., 1995; Tremblay, Ruffet and Castonguay, 2000). More specifically, the promontory-embayment geometry of the margin has been inferred to influence the distribution of the rift and passive margin sequences (Lavoie et al., 2003; Cousineau and Longuépée, 2003), the formation, obduction and preservation of the ophiolites (Cawood and Suhr, 1992), the along-strike variations in the timing and structural style of both the Taconian and Acadian orogenies (Lavoie, 1994; Tremblay, Ruffet and Castonguay, 2000; Sacks et al., 2004; Malo et al., 1992, 1995, 2008) and the Lower Paleozoic syn-tectonic sedimentation (Hiscott, 1995).

Geochronological studies, mostly from the Thetford-Mines and Mont-Albert ophiolitic complexes (figs. 3.1b, 3.2; Lux, 1986; Dunning and Pedersen, 1988; Whitehead, Reynolds and Spray, 1995; Whitehead, Dunning and Spray, 2000; Pincivy et al., 2003; Malo et al., 2008; Tremblay, Ruffet and Bédard, 2011), have provided significant isotopic age constraints for ophiolite formation and/or obduction. However, little is known about the tectonic history and timing of obduction and mélange formation along the entire length of the Baie Verte-Brompton line in the Quebec Appalachians. In this contribution, we combine detailed field observations and new $^{40}Ar/^{39}Ar$ age data from granites, ophiolitic rocks and infraophiolitic metamorphic rocks of the Lac-Brompton ophiolite and Rivière-des-Plante ultramafic Complex in southern Quebec, and the Nadeau Ophiolitic Mélange in the Gaspé Peninsula (figs. 3.1b, 3.2). These data are then synthesized with published geochronological constraints for ophiolitic complexes of the Quebec Appalachians in order to document orogenic processes related to ophiolite obduction, syn-tectonic sedimentation and mélange formation in the context of an irregular collision zone.

3.1.1 Geological setting

The Quebec Appalachians comprise three main lithotectonic assemblages: the Cambrian-Ordovician Humber and Dunnage zones (Williams, 1979), and the overlying Silurian-Devonian succession of the Gaspé Belt (fig. 3.1; Bourque, Malo and Kirkwood, 2000; Lavoie and Asselin, 2004).

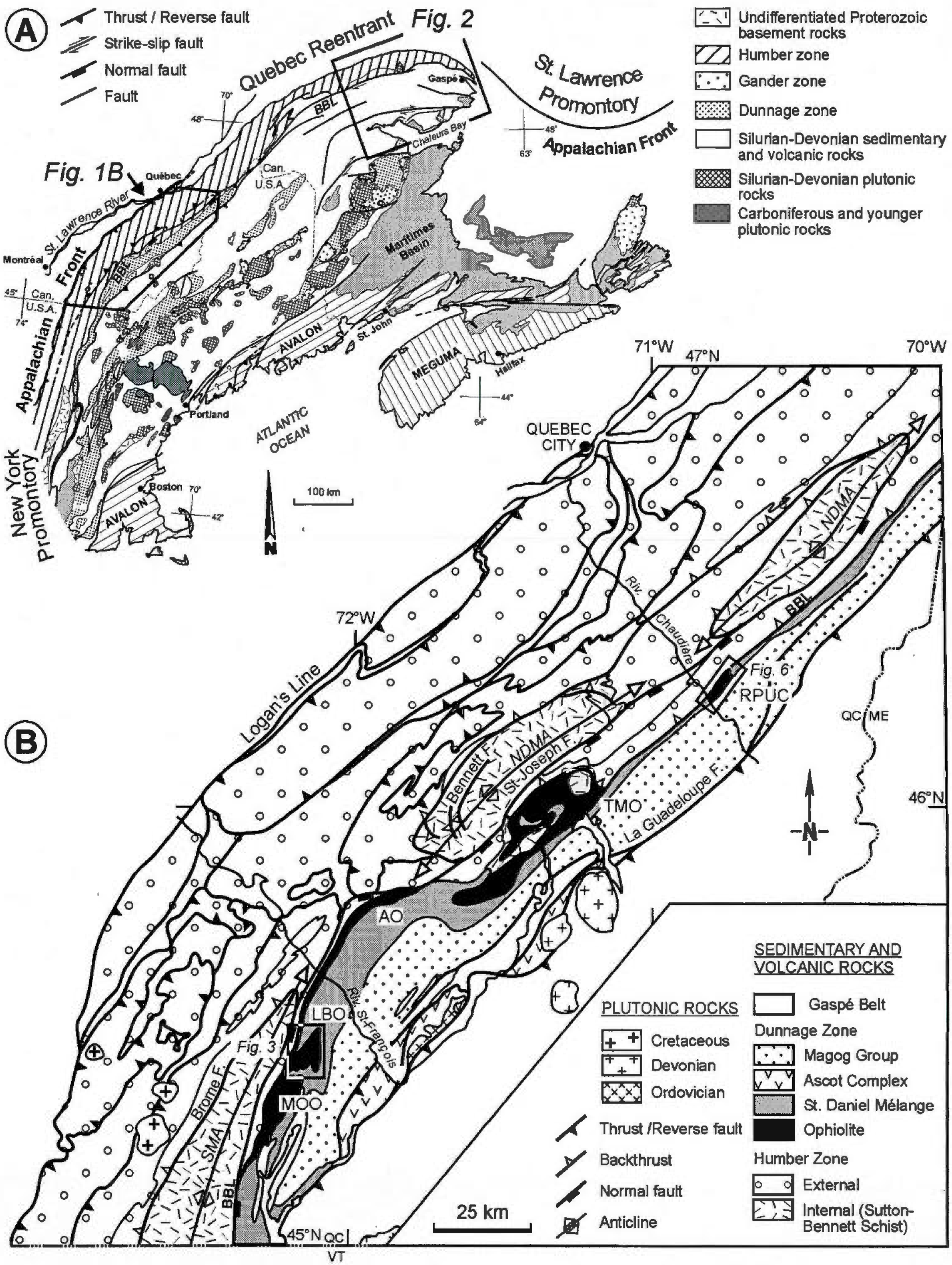

Figure 3.1 : a) Lithotectonic map of the Northern Appalachians of mainland Canada and New England showing embayment-promontory geometry of the region (modified from Tremblay and Pinet, 2005). BBL – Baie Verte-Brompton line. b) Generalized geological map of the southern Quebec Appalachians (modified from Schroetter, Tremblay and Bédard, 2005). SMA – Sutton Mountains anticlinorium; NDMA – Notre-Dame Mountains anticlinorium; MOO – Mont-Orford ophiolite; LBO – Lac-Brompton ophiolite; AO – Asbestos ophiolite; TMO – Thetford-Mines ophiolite; RPUC – Rivière-des-Plante ultramafic Complex. See fig. 3.1a for location.

The Humber and Dunnage zones were amalgamated during the Taconian orogeny and represent the remnants of the Laurentian continental margin and the adjacent oceanic domain, respectively (Williams, 1979; van Staal et al., 1998). In Quebec, penetrative Taconian deformation is essentially confined to the Humber zone, and has been attributed to the closure of the Iapetus ocean and obduction of a large ophiolitic nappe, now preserved as the southern Quebec and Gaspé Peninsula ophiolites (Pinet and Tremblay, 1995b; Tremblay and Castonguay, 2002; Malo et al., 2008).

3.1.1.1 Stratigraphy and structure of the Dunnage zone in the southern Quebec Appalachians

In the Quebec Appalachians, the Dunnage zone is best exposed in southern Quebec (fig. 3.1b), where it consists of : 1) a series of well-preserved to dismembered ophiolite complexes, namely, from south to north, the Mont-Orford, Lac-Brompton, Asbestos and Thetford-Mines ophiolites, and the Rivière-des-Plante ultramafic Complex; 2) the Saint-Daniel Mélange; 3) the Magog Group; and 4) the Ascot Complex (fig. 3.2; Tremblay, 1992a). The ophiolite complexes of southern Quebec are all believed to represent remnants of a composite slab of suprasubduction zone oceanic lithosphere (Schroetter, Tremblay and Bédard, 2005; De Souza et al., 2008; De Souza and Tremblay, 2010a). Oceanic plagiogranites from the Thetford-Mines ophiolite yielded U-Pb zircon ages of 478 $^{+3}/_{-2}$ and 480 ± 2 Ma (Whitehead, Dunning and Spray, 2000). The Mont-Orford ophiolite is presumably older (U-Pb zircon age of 504 ± 3 Ma; David and Marquis, 1994) and is believed to represent a magmatic arc basement that rifted to form the other complexes (De Souza et al., 2008). Mantle peridotites of the Thetford-Mines, Asbestos, Lac-Brompton and Rivière-des-Plante complexes, are crosscut by a series of granitoids that are referred to as peridotite-hosted granites (Whitehead, Dunning and Spray, 2000; De Souza et al., 2008; De Souza and Tremblay, 2010a; Tremblay, Ruffet and Bédard, 2011). These granitic rocks were shown to be derived from the anatectic partial melting of Laurentian continental margin sedimentary rocks during ophiolite emplacement as a result of shear heating and residual heat transfer from the ophiolite to the margin (Whitehead, Dunning and Spray, 2000; Tremblay, Ruffet and Bédard, 2011). In the Thetford-Mines area, these granitic rocks yielded U-Pb zircon ages of 470 $^{+5}/_{-3}$ Ma and 469 ± 4 Ma (Whitehead, Dunning and Spray, 2000), and $^{40}Ar/^{39}Ar$ muscovite ages varying between ca. 466 Ma and 460 Ma (Tremblay, Ruffet and Bédard, 2011). A $^{40}Ar/^{39}Ar$ hornblende isochron age of 477 ± 5 Ma (Whitehead, Reynolds and Spray, 1995) for the infraophiolitic metamorphic sole of the Thetford-Mines ophiolite, was recently shown to be partly inherited from its basaltic protolith, and revised to a younger age of ca. 471 Ma (Tremblay, Ruffet and Bédard, 2011).

Debris flows and conglomerates characterizing the base of the Saint-Daniel Mélange (fig. 3.1b) form the lower part of a syn-collisional sedimentary basin unconformably overlying the ophiolitic basement and underlying metamorphic rocks (Schroetter et al., 2006; De Souza et al., 2008; Tremblay, Meshi and Bédard, 2009). The Magog Group is a Caradoc (the Ordovician time scale of Sadler, Cooper and Melchin (2009) is used throughout this article) flysch-dominated turbiditic succession that unconformably overlies the Saint-Daniel Mélange (Cousineau and St-Julien, 1994; Schroetter et al., 2006), whereas the Ascot Complex is thought to represent the remnants of a Middle to Late Ordovician peri-Laurentian volcanic arc sequence (fig. 3.1b; Tremblay, Hébert and Bergeron, 1989; Tremblay, 1992a).

In southern Quebec, the Humber and Dunnage zones share a similar structural evolution (Schroetter, Tremblay and Bédard, 2005). A Middle to Late Ordovician (471-456 Ma; Castonguay et al., 2001; Tremblay and Castonguay, 2002; Castonguay, Ruffet and Tremblay, 2007; Tremblay, Ruffet and Bédard, 2011) S_{1-2} schistosity is associated with ophiolite emplacement during the Taconian orogeny, and is only developed in rocks of the Humber zone and the infraophiolitic metamorphic sole rocks (Tremblay and Castonguay, 2002; Schroetter, Tremblay and Bédard, 2005; Daoust, 2007). Taconian deformation culminated with the emplacement of the taconic allochthons between 460 and 445 Ma as part of a foreland-propagating thrust system (St-Julien and Hubert, 1975; Sasseville et al., 2008). Two generations of post-obduction structures are recognized: 1) D_3 Silurian – Early Devonian SE-verging folds and faults that culminated with the formation of steep southeast-dipping normal faults, such as the St-Joseph fault, and 2) D_4 Late Devonian NW-verging folds and reverse faults related to the Acadian orogeny (fig. 3.1b; Tremblay and Pinet, 1994; Pinet, Castonguay and Tremblay, 1996; Castonguay et al., 2001; Tremblay and Castonguay, 2002; Castonguay and Tremblay, 2003; Schroetter, Tremblay and Bédard, 2005; Castonguay, Ruffet and Tremblay, 2007).

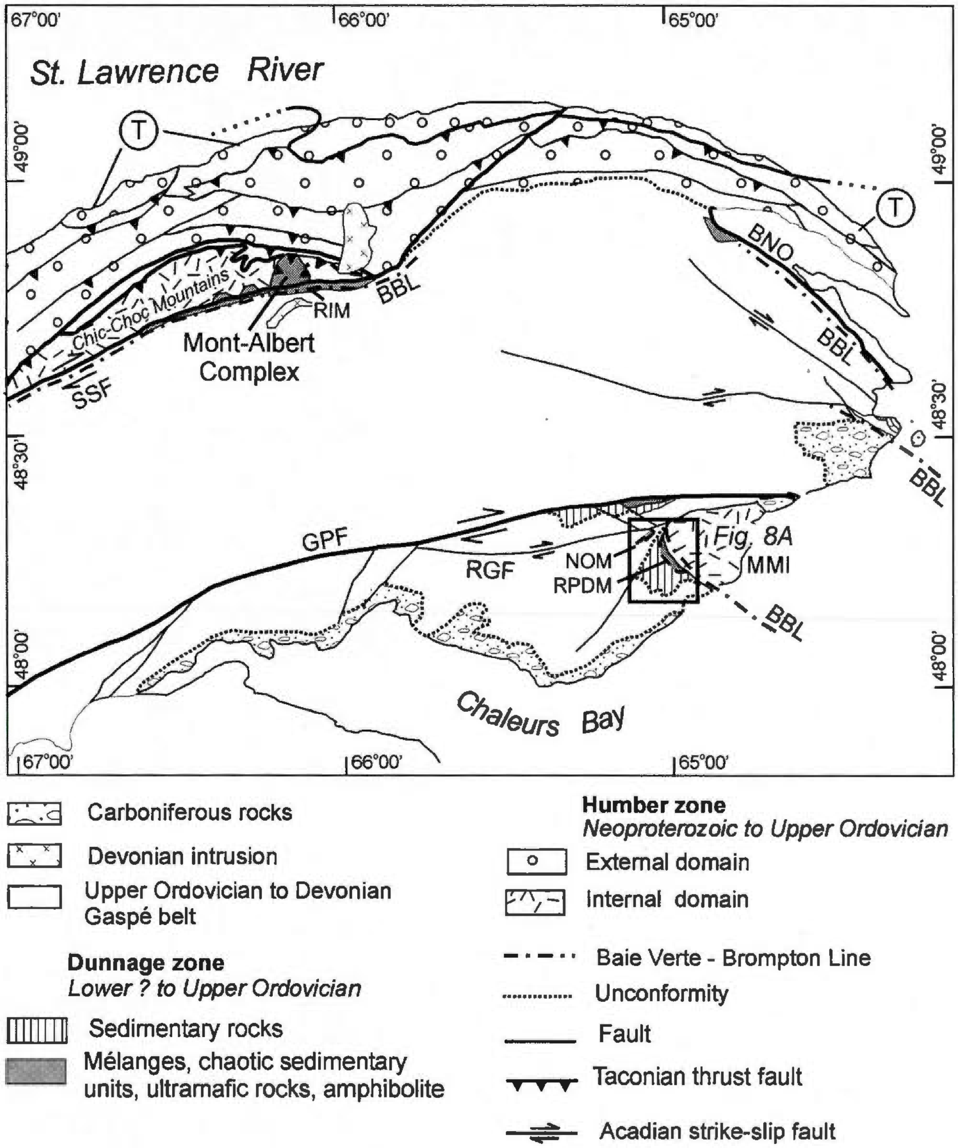

Figure 3.2 : Generalized geological map of the Gaspé Peninsula modified from Malo et al. (2008). BBL – Baie Verte-Brompton line; BNOF – Bras Nord-Ouest fault; MMI – Maquereau-Mictaw inlier; GPF – Grand Pabos fault; NOM – Nadeau Ophiolitic Mélange; RGF – Rivière Garin fault; RPDM – Rivière-Port-Daniel Mélange; SSF – Shickshock Sud fault; T – Tectonic slices containing Tourelle, Deslandes and Cloridome formations or other correlative units. The geology of New-Brunswick is undifferentiated. See fig. 3.1a for location.

3.1.1.2 Stratigraphy and structure of the Dunnage zone in the Gaspé Peninsula

In the Gaspé Peninsula (fig. 3.2), Cambrian(?)-Ordovician ophiolitic rocks, mélanges and sedimentary rocks assigned to the Dunnage zone occur in the Mont-Albert Complex, as a series of tectonic slivers along major faults and as structural inliers beneath the sedimentary cover of the Gaspé Belt (see Tremblay, Malo and St-Julien, 1995 for a review). The Mont-Albert Complex (Beaudin, 1980) consists of mantle peridotites and of an infraophiolitic metamorphic sole (Gagnon and Jamieson, 1986; Pincivy et al., 2003; Malo et al., 2008). As in southern Quebec, olistostromal breccias comprising ophiolitic rock fragments occur in the Gaspé Peninsula as part of the Rivière-Port-Daniel Mélange, which is associated with ultramafic rock slivers and delineates the Baie Verte – Brompton line in the Maquereau-Mictaw inlier (fig. 3.2; De Brouker, 1987). Williams and St-Julien (1982) have suggested, based on fragment and matrix petrography and stratigraphic relationships with adjacent pre-Silurian rock units, that the Rivière-Port-Daniel Mélange correlates with the Saint-Daniel Mélange, and that both were probably formed in similar stratigraphic settings. The Mictaw Group is a Llanvirn to Caradoc flysch-dominated turbiditic succession that unconformably overlies the Rivière-Port-Daniel Mélange (De Brouker, 1987).

Taconian accretionary events in the Gaspé Peninsula include the formation of high-grade metamorphic rocks, such as the metamorphic sole of the Mont-Albert Complex (Pincivy et al., 2003; Malo et al., 2008), and the emplacement of northwest-directed thrust sheets during ophiolite obduction (St-Julien and Hubert, 1975; Malo et al., 2008). The metamorphic sole of the Mont-Albert Complex and the underlying metamorphic rocks of the Humber zone yielded $^{40}Ar/^{39}Ar$ amphibole and muscovite ages of ca. 465 Ma to 457 Ma and 459 to 456 Ma, respectively, with evidence of a late increment at ca. 449 Ma in both of these assemblages (Lux, 1986; Pincivy et al., 2003; Malo et al., 2008). These ages were interpreted to record the progressive transfer of deformation from the oceanic domain to the continental margin (Malo et al., 2008). In the Taconian foreland of the Gaspé Peninsula, synorogenic chromite-bearing flysch units provide evidence for ophiolite obduction and erosion, and the foundering of the Laurentian margin as early as the late Arenig-early Llanvirn (fig. 3.2; Hiscott, 1978, 1995). Acadian regional structures are related to a transpressional deformation regime and the development of dextral strike-slip faults and oblique folds (Malo and Béland, 1989; Malo et al., 1992; Pinet et al., 2008, 2010). Palinspastic restoration of the pre-Acadian Cambrian-Ordovician features in the Gaspé Peninsula suggests that rocks of the Nadeau Ophiolitic Mélange and the Maquereau-Mictaw inlier were originally located ca. 120 km to the east of their current location, along the Baie Verte-Brompton line (Malo et al., 1992).

3.2 $^{40}Ar/^{39}Ar$ geochronological data

For the purpose of this study, a series of samples of igneous and metamorphic rocks collected in the Lac-Brompton ophiolite, the Rivière-des-Plante ultramafic Complex and the Nadeau Ophiolitic Mélange, was selected for white mica and amphibole laser step-heating $^{40}Ar/^{39}Ar$ analyses. Single grains of sericite, muscovite and amphibole were separated from 0.25 to 0.5 mm fractions of crushed rock samples and handpicked using a binocular microscope. The sample preparation and analytical procedures for the laser step-heating measurements are detailed by Ruffet, Féraud and Amouric (1991), Ruffet et al. (1995), Castonguay et al. (2001) and Castonguay, Ruffet and Tremblay (2007), whereas the tectonic setting of the dated samples and analytic results are presented in the following sections.

3.2.1 The Lac-Brompton ophiolite

The Lac-Brompton ophiolite (De Souza et al., 2008) mostly consists of harzburgitic mantle peridotites with minor occurrences of a discontinuous crustal unit preserved as pyroxenitic to gabbroic plutonic rocks and boninitic mafic volcanic rocks (fig. 3.3). It has been correlated with both the Asbestos and Thetford-Mines ophiolites based on the geochemistry of volcanic rocks and dykes, ophiolite stratigraphy, and relationships with adjacent rock units (De Souza et al., 2008). The lowermost contact of the Lac-Brompton ophiolite is marked by a discontinuous unit of metavolcanic and metasedimentary rocks that have been collectively interpreted as the remnants of an infraophiolitic metamorphic sole (fig. 3.3; Daoust, 2007). The ophiolite and its metamorphic sole are both unconformably overlain by the Saint-Daniel Mélange, which consists of a basal unit made up of polymictic breccia and laminated chert/mudstone that is overlain by pebbly mudstone, black shale and sandstone. The breccia unit mostly consists of ophiolite-derived mafic and ultramafic lithologies, granitoid and sedimentary rock fragments, but is also locally entirely made up of foliated sole-type metamorphic rock clasts, suggesting that the metamorphic sole was formed and then uplifted during or prior to the sedimentation of the Saint-Daniel Mélange.

The metamorphic sole itself is made up of amphibolites, phyllite and mica schist (Daoust, 2007). The metamorphic mineral assemblages are, in order of increasing metamorphic grade: (1) epidote + plagioclase + hornblende ± sphene ± chlorite; (2) hornblende + epidote + plagioclase ±

sphene and (3) hornblende + plagioclase ± sphene in the amphibolites, and muscovite + quartz + chlorite ± plagioclase ± garnet ± zoisite in the phyllites and mica schists. The amphibole compositions and the nature of mineral assemblages indicate maximum pressure and temperature conditions of 5-8 kbar and 700°C (Daoust, 2007). The geochemical composition of the mafic protolith to the amphibolite corresponds to tholeiitic-transitional MORB-like to alkali basalts, which have been interpreted as metamorphosed passive margin to continental rift-related volcanic rocks (Daoust, 2007). The dominant fabric in the metamorphic sole is a S_{1-2} composite foliation marked by amphibole, plagioclase, epidote and micaceous minerals, that has been attributed to ophiolite obduction (Daoust, 2007). This foliation is absent in the Saint-Daniel Mélange and is only developed in ultramafic rocks that are immediately adjacent to the metamorphic sole. The ophiolitic and metamorphic sole rocks, as well as the Saint-Daniel Mélange are, however, all affected by northwest- to northeast-trending F_3 folds and faults belonging to the D_3 phase of Tremblay and Pinet (1994), and Tremblay and Castonguay (2002). These are in turn overprinted by Acadian-related, upright and steeply-plunging F_4 folds (fig. 3.3).

$^{40}Ar/^{39}Ar$ results

Published isotopic age constraints for the Lac-Brompton ophiolite and related metamorphic rocks are rare. A single K-Ar biotite age of 358 ± 16 Ma has been measured for a biotite- and muscovite-bearing granitoid crosscutting the mantle peridotites (Wanless, 1963), which has been later interpreted as a crystallization age (St-Julien and Hubert, 1975). However, De Souza et al. (2008) have shown that the dated rock is a peridotite-hosted granite similar to those of the Thetford-Mines ophiolite, suggesting that this Late Devonian age is related to thermal resetting during the Acadian orogeny, or more simply to weathering.

For this study, the analyzed samples of metamorphic and igneous rocks were collected in five different locations in the Lac-Brompton ophiolite (see fig. 3.3). Amphibole sample 6304 was taken from a dyke of amphibolitized and coarse-grained pyroxenitic gabbro crosscutting the mantle peridotite. This gabbro is not foliated and is composed of magnesio-hornblende, zoisite, albite and relict clinopyroxene, a mineral assemblage typical of oceanic hydrothermal metamorphism commonly developed in ophiolitic crustal rocks (Juteau and Maury, 1999). The amphibole yielded a $^{40}Ar/^{39}Ar$ plateau age of 476.7 ± 6.2 Ma (fig. 3.4a). The large error is correlative of the very high CaO/K$_2$O calculated ratios (ca. 130; CaO/K$_2$O = 2.179 x ($^{37}Ar_{Ca}/^{39}Ar_K$)). There is no sign of a post-cooling disturbing event.

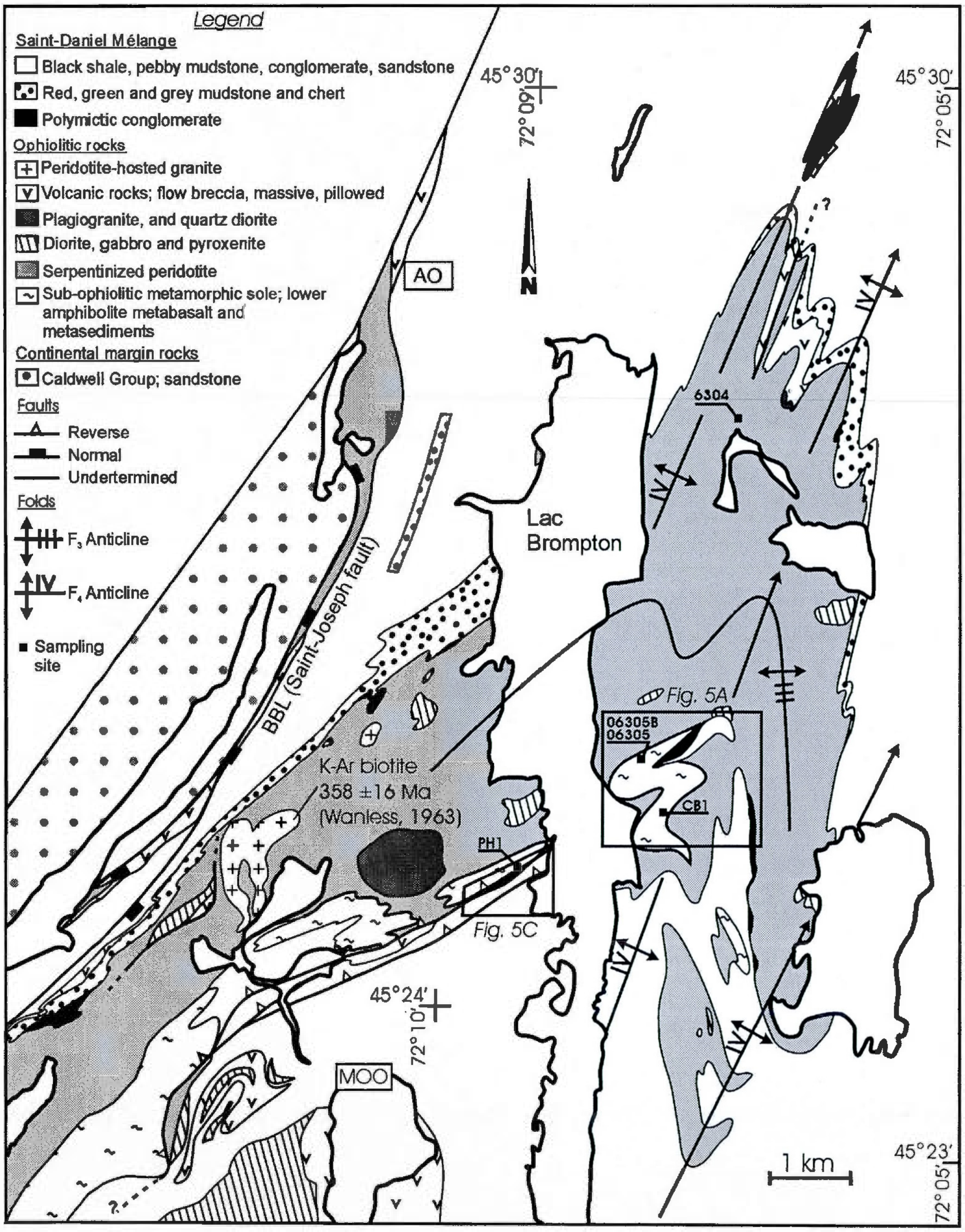

Figure 3.3 : Generalized geological map of the Lac-Brompton area showing the location of the samples used for $^{40}Ar/^{39}Ar$ analyses. MOO – Mont-Orford ophiolite; AO – Asbestos ophiolite; BBL – Baie Verte-Brompton line. Modified from De Souza et al. (2008). See fig. 3.1b for location.

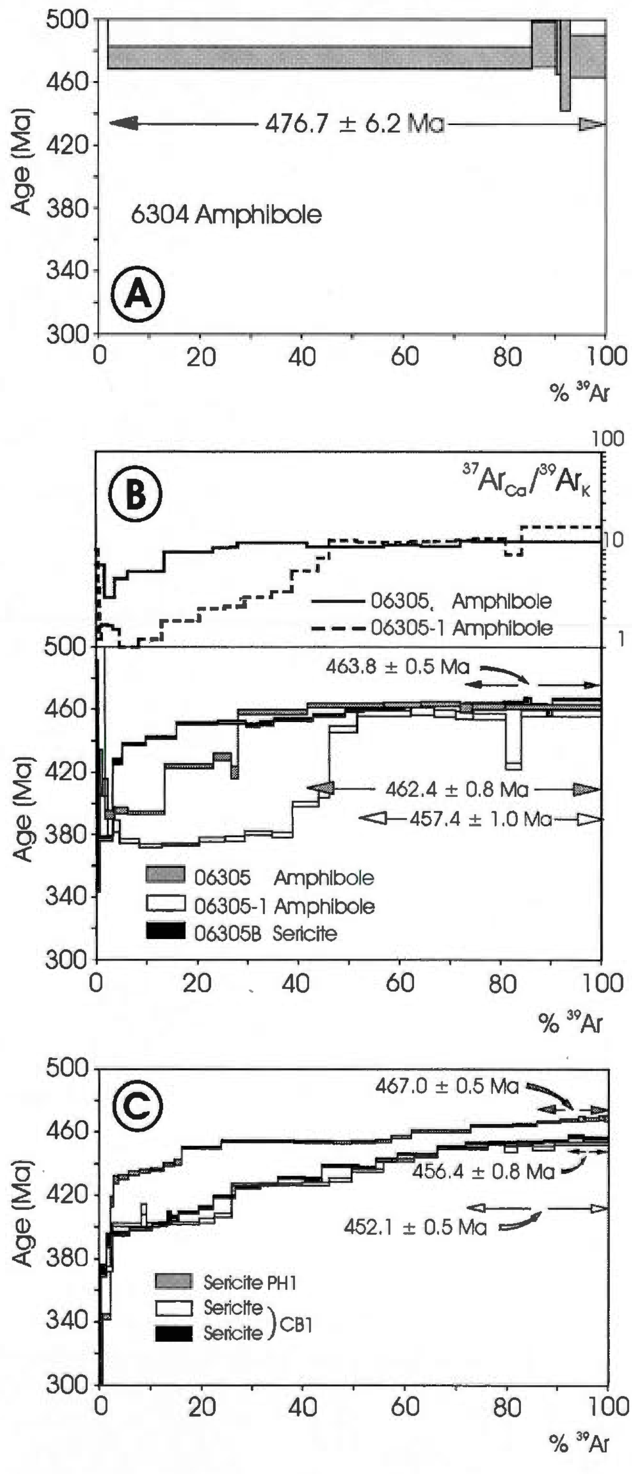

Figure 3.4 : $^{40}Ar/^{39}Ar$ age spectra for a) amphibole of sample 6304; b) amphibole and sericite from sample 06305 and 06305B, respectively; c) sericite from samples PH1 and CB1. See figs. 3.3 and 3.5 for location of sampling sites.

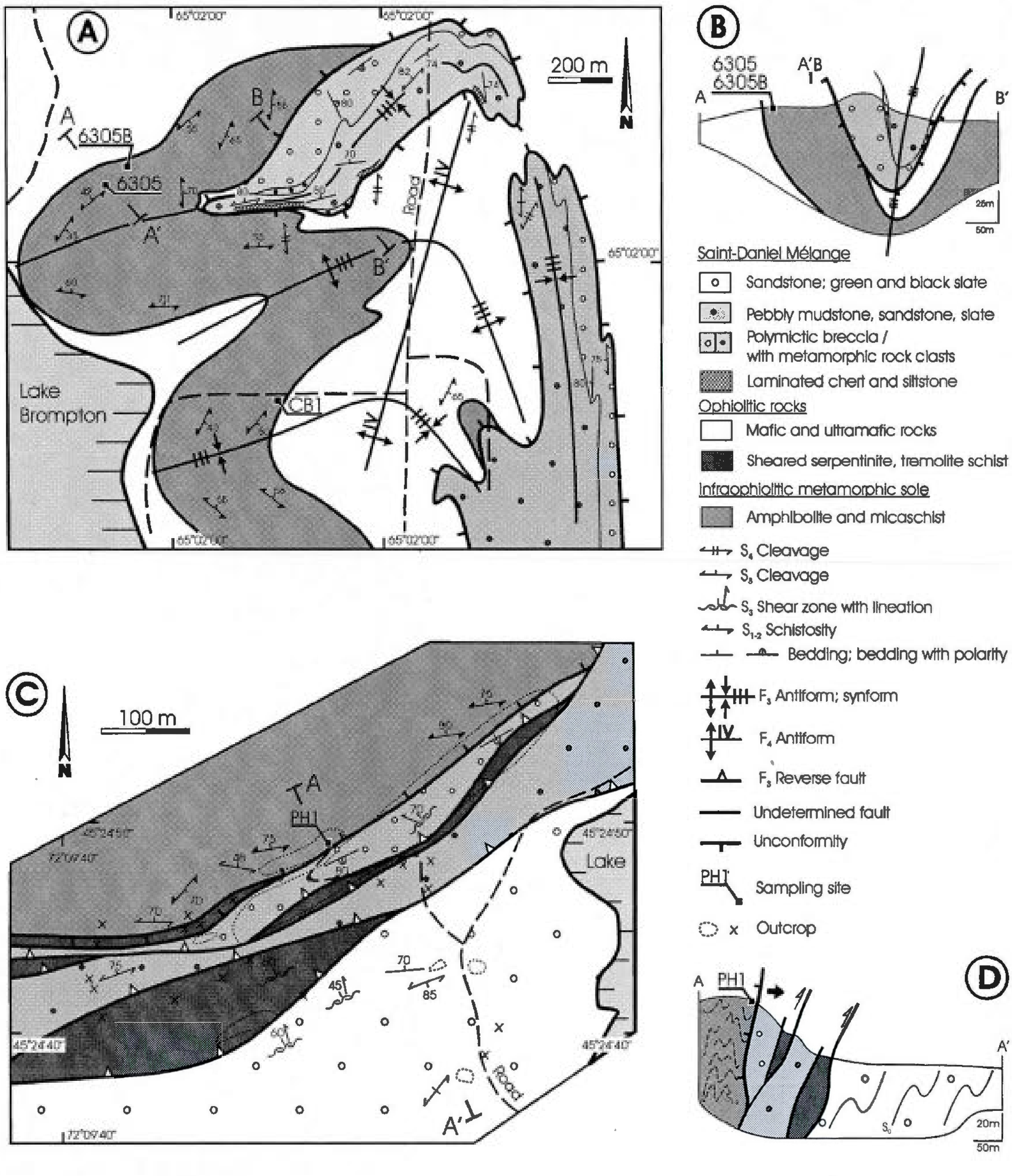

Figure 3.5 : a) Detailed geological map of the metamorphic sole of the Lac-Brompton ophiolite and Saint-Daniel Mélange (see fig. 3 for location); b) cross-section showing the location and tectono-stratigraphic setting of samples 06305, 06305B; c) Detailed geological map of the Saint-Daniel Mélange and metamorphic sole of the Lac-Brompton ophiolite (see Fig. 3.3 for location); d) cross-section showing the location and tectono-stratigraphic setting of sample PH1. Thick arrow indicates topping direction in the polymictic breccia. The same legend applies to figs. 3.5a, 3.5b, 3.5c and 3.5d.

Four samples were collected from the metamorphic sole of the ophiolite. Samples 06305 and 06305B consist of amphibolite and mica schist, respectively, and were collected 70 metres apart (figs. 3.5a, 3.5b). The amphibolite sample is medium-grained and consists of brown to light green zoned amphibole, plagioclase and quartz. The mica schist sample shows a penetrative foliation marked by fine-grained muscovite (sericite), quartz, garnet and chlorite. Both amphibole experiments from sample 06305 show variably staircase-shaped age spectra in the first 40-50% of $^{39}Ar_K$ degassing, with a pseudo-plateau (24.7% of $^{39}Ar_K$) at ca. 375 Ma for the most disturbed one, which is also linked to the lowest measured $^{37}Ar_{Ca}/^{39}Ar_K$ ratios (fig. 3.4b). The less disturbed amphibole experiment yields a high-temperature pseudo-plateau age at 462.4 ± 0.8 Ma supported by a flat $^{37}Ar_{Ca}/^{39}Ar_K$ segment with highest measured values. The most disturbed experiment yields a slightly younger pseudo-plateau age at 457.4 ± 1.0 Ma corresponding to a less regular $^{37}Ar_{Ca}/^{39}Ar_K$ segment, which could suggest a slight persistence of a disturbance in the high-temperature steps (fig. 3.4b). These results are coherent with petrographic observation and microprobe analysis of the dated sample (Daoust, 2007), which indicate that the amphiboles are zoned and characterized by Ca-poor and Ca-rich end members.

Sericite from sample 06305B also displays a staircase-shaped age spectrum, with low-temperature steps at 376.8 ± 1.6 Ma (fig. 3.4b), whereas the high-temperature steps define a pseudo-plateau age at 463.8 ± 0.5 Ma, that is concordant with the amphibole age.

Samples PH1 and CB1 consist of mica schist similar to the one of sample 06305B. Sample PH1 was collected ca. 3 metres beneath the unconformity marking the base of the Saint-Daniel Mélange (figs. 3.5c, 3.5d). Both samples, PH1 and CB1, yield straircase-shaped age spectra, with an overprint that is much more pronounced for sample CB1 (fig. 3.4c). Both experiments from sample CB1 yield reproducible age spectra that define high-temperature pseudo-plateau ages at 452.1 ± 0.5 Ma and 456.4 ± 0.8 Ma (fig. 3.4c). On the other hand, sample PH1, with a less disturbed age spectrum, displays older high-temperature pseudo-plateau ages, up to 467.0 ± 0.5 Ma (fig. 3.4c).

3.2.2 The Rivière-des-Plante ultramafic Complex

The Rivière-des-Plante ultramafic Complex (De Souza and Tremblay, 2010a) lies along the Baie Verte-Brompton line (fig. 3.1). It is bounded on the northwest by a northwest-dipping D_3

backthrust fault, and is unconformably overlain by sedimentary rocks belonging to the Saint-Daniel Mélange to the southeast (fig. 3.6). It comprises harzburgite, serpentinite, ophicalcite, gabbro and granite, and has been interpreted as an eroded ophiolitic remnant comprising mantle peridotites that correlate with those of the Thetford-Mines ophiolite (De Souza and Tremblay, 2010a). The Rivière-des-Plante ultramafic Complex is also characterized by the occurrence of peraluminous granites that have been divided into two textural sub-types: xenolith-free type 1 and xenolith-bearing type 2 granites (De Souza and Tremblay, 2010a).

Type 1 granites occur as small intrusions (< 100 m) in the southwestern part of the complex (fig. 3.6), or as dykes and fault-bounded bodies less than 10 metres wide. They are equigranular to porphyritic, medium- to coarse-grained, locally foliated and often rodingitized along their margins. In thin section, these rocks are made up of euhedral to sub-euhedral plagioclase, quartz, K-feldspar, biotite, muscovite and minor zircon, apatite and oxides. Type 1 granites are peraluminous and have a normative composition of granite *sensu stricto*. Type 2 granites consist of randomly-oriented xenoliths dispersed in a medium-grained felsic matrix composed of K-feldspar, plagioclase, muscovite, reddish-brown biotite, zircon and apatite. Chlorite and sericite aggregates are found as pseudomorphs after cordierite (Cousineau, 1991; Trzcienski, Rodgers and Guidotti, 1992), and garnet has been locally observed. The xenoliths consist of subrounded-to-angular, schistose and gneissic metasedimentary rocks and amphibolite. The mineralogical assemblages of the matrix and xenoliths suggest emplacement of these granitic rocks under low-pressure conditions, i.e. less than 3 kbar (Trzcienski, Rodgers and Guidotti, 1992). Type 2 granite is frequently foliated and locally deformed into a gneissic mylonitic facies (De Souza and Tremblay, 2010a). The sedimentary breccias marking the base of the Saint-Daniel Mélange are locally entirely made up of pebble- to boulder-size fragments of foliated to undeformed type 1 and type 2 granites, indicating that cooling and deformation of the granitic rocks must have preceded the sedimentation of the mélange.

$^{40}Ar/^{39}Ar$ results

The two analyzed samples that were collected in the Rivière-des-Plante ultramafic Complex consist of a type 1 (sample 08M31B) and a mylonitic type 2 (RPOM02) granite (see fig. 3.6 for location). Sample 08M31B is from an undeformed, coarse-grained porphyritic granite made up of quartz, sericitized plagioclase, K-feldspar, muscovite and chloritized biotite. Muscovite forms randomly-oriented interstitial crystals with respect to feldspar and does not show evidence of dynamic recrystallization. The two analyzed muscovite grains from this sample yield distinct age spectra, a flat one with a plateau age at 465.5 ± 0.5 Ma, whereas the second one displays a slightly disturbed age spectrum with a high-temperature pseudo-plateau age at 461.2 ± 0.4 Ma (fig. 3.7a).

Sample RPOM02 was collected in a site where the mylonitized facies of type 2 granite is best exposed and developed, and in fault contact with serpentinite. The sample shows a porphyroclastic mylonitic texture that is overprinted by a northwest-dipping cleavage genetically-related to SE-directed backthrusting (De Souza and Tremblay, 2010a). Two muscovite grains from sample RPOM02 were analyzed. One of them yields a staircase-shaped age spectrum that shows an increase of apparent ages from ca. 461.5 Ma in the low-temperature steps, and up to ca. 466 Ma (pseudo-plateau at 465.8 ± 1.0 Ma) in the high-temperature steps (fig. 3.7b). The duplicated experiment displays a flat age spectrum with a plateau age at 461.1 ± 0.4 Ma, perfectly concordant with the low-temperature pseudo-plateau age (461.5 ± 0.6 Ma) of its alter ego (fig. 3.7b).

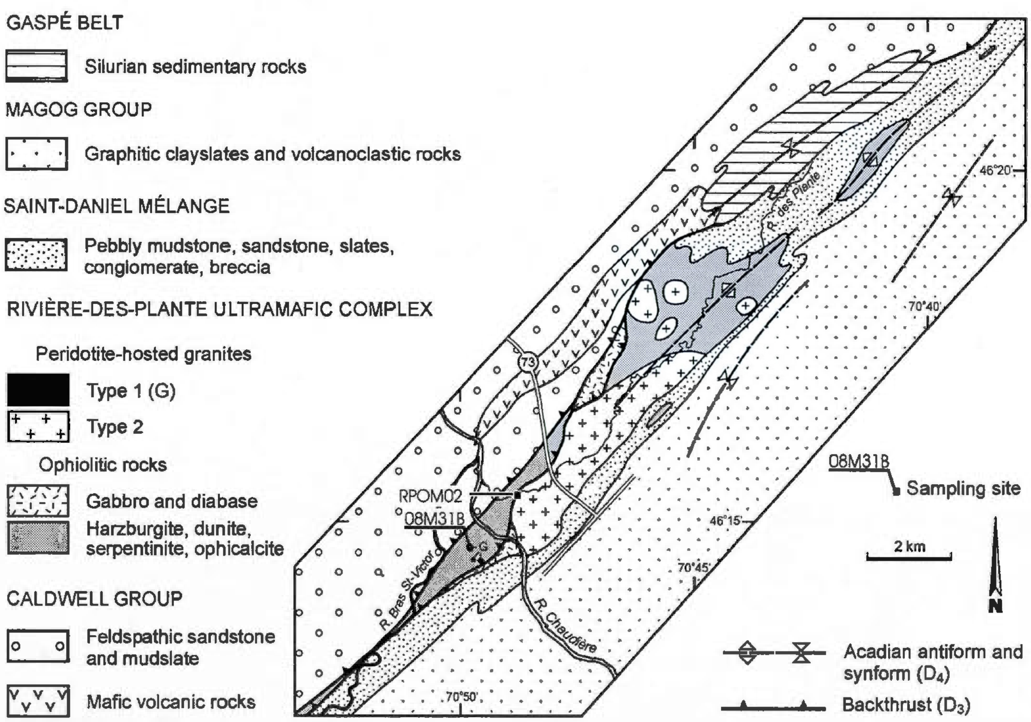

Figure 3.6 : Geological map of the Rivière-des-Plante ultramafic Complex showing the location of samples RPOM02 and 08M31B (modified from De Souza and Tremblay, 2010a).

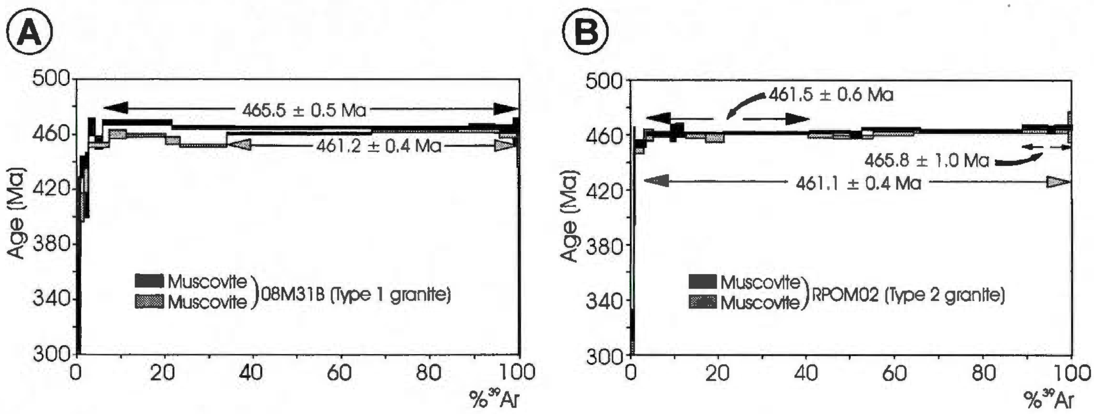

Figure 3.7 : $^{40}Ar/^{39}Ar$ age spectra for muscovite from samples 08M31B and RPOM02 (see fig. 3.6 for sample location).

3.2.3 The Nadeau Ophiolitic Mélange

The Nadeau Ophiolitic Mélange forms a 5 km-long and 700 m-wide lens-shaped inlier beneath the cover sequence of the Gaspé Belt, in the vicinity of the Maquerau-Mictaw inlier (fig. 3.8; De Brouker, 1987). It is made up of serpentinized peridotite, amphibolite, granitoid, mica schist and quartzite. De Brouker (1987) interpreted the peridotite as sheared serpentinite forming the matrix of a mélange, but our mapping of the area rather suggests that the ultramafic rocks and granitoids form part of an ophiolitic massif that is underlain by sole-type amphibolite and mica schist (figs. 3.8a, 3.8b, 3.8c). Peridotites of the Nadeau Ophiolitic Mélange consist of massive chromite-bearing dunite and harzburgite showing a high-temperature foliation typical of mantle rocks. The granitoids are massive or foliated, coarse- to fine-grained, and locally rodingitized at the contact with the host ultramafic rocks (De Brouker, 1987). They have the modal composition of granite, granodiorite or tonalite, with varying amounts of muscovite and biotite. Their overall composition and petrographic characteristics suggest that they are peridotite-hosted granites similar to those crosscutting the mantle rocks of the Thetford-Mines ophiolite. Granitoids intrude the metamorphic sole rocks and the ultramafics, a critical relationship that is not documented in southern Quebec (fig. 8d).

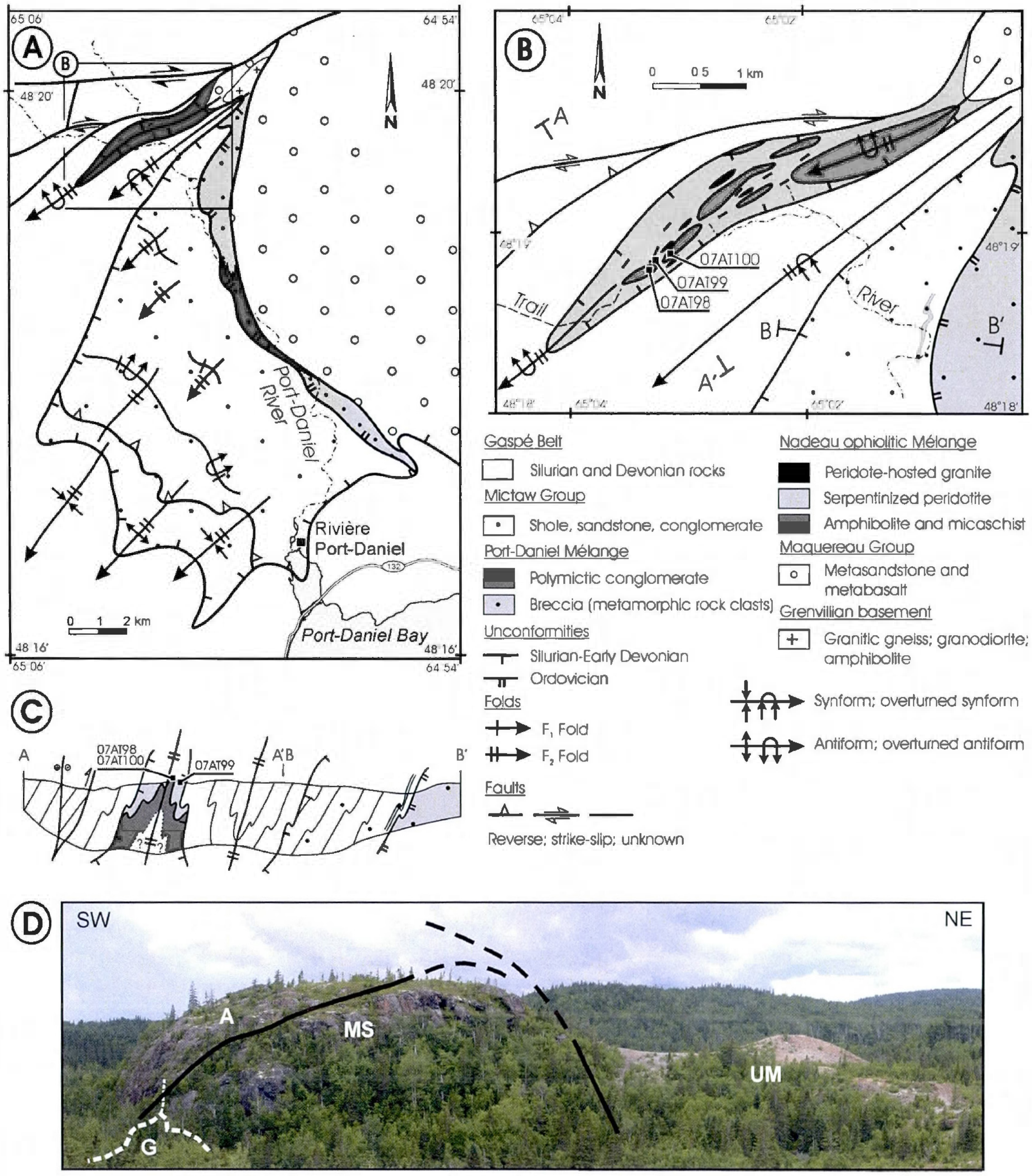

Figure 3.8 : a) Simplified geological map of the Maquereau-Mictaw inlier and Nadeau Ophiolitic Mélange (modified from De Brouker (1987)). b) Geological map and c) cross-section of the Nadeau Ophiolitic Mélange and adjacent rock units showing the location of samples 07AT98, 07AT99 and 07AT100 (geology from this study and compiled from De Brouker (1987)). d) Photomontage showing the ultramafic rocks (UM), mica schist (MS), amphibolite (A) and peridotite-hosted granite (G) of the Nadeau Ophiolitic Mélange.

The antiformal culmination in which is exposed the Nadeau Ophiolitic Mélange is attributed to the Acadian orogeny. It corresponds to northeast-trending folds in the Mictaw Group and Gaspé Belt (fig. 3.8). The metamorphic fabric in the mica schist and amphibolite is interpreted as the result of Taconian, obduction-related metamorphism. A breccia unit located along the strike of, and bounding the Rivière-Port-Daniel Mélange, was previously interpreted as a tectonic breccia related to the Rivière-Port-Daniel fault (Williams and St-Julien, 1982; De Brouker, 1987). It consists of angular to rounded fragments of foliated metasandstone, granitic gneiss and chloritic schist, and is interlayered with sedimentary rocks of the Mictaw Group (De Brouker, 1987). These gradational contacts rather suggest that the breccia is of sedimentary origin and formed close to a topographic high including metamorphic rocks of the Maquereau Group. The breccia unit should thus be included in the Rivière-Port-Daniel Mélange, as illustrated in fig. 3.8a. Moreover, the coexistence of ophiolitic and metasandstone fragments in the Rivière-Port-Daniel Mélange indicates that the source rocks consisted of both Humber and Dunnage zones lithologies.

$^{40}Ar/^{39}Ar$ results

The samples that were collected in the Nadeau Ophiolitic Mélange for $^{40}Ar/^{39}Ar$ analysis consisted of coarse-grained muscovite-bearing granite (07AT99), mica schist (07AT100) and amphibolite (07AT98). The mica schist is intimately associated with the amphibolite, and consists of quartz, plagioclase, alkali feldspar, muscovite, biotite, chlorite, zircon, sphene, apatite and garnet (De Brouker, 1987). The two analyzed amphiboles from sample 07AT98 have yielded strongly disturbed and inconclusive age spectra (not shown). The muscovite from the mica schist displays, however, a perfect plateau age at 470.0 ± 0.4 Ma (fig. 3.9a). The two analyzed muscovites from the granite provided slightly but notably distinct age spectra, a flat one with a plateau age at 475.1 ± 0.3 Ma (fig. 3.9a), whereas the second one presents a subtle but characteristic saddle shape with low- and high-temperature apparent ages at ca. 474.5 Ma (mean at 474.5 ± 0.6 Ma) and a saddle minimum at 471.9 ± 0.7 Ma (fig. 3.9b).

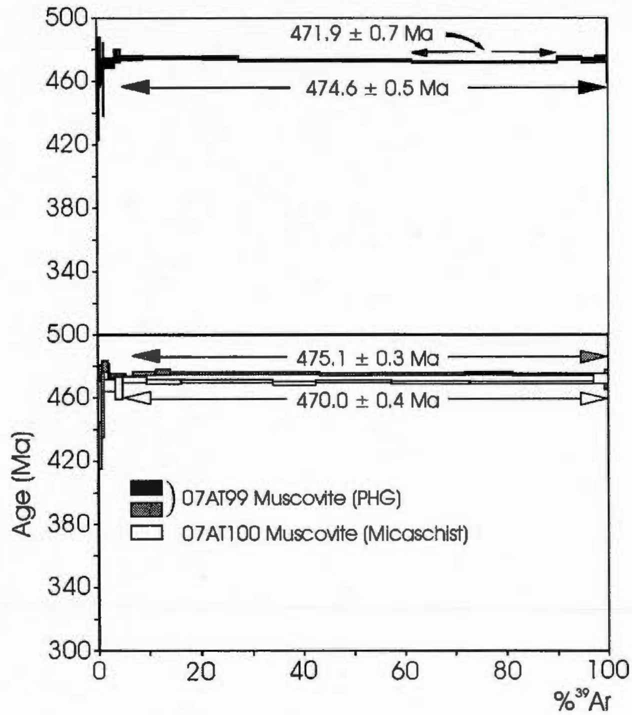

Figure 3.9 : $^{40}Ar/^{39}Ar$ age spectra for muscovite from samples 07AT99 and 07AT100.

3.3 Interpretation and synthesis

A compilation of isotopic age data for peridotite-hosted granites, ophiolitic and infraophiolitic metamorphic rocks, as well as biostratigraphic age constraints for the Magog and Mictaw groups is shown in fig. 3.10, and will be used in the following sections to synthesize and discuss the $^{40}Ar/^{39}Ar$ data presented herein.

3.3.1 Ophiolitic gabbros

There are no U-Pb ages for the crustal sequence of the Lac-Brompton ophiolite, but valuable time constraints for its crystallization and cooling history can be inferred from our $^{40}Ar/^{39}Ar$ age data and correlation with the Thetford-Mines ophiolite. An amphibole plateau age of 476.7 ± 6.2 Ma for the ophiolitic gabbro (6304) provides the best minimum estimate for its formation. The Thetford-Mines and Lac-Brompton ophiolites have been interpreted as originating from the same oceanic slab (e.g., Schroetter, Tremblay and Bédard, 2005; De Souza et al., 2008; Tremblay, Meshi and Bédard, 2009). Such an interpretation and correlation are supported by the concordance of our ca. 477 Ma amphibole age with the mean $^{40}Ar/^{39}Ar$ amphibole cooling age of

477.6 ± 3.5 Ma and U-Pb zircon crystallization age of 479.2 ± 1.6 Ma yielded by plagiogranites and gabbros of the Thetford-Mines ophiolite (fig. 3.10; Tremblay, Ruffet and Bédard, 2011; Whitehead, Dunning and Spray, 2000). This suggests that both ophiolitic bodies share a similar cooling history and, therefore, formed more-or-less synchronously at ca. 479 Ma. However, the lack of plutonic facies in the ophiolitic complexes of the Gaspé Peninsula inhibits further correlations and comparisons with the cooling and crystallization history of the Mont-Albert Complex and Nadeau Ophiolitic Mélange.

3.3.2 Metamorphic sole rocks

Samples from the metamorphic sole of the Lac-Brompton ophiolite consistently yield amphibole and sericite age spectra showing evidence of thermal overprinting and/or recrystallization. In spite of such overprint, these minerals still yield older increments with amphibole and sericite pseudo-plateau ages at ca. 463 Ma for samples 06305 and 06305B, and as old as ca. 467 Ma for mica schist PH1. The slightly younger pseudo-plateau ages between ca. 457 and 452 Ma yielded by sericite and amphibole of samples CB1 and 06305, together with pseudo-plateau and low-temperature step ages at ca. 376 Ma calculated from samples 06305 and 06305B, indicate that such young ages are clearly related to an Acadian disturbance (fig. 3.4c). Moreover, the ca. 4 Ma age discrepancy between the sericite pseudo-plateau ages of sample CB1 and the fact that they are younger than the inferred age of both the Magog Group and the Saint-Daniel Mélange (see below) suggest that, probably due to $^{40}Ar^*$ loss, these ages represent minimum estimates. This is also consistent with the compositional zonation of the amphiboles and the persistence of variable $^{37}Ar_{Ca}/^{39}Ar_K$ ratios into the high-temperature steps of experiment 06305-1 (fig. 3.4b). The Ca-poor component observed in the low-temperature steps probably represents partial recrystallization of previously-formed Ca-rich amphibole. On the other hand, sericite samples PH1 and 06305B, and amphibole experiment 06305 display less disturbed age spectra with older high-temperature pseudo-plateau ages up to 467.0 ± 0.5 Ma (fig. 3.4c). The latter sericite age represents the best estimate for the cooling of the metamorphic sole rocks below the closure temperature of muscovite. The correlation of a medium to high-temperature pseudo-plateau at 462.4 ± 0.8 Ma with a flat $^{37}Ar_{Ca}/^{39}Ar_K$ segment for amphibole experiment 06305, and its concordance with the high-temperature sericite age of 463.8 ± 0.5 Ma for sample 06305B (fig. 3.4b), indicate that the sole rocks were locally affected by an episode of thermal overprinting and/or recrystallization at ca. 463 Ma. It can thus be suggested that the sole amphibolites and mica schists initially crystallized in Early (?) to Middle Ordovician times, were exhumed at or prior to ca. 467 Ma, and then transported over the Laurentian margin and locally metamorphosed at ca. 463 Ma.

The $^{40}Ar/^{39}Ar$ ages and the inferred series of events for the infraophiolitic metamorphic rocks of the Lac-Brompton ophiolite are consistent with the crystallization and cooling history of the metamorphic sole of the Thetford-mines ophiolite, as reported by Tremblay, Ruffet and Bédard (2011). The latter rocks yield $^{40}Ar/^{39}Ar$ ages of ca. 471 Ma on amphibole and of 466 Ma on muscovite, with evidence for recrystallization as young as ca. 460-457 Ma. Such ages have been interpreted as representing the successive cooling of the metamorphic sole below the closure temperature of amphibole and muscovite, respectively, and later recrystallization of the muscovites during continental thrusting of the ophiolite and uplifting of the collisional orogenic wedge (Tremblay, Ruffet and Bédard, 2011).

The 470.0 ± 0.4 Ma muscovite plateau age from mica schists of the Nadeau Ophiolitic Mélange (07AT100), may result either from recrystallization or cooling below the closure temperature of muscovite, indicating that these rocks were metamorphosed and thrust onto the Laurentian margin with the overlying ophiolitic rocks as early as late Arenig. Data presented in this study for the Nadeau Ophiolitic Mélange are therefore significantly older than the available isotopic age constraints for the Mont-Albert Complex (fig. 3.10), where the metamorphic sole rocks yield amphibole and muscovite cooling ages of ca. 465 Ma to 459 Ma (fig. 3.10; c.f., Malo et al., 2008).

3.3.3 Peridotite-hosted granites

Petrographic and $^{40}Ar/^{39}Ar$ data on muscovites from the peridotite-hosted granites of the Rivière-des-Plante ultramafic Complex, and comparison with age data for those of the Thetford-Mines ophiolite (fig. 3.10), clearly show that they have undergone a polyphase cooling and crystallization history. The 08M31B plateau age at 465.5 ± 0.5 Ma and the concordant RPOM02 high-temperature pseudo-plateau at 465.8 ± 1.0 Ma for undeformed type 1 and mylonitic type 2 granites, respectively, suggest that the Rivière-des-Plante granites cooled below the closure temperature of muscovite at ca. 466 Ma. In both cases, the age and spectra shape discrepancies of duplicated analyses and the staircase-shaped age spectrum of the first RPOM02 muscovite experiment, indicate that the studied samples record a heterogeneous disturbance at ca. 460 Ma, regardless of deformation. Although it remains to be confirmed, such a disturbance could be the result of heterogeneous fluid-rock interactions with partial to complete induced recrystallizations (Tartèse et al., 2011).

The age spectra yielded by the granitic rocks of the Rivière-des-Plante ultramafic Complex are almost identical to those obtained for similar granitoids of the Thetford-Mines ophiolite, which suggest initial cooling of muscovite and later recrystallization at ca. 466 Ma and 461 Ma, respectively (Tremblay, Ruffet and Bédard, 2011). Such similarities in the $^{40}Ar/^{39}Ar$ ages, not only indicate a common cooling and crystallization history for these granitic rocks, but further support the inferred correlation of ophiolitic rocks and granitoids of both complexes. This, as well as the mean 469.5 ± 2.8 Ma U-Pb age yielded by peridotite-hosted granites of the Thetford-Mines ophiolite (Whitehead, Dunning and Spray, 2000), suggests that granitic rocks of the Rivière-des-Plante ultramafic Complex also initially crystallized at ca. 470 Ma.

However, muscovites from a peridotite-hosted granitoid of the Nadeau Ophiolitic Mélange yielded significantly older plateau, and high- to low-temperature step ages of ca. 475 Ma. The saddle-shaped age spectrum of the duplicate experiment of sample 07AT99 is evidence for a disturbance at or slightly after ca. 472 Ma. The partial re/neocrystallization of white micas can generate saddle-shaped age spectra that are the result of distinctive degassing patterns of initial/inherited and re/neocrystallized domains for a given crystal (e.g., Cheilletz et al., 1999; Alexandrov, Ruffet and Cheilletz, 2002). According to these authors and in good agreement with plateau ages yielded by (i) duplicated muscovite experiment and (ii) mica schist muscovite from the metamorphic sole, the initial crystallization or cooling of the 07AT99 muscovite would have occurred at ca. 475 Ma or earlier, and it would have recorded a partial recrystallization history linked to deformation and/or fluid circulation at or slightly later than ca. 472 Ma.

Our data and compiled U-Pb and $^{40}Ar/^{39}Ar$ age constraints for peridotite-hosted granites of southern Quebec and the Nadeau Ophiolitic Mélange highlight significant discrepancies in their crystallization and cooling history from one area to the other. Considering the nature of the dated mineral species and the isotopic dating methods that were used, it can be concluded that granitic rocks of the Nadeau Ophiolitic Mélange cooled below the closure temperature of muscovite ca.

10 Ma before those of southern Quebec and probably crystallized more than 3-to-5 Myr earlier (fig. 3.10).

3.4 Discussion

As previously mentioned, the obduction of suprasubduction, Tethyan-type ophiolites, can be defined as a series of processes leading to the final emplacement of oceanic lithosphere onto an adjacent continental margin. These are (1) intra-oceanic subduction and formation of an ophiolite and its metamorphic sole, (2) exhumation of the metamorphic sole, (3) thrusting of the ophiolitic nappe over the continental margin, and (4) its subaerial exposure as part of an orogenic belt (Wakabayashi and Dilek, 2003). Obduction *sensu stricto* can be considered as ending with the syncollisional uplift and erosion of the ophiolite and underlying metamorphic rocks (Tremblay, Ruffet and Bédard, 2011). This can be accomplished once the ophiolitic nappe is being uplifted and transported passively on top of a foreland-propagating thrust system, as continental and/or oceanic material is progressively underplated beneath it (Searle and Cox, 1999; Bortolotti et al., 2005; Cloos et al., 2005; Schroetter et al., 2006; Tremblay, Ruffet and Bédard, 2011). In this tectonic framework, the main issues that will be discussed in the following sections are the correlation of ophiolitic complexes along the Baie Verte-Brompton line, the chronology of events related to ophiolite obduction onto Laurentia, and the exhumation processes leading to mélange formation and termination of obduction.

3.4.1 Obduction diachronism in the Quebec Appalachians

In the Quebec Appalachians, major ophiolitic complexes and serpentinite bodies marking the Baie Verte-Brompton line have been interpreted as more-or-less correlative fragments of obducted oceanic lithosphere formed in short-lived subduction-related pericontinental marginal basins (fig. 3.11; Hébert and Bédard, 2000; Huot, Hébert and Turcotte, 2002; Schroetter et al., 2003; De Souza et al., 2008; Malo et al., 2008; Tremblay, Meshi and Bédard, 2009). The petrography, geochemical composition and isotopic ages of infraophiolitic metamorphic rocks, all suggest that these are mostly derived from oceanic and/or continental margin mafic igneous protoliths with minor amounts of interlayered sedimentary rocks (Clague, Rubin and Brackett, 1981; Feininger, 1981; Gagnon and Jamieson, 1986; De Brouker, 1987; Daoust, 2007). Since $^{40}Ar/^{39}Ar$ ages for metamorphic soles are considered as cooling ages, and that sole amphibolites cool rather rapidly below the closure temperature of amphibole, the $^{40}Ar/^{39}Ar$ amphibole ages for such metamorphic rocks are more probably 1 to 5 Myr younger than peak metamorphism (Hacker, 1990, 1991; Hacker, Mosenfelder and Gnos, 1996). The sole protoliths were likely overthrust and metamorphosed during nascent stages of obduction, after the crystallization of the ophiolites between ca. 479 and 472 Ma in southern Quebec, and between ca. 470 and 466 Ma for the Mont-Albert Complex. These metamorphic sole rocks were then exhumed and successively cooled below the closure temperatures of amphibole and muscovite, from ca. 471 to 467-466 Ma in southern Quebec, and 465 Ma to 459 Ma in the Mont-Albert Complex. Although no $^{40}Ar/^{39}Ar$ amphibole age has been obtained for the Nadeau Ophiolitic Mélange, our $^{40}Ar/^{39}Ar$ ca. 470 Ma muscovite age of the micaschists (sample 07AT99) suggests that the sole amphibolite there was uplifted and cooled below the muscovite closure temperature 3-to-12 Myr earlier than in other infraophiolitic metamorphic rocks of the Quebec Appalachians.

A minimum age for underthrusting and anatectic melting of the Laurentian margin beneath the obducting ophiolites is provided by the mean 469.5 ± 2.8 Ma U-Pb zircon crystallization age of the peridotite-hosted granites belonging to the Thetford-Mines ophiolite (Whitehead, Dunning and Spray, 2000), and the ca. 475 Ma $^{40}Ar/^{39}Ar$ muscovite cooling age we obtained for similar rocks of the Nadeau ophiolitic Mélange. As emphasized by Whitehead, Dunning and Spray (2000) and Tremblay, Ruffet and Bédard (2011), up to ca. 2 Myr of shear heating at the base of the ophiolite nappe may have been necessary to account for partial melting of continental margin siliciclastic rocks. This suggests that, in southern Quebec, underthrusting of the Laurentian margin would have been initiated at ca. 471 Ma or slightly earlier, at least 5 Myr prior to the cooling of the peridotite-hosted granites below the muscovite closure temperature at 465-466 Ma. By applying the same reasoning to the Nadeau Ophiolitic Mélange granitoids, it can be inferred that underthrusting and anatectic melting of the continental margin there was initiated well before 475 Ma, possibly earlier than ca. 480 Ma. Following their initial emplacement onto Laurentia, the ophiolitic nappes were actively translated over the margin at least until ca. 460-457 Ma in southern Quebec and in the Mont-Albert Complex, as suggested by $^{40}Ar/^{39}Ar$ muscovite and amphibole ages related to the latest increments of recrystallization and/or cooling of the infraophiolitic metamorphic rocks from both areas (fig. 3.10). By the latest Arenig-early Caradoc, active thrusting and shear deformation of the ophiolitic nappes were progressively transferred to lower thrust slices within the continental margin, whereas the uplifting of the ophiolites and underlying metamorphic sole rocks was more-or-less completed.

The U-Pb and $^{40}Ar/^{39}Ar$ geochronological constraints for ophiolitic complexes of the Quebec Dunnage zone therefore suggest that orogenic processes leading to the Taconian obduction of ophiolitic nappes onto Laurentia may have been diachronous along the strike of the orogenic belt. Unequivocal evidence for interactions between Laurentia and ophiolitic nappes between ca. 480 and 475 Ma in the Nadeau Ophiolitic Mélange suggests that ophiolite emplacement was initiated earlier in the vicinity of the St-Lawrence promontory than in the central segment of the Quebec embayment. $^{40}Ar/^{39}Ar$ muscovite ages also indicate a significant ca. 3-to-10 Myr delay in the cooling history of the metamorphic soles and peridotite-hosted granites from both areas, and ca. 5 Myr between amphibole and muscovite ages for the sole rocks of the southern Quebec ophiolites and the Mont-Albert Complex. Such variations in the age data can be tentatively attributed to the inherited irregular geometry of the Laurentian margin that developed in Early Paleozoic times (Thomas, 1977). Obduction-related deformation and metamorphism were likely older at the periphery of the Quebec embayment than in its central segment (fig. 3.1a), where obduction was vanishing by ca. 460-457 Ma and lasted between approximately 10 to 22 Myr. Using a conservative convergence rate of 1 cm/year, it can be extrapolated that the Quebec ophiolites are therefore rooted at a minimum distance of 100 km to the southeast (present coordinates) of their current location.

3.4.2 Syn-Taconian exhumation and sedimentation in the Quebec Appalachians

Emplacement of ophiolites onto a continental margin ultimately leads to their exhumation and subaerial exposure, and to the formation of syncollisional sedimentary basin(s) and olistostromal mélanges derived from the erosion of the uplifted orogenic wedge (Gray, Gregory and Miller, 2000; Bortolotti et al., 2005; Cloos et al., 2005; Tremblay, Ruffet and Bédard, 2011). Examples of such sedimentary units are the Saint-Daniel and Rivière-Port-Daniel mélanges. Both are characterized by sedimentary breccias and conglomerates that comprise ophiolitic, metamorphic and sedimentary rock fragments, and that are interlayered with mudstone and chert (De Brouker, 1987; Schroetter et al., 2006; De Souza et al., 2008; Tremblay, Meshi and Bédard, 2009). Metasedimentary rock fragments in the Saint-Daniel Mélange yielded $^{40}Ar/^{39}Ar$ muscovite ages varying between 467 Ma and 463 Ma in the Thetford-Mines area (fig. 3.10; Schroetter et al., 2006; Tremblay, Ruffet and Bédard, 2011), suggesting rapid uplifting and recycling of the Taconian orogenic wedge. Graptolite-bearing slates and sandstones overlying the Saint-Daniel and Rivière-Port-Daniel mélanges belong to the Llanvirn and early Caradoc Magog and Mictaw groups (fig. 3.10; Riva, 1974; De Brouker, 1987; Cousineau, 1990). The age of the Saint-Daniel Mélange can be therefore tightly bracketed in the 463-461 to 456 Ma interval, whereas a minimum age of ca. 462 Ma can be suggested for the Rivière-Port-Daniel Mélange (fig. 3.10). Mélange formation and syncollisional basin development were thus initiated shortly earlier in the Gaspé Peninsula than in southern Quebec, which is consistent with the inferred diachronism in the obduction history of both areas.

Further evidence for syn-obduction sedimentation is provided by comparing the isotopic age constraints presented herein with the stratigraphic record of the Appalachian foreland of the Gaspé Peninsula. There, the late Arenig – early Llanvirn Tourelle Formation (fig. 3.10; Biron, 1972), located ca. 35 km to the NE of the Mont-Albert Complex, is the oldest of a series of chromite- and mafic detritus-bearing flysch units (figs. 3.1a, 3.2; Hiscott, 1978, 1995). Paleocurrent and petrographic provenance data for the Tourelle Formation suggest that detritus was transported westward into a foredeep basin, away from the St-Lawrence promontory, and was derived from the erosion of a Grenville-type basement, variously-metamorphosed continental margin rocks and ophiolitic rocks (Hiscott, 1978). The youngest ophiolite clast-bearing flysch units of the Gaspé Peninsula foreland are the Caradoc Deslandes Formation and the lower part of the Cloridorme Formation (figs. 3.1a, 3.2; Prave et al., 2000). Again, this is consistent with the onset of ophiolite exhumation at 475-470 Ma in the vicinity of the St-Lawrence promontory, the final emplacement of the Mont-Albert Complex at 459-457 Ma, and with the obduction diachronism highlighted above (fig. 3.10).

We believe that a similar suturing history took place along the entire length of the Quebec Appalachians. Obduction of the ophiolites onto the Laurentian margin occurred along northwest-directed and shallow-dipping thrust surfaces, thereby creating an almost flat-lying suture zone (fig. 3.11). The syn-obduction uplift and erosion of an accretionary ridge(s) made up of ophiolitic and metamorphic rocks led to the formation of olistostromal mélanges as a result of mass wasting on top of the advancing ophiolite nappe(s), and to the sedimentation of foredeep successions. Finally, the progressive deepening and stabilization of the basin during the waning stages of obduction led to the deposition of thick onlapping flysch sequences represented by the Magog and Mictaw groups. As shown on fig. 3.11, the existence of late- to post-Ordovician unconformities, the occurrence of Silurian – Early Devonian normal faults, as well as the superposed Acadian deformation (Tremblay and Pinet, 2005; Malo et al., 2001, 2008) account for missing sections and additional structural complexities in the pre-Silurian stratigraphic record of the Quebec Appalachians.

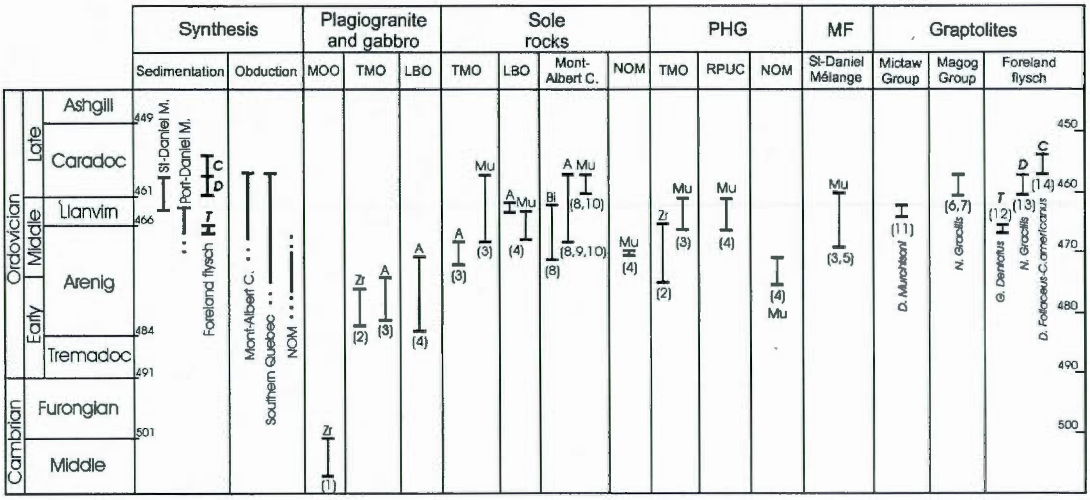

Figure 3.10 : Compilation and synthesis of U-Pb and $^{40}Ar/^{39}Ar$ isotopic age data for ophiolitic rocks, infraophiolitic metamorphic sole rocks and peridotite-hosted granites of the southern Quebec and Gaspé Peninsula Appalachians, as well as biostratigraphic age constraints for the Magog and Mictaw groups, and the Trourelle, Deslandes and Cloridorme formations. Age intervals shown in this figure take into account the uncertainties related to the isotopic age data. LBO – Lac-Brompton ophiolite; MF – Metamorphic rock fragments; MOO – Mont-Orford ophiolite; NOM – Nadeau Ophiolitic Mélange; PHG – Peridotite-hoste granites; RPUC – Rivière-des-Plante ultramafic Complex; TMO – Thetford-Mines ophiolite. Stratigraphic units : C – lower Cloridorme Formation; D – Deslandes Formation; T – Trourelle Formation; $^{40}Ar/^{39}Ar$ data : A – Amphibole; Mu – Muscovite; U-Pb data : Zr – Zircon. Data sources : (1) David and Marquis (1994); (2) Whitehead, Dunning and Spray (2000); (3) Tremblay, Ruffet and Bédard (2011); (4) this study; (5) Schroetter et al. (2006); (6) Riva (1974); (7) Cousineau (1990); (8) Malo et al. (2008); (9) Lux (1986); (10) Pincivy et al. (2003); (11) De Brouker (1987); (12) Hiscott (1978); (13) Riva (1968) and Bloechl (1996); (14) Prave et al. (2000). Time scale from Sadler, Cooper and Melchin (2009) and Gradstein, Ogg and Smith (2004), and graptolite zones from Webby et al. (2004).

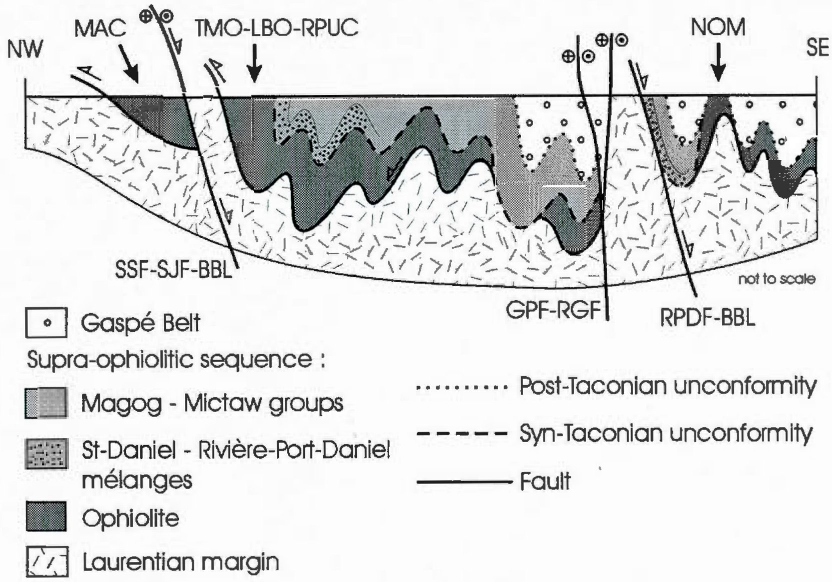

Figure 3.11 : Composite and schematic interpretative section across the post-Acadian Laurentian margin of the Quebec Appalachians showing the location and stratigraphic position of the various ophiolitic complexes, supra-ophiolitic sedimentary sequences and Gaspé Belt basin, as well as major structural features and unconformities. BBL – Baie Verte-Brompton line; GPF – Grand Pabos fault; LBO – Lac-Brompton ophiolite; MAC – Mont-Albert Complex; NOM – Nadeau Ophiolitic Mélange; RGF – Rivière Garin fault; RPDF – Rivière-Port-Daniel fault; RPUC – Rivière-des-Plante ultramafic Complex; SJF – Saint-Joseph fault; SSF – Shickshock Sud fault; TMO – Thetford-Mines ophiolite.

3.4.3 Comparisons with northern New England and Newfoundland

Ophiolites, mafic-to-ultramafic complexes, arc-related rocks and serpentinite slivers that occur in a structural setting similar to those of southern Quebec are also known along the strike of the Baie Verte-Brompton line (and correlative structures) in Newfoundland and New England (Church, 1977; Doolan et al., 1982; Williams and St-Julien, 1982; Karabinos et al., 1998; van Staal et al., 1998; Kim and Jacobi, 2002; van Staal, 2007). In the Baie-Verte Peninsula of western Newfoundland, for instance, Tremadocian ophiolitic rocks are overlain by the Flatwater Pond and Snooks Arm groups (Hibbard, 1983; Bédard et al., 2000; Skulski et al., 2010), both representing correlative units of an ophiolite cover sequence to which has been attributed a maximum age of ca. 479 Ma (Skulski et al., 2010). The Flatwater Pond and Snooks Arm groups comprise abundant mafic and felsic volcanic rocks that overlie iron formations and conglomerates. Megabreccias and olistostromes marking the base of the Flatwater Pond Group comprise ophiolite-derived fragments and metamorphic rock clasts attributed to erosion of the metamorphosed Laurentian continental margin (Kidd, 1974; Williams and St-Julien, 1982;

Skulski et al., 2010). West of the Baie Verte-Brompton line, Middle Cambrian to Tremadocian ophiolitic rocks belonging to the Bay of Islands Ophiolite and St-Anthony Complex (Williams and Smyth, 1973; Karson and Dewey, 1978; Suhr and Cawood, 2001) are underlain by metamorphic sole amphibolites that yielded $^{40}Ar/^{39}Ar$ amphibole ages of 469 ± 5 Ma and 489 ± 5 Ma, respectively (Dallmeyer and Williams, 1975; Dallmeyer, 1977; new decay constant), suggesting that obduction may have been initiated earlier in western Newfoundland than in Quebec, or that it overlaps in both areas. As for the Quebec Appalachians, stratigraphic relationships and age data from the Baie-Verte Peninsula indicate that mélange formation has been coeval with ophiolite obduction, although it was followed there by a much more voluminous amount of mafic volcanism (Bédard et al., 2000; Skulski et al., 2010) rather than Magog- or Mictaw-type flysch sedimentation.

In western New England, the tectonic units of the Baie Verte-Brompton line extend into the Rowe-Hawley belt, where arc-related and ophiolitic rocks are poorly-preserved as strongly dismembered amphibolite, peridotites and/or serpentinite slivers, gneisses and schists that were accreted to Laurentia during the Early to Middle Ordovician (Doolan et al., 1982; Stanley and Ratcliffe, 1985; Kim and Jacobi, 1996; Karabinos et al., 1998; Coish and Gardner, 2004; Coish, 2010). Moreover, the Saint-Daniel Mélange and the Magog Group are only locally present to totally absent from the geological record south of the Quebec-Vermont border. The lack of well-preserved ophiolites and related syncollisional deposits in western New England suggests that, in contrast to the Quebec and Newfoundland Appalachians, large-scale ophiolite obduction was probably of minor importance there during the Taconian orogeny, such an hypothesis being supported by a south-directed decrease in the Cr and Ni content of Ordovician foreland flysch units from the Canadian to the US Appalachians (Hiscott, 1984).

3.5 Conclusion

Geological and $^{40}Ar/^{39}Ar$ geochronological data for the Lac-Brompton ophiolite, the Rivière-des-Plante ultramafic Complex in southern Quebec, and the Nadeau Ophiolitic Mélange in the Gaspé Peninsula suggest that they form, along with the Thetford-Mines and Mont-Albert ophiolitic complexes, eroded remnants of a composite ophiolitic slab. Ophiolite obduction was initiated with the formation of metamorphic sole rocks between ca. 479 and 472 Ma in southern Quebec and possibly as late as 470-466 Ma in the Mont-Albert Complex, shortly after the crystallization of the ophiolitic rocks. Underthrusting of the Laurentian margin beneath the obducting ophiolites was first initiated along the margin of the St-Lawrence promontory between ca. 480 and 475 Ma, earlier than in southern Quebec at 472-470 Ma. The progressive thrusting of the ophiolites over the margin, between 471-457 Ma in southern Quebec and 465-457 Ma in the Mont-Albert Complex, has ultimately led to the uplift, erosion and unroofing of the orogenic wedge. This has resulted in the formation of the Saint-Daniel and Rivière-Port-Daniel mélanges as part of syncollisional sedimentary basins on top of the ophiolitic nappes, and foreland flysch successions between the latest Arenig(?) to early Caradoc. Diachronism in the tectono-sedimentary evolution along the strike of the Taconian orogeny in the Quebec Appalachians may be likely attributed to the irregular geometry of the Early Paleozoic Laurentian margin. Also, lithotectonic variations along the Baie Verte-Brompton line, from western New England to western Newfoundland, highlight additional complexities in the Taconian obduction history that may also be the result of an irregular collision zone.

3.6 Acknowledgements

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CHAPITRE IV

TACONIAN OROGENESIS, SEDIMENTATION AND MAGMATISM IN THE SOUTHERN QUEBEC – NORTHERN VERMONT APPALACHIANS : STRATIGRAPHIC AND DETRITAL MINERAL RECORD OF IAPETAN SUTURING

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ABSTRACT

Cambrian-Ordovician units of the southern Quebec Appalachians oceanic domain include obducted ophiolites and an overlying syncollisional sedimentary basin represented by the Saint-Daniel Mélange and flysch units of the Magog Group. These terrains were thrust onto the Laurentian continental margin during the Taconian orogeny and are unconformably overlain by Silurian-Devonian successor basins. In this article we present stratigraphic data mainly acquired in the oceanic domain, as well as U-Pb and $^{40}Ar/^{39}Ar$ geochronological data on detrital zircons and muscovites sampled at various stratigraphic levels, from the base of the sedimentary units of the oceanic domain, to the Silurian-Devonian units. This work, combined with a compilation of existing data for the southern Quebec and western New England Appalachians, allows for a better definition of the tectono-stratigraphic evolution of the studied units, and to revisit their correlation with rocks of the northern Vermont Rowe-Hawley belt, which represents a collage of Laurentian margin and oceanic units. It is shown that although well-preserved ophiolites and flysch units are lacking in the Rowe-Hawley belt, Cambrian-Ordovician rocks of southern Quebec and northern Vermont have undergone a similar tectonic evolution. The ophiolites were obducted onto the Laurentian margin as a more or less coherent slab of oceanic lithosphere. Both the ophiolites and continental margin rocks were uplifted, eroded and unconformably overlain by Middle to Late Ordovician units of the Saint-Daniel Mélange and Magog Group in a forearc setting. The presence of mafic and felsic volcanic flows that are interstratified within the Saint-Daniel Mélange can be correlated with an episode of delamination and slab breakoff during the final stages of obduction and exhumation of the orogenic wedge. The Magog Group was deposited during the emplacement and uplift of the Taconian nappes, as the outboard volcanic arc was still active but progressively shutting down. The provenances suggested by the age of the dated detrital minerals can be entirely reconciled with the peri-Laurentian context of the studied rock units and testify of the tectonic evolution and sedimentary recycling of the Taconian orogen, from ophiolite obduction onto the Laurentian margin, to the collapse of the orogenic belt and the formation of successor bassins.

4.1 Introduction

The Cambrian-Ordovician Taconian orogeny is a first order lithotectonic feature of the Northern Appalachians that involves the emplacement of large ophiolitic nappes onto Laurentia and the accretion of volcanic arcs and continental blocks as a result of the partial closure of the Iapetus Ocean and ongoing convergence between Laurentia and Gondwana (St-Julien and Hubert, 1975; Williams, 1979; Stanley and Ratcliffe, 1985; Tremblay, 1992a; Pinet and Tremblay, 1995b; Karabinos et al., 1998; van Staal et al., 1998; van Staal, 2007). Evidence for Taconian orogenesis and suturing is particularly well-preserved in the southern Quebec and northern Vermont Appalachians, which can be divided into three main lithotectonic assemblages: the Early Paleozoic Humber and Dunnage zones, and the overlying Silurian-Devonian successions of the Connecticut Valley-Gaspé trough (fig. 4.1; Williams, 1979; Bourque, Malo and Kirkwood, 2000; Lavoie and Asselin, 2004; Tremblay and Pinet, 2005; Rankin et al., 2007). The Humber and Dunnage zones were amalgamated during the taconic orogeny and represent, respectively, the vestiges of the Laurentian continental margin and the adjacent oceanic basin (Williams, 1979). Although many plate tectonic models have been proposed for the Early Paleozoic evolution of the southern Quebec-northern Vermont Appalachians, these are hindered by scarce geochronological age constraints and poor understanding of various lithostratigraphic components of the Dunnage zone, and of their significance in paleotectonic reconstructions (Doolan et al., 1982; Tremblay, 1992a; Pinet and Tremblay, 1995b).

In this contribution we present detailed observations and detrital zircon U-Pb and muscovite $^{40}Ar/^{39}Ar$ geochronological data for volcanic and sedimentary rocks of the Dunnage zone and Connecticut Valley-Gaspé trough of southern Quebec. It will be shown that these syn- to post-orogenic units form an almost continuous lithostratigraphic record of Ordovician to Devonian basin formation and ongoing tectonism. The stratigraphy, age, and provenance of the studied rock units can be reconciled with the regional stratigraphic framework of the southern Quebec-northern Vermont Appalachians, in order to propose a comprehensive tectonic model for this part of the Taconian orogen, from the emplacement of ophiolitic nappes onto Laurentia and formation of an onlapping forearc basin during the Ordovician, to the collapse of the thickened orogenic wedge and formation of successor basins in the Silurian to Early Devonian.

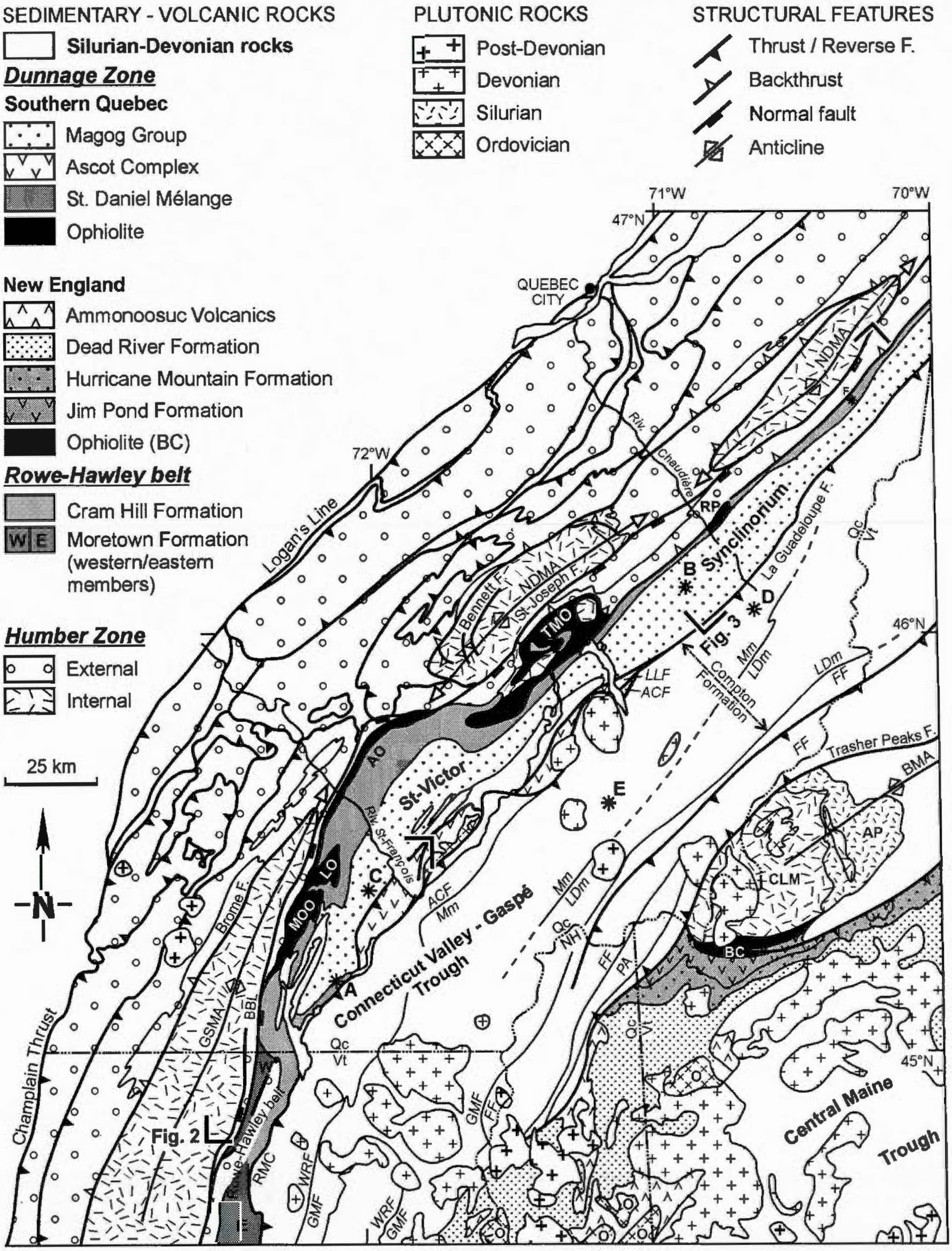


Figure 4.1 : Generalized geological map of the southern Quebec and western New England Appalachians (after Slivitzky and St-Julien, 1987; Moench et al., 1995; Tremblay and Castonguay, 2002; Moench and Aleinikoff, 2002; Castonguay, Tremblay and Lavoie, 2000, 2002; Hibbard et al., 2006; McWilliams, Walsh and Wintsch, 2010). The Rowe-Hawley belt of northern Vermont, as shown in this map, includes the Stowe, Moretown and Cram Hill formations. Plutonic and volcanic rocks attributed to the Bronson Hill arc include the Oliverian Plutonic Suite and Ammonoosuc Volcanics. Stratigraphic units of the Connecticut Valley – Gaspé trough are identified by labels in italic. Gaspé Belt : *ACF* – Ayers Cliff Formation; *FF* – Frontenac Formation; *LDm* – Lac Drolet member; *LL* – Lac Lambton Formation; *Mm* – Milan member. Connecticut Valley sequence : *GMF* – Gile Mountain Formation; *PA* –

Piermont allochthon; *WRF* – Waits River Formation. AO – Asbestos ophiolite; BBL – Baie Verte-Brompton line; BC – Boil Mountain Complex; BMA – Boundary Mountains Anticlinorium; GSMA – Green and Sutton Mountains anticlinorium; LO – Lac-Brompton ophiolite; MOO – Mont-Orford ophiolite; NDMA – Notre-Dame Mountains Anticlinorium; O – Oliverian Plutonic Suite; RMC – Richardson-Memorial contact; RP – Rivière-des-Plante ultramafic Complex; Thetford-Mines ophiolite; UT – Underhill Thrust. Asterisks indicate the location of the sampling sites for the U-Pb and $^{40}Ar/^{39}Ar$ data presented in this study: A – samples 07BUNKER and 10BUNKER01; B – 09SV01; C – 09SV02; D – 09MILAN01; E – 09MILAN02; F – 08M44.

4.1.1 The Southern Quebec – Northern Vermont Appalachians

In southern Quebec and northern Vermont, the Humber zone consists of a fold-and-thrust belt that is mostly represented by Neoproterozoic to Early Ordovician siliciclastic and minor basaltic rift and passive margin facies, as well as Middle to Late Ordovician syn-orogenic flysch units. In southern Quebec, these are further subdivided into external and internal zones composed of low- to subgreenschist-grade rocks and upper greenschist to amphibolite facies metamorphic rocks, respectively (fig. 4.1). The internal Humber zone is limited to the southeast by a major normal fault zone known as the Baie Verte-Brompton line – Saint-Joseph fault in southern Quebec (Pinet, Tremblay and Sosson, 1996; Tremblay and Castonguay, 2002; Castonguay and Tremblay, 2003), and the Burgess Branch fault zone in northern Vermont (Kim et al., 1999). Dunnage zone rocks are exposed to the southeast of the Baie Verte – Brompton line as part of the Saint-Victor synclinorium in southern Quebec and of the Rowe-Hawley belt in northern Vermont (fig. 4.1; Doolan et al., 1982; Tremblay, 1992a). The Connecticut Valley – Gaspé trough is located to the southeast of the La Guadeloupe fault and is underlain by Silurian-Devonian rocks that form part of the Gaspé Belt in southern Quebec (Tremblay and Pinet, 2005) and that can be mapped continuously southward into western New England, where it is informally referred to as the Connecticut Valley sequence (fig. 4.1; Rankin et al., 2007; McWilliams, Walsh and Wintsch, 2010).

4.1.1.1 The southern Quebec Dunnage zone

The Dunnage zone of southern Quebec includes (1) a series of well-preserved to dismembered ophiolite complexes; 2) the Saint-Daniel Mélange; 3) the Magog Group; and 4) the Ascot Complex (fig. 4.1).

The ophiolites occur on the northwestern limb of the Saint-Victor synclinorium and represent correlative remnants of obducted suprasubduction zone ophiolitic nappe(s) made up of oceanic crust and mantle, and infraophiolitic metamorphic rocks (Schroetter et al., 2003; Schroetter, Tremblay and Bédard, 2005; De Souza et al., 2008; De Souza and Tremblay, 2010a; Pagé et al., 2008; Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012). U-Pb zircon ages for ophiolitic plagiogranites vary from 504 ± 3 Ma in the Mont-Orford ophiolite (David and Marquis, 1994), to 478 +3/-2 and 480 ± 2 Ma in the Thetford-Mines ophiolite (Whitehead, Dunning and Spray, 2000). Syn-obduction anatectic granites that crosscut the mantle unit of the ophiolites yielded U-Pb zircon ages of 470 +5/-3 Ma and 469 ± 4 Ma in the Thetford-Mines area (Whitehead, Dunning and Spray, 2000). $^{40}Ar/^{39}Ar$ amphibole and muscovite cooling ages of ca. 471 Ma and ca. 466-467 Ma, respectively, were obtained for the infraophiolitic metamorphic rocks, which also show evidence of recrystallization down to ca. 460-457 Ma (Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012).

The Saint-Daniel Mélange is exposed on both limbs of the Saint-Victor synclinorium and was previously interpreted as an accretionary complex dominantly composed of an argillaceous-to-conglomeratic matrix with deca- to kilometre-scale olistoliths and tectonic slivers of ophiolitic, volcanic and Humber-like rock units, such as the Bolton Igneous Group and Bunker Hill sequence in the Lake Memphremagog area, and the Ware Volcanics in the Beauce area (figs. 4.1, 4.2, 4.3; Doolan et al., 1982; Cousineau, 1990; Cousineau and St-Julien, 1992; Tremblay, 1992). Recent work has rather shown that the Saint-Daniel Mélange represents the lower part of a syncollisional sedimentary basin that unconformably overlies various stratigraphic levels of the ophiolites including crustal, mantle and infraophiolitic metamorphic rocks (Schroetter et al. 2006; De Souza et al., 2008; De Souza and Tremblay, 2010a). It comprises a discontinuous basal unit consisting of debris flow breccias and conglomerates that are overlain by green and black slate and sandstone, and a characteristic pebbly mudstone facies (St-Julien 1987; Lavoie 1989; Cousineau, 1990; Cousineau and St-Julien, 1992; Schroetter et al., 2006; De Souza and Tremblay, 2010a). $^{40}Ar/^{39}Ar$ muscovite ages on micaschist clasts from the basal debris flows vary between ca. 467 Ma and 463 Ma (Schroetter et al., 2006; Tremblay, Ruffet and Bédard, 2011), which corresponds to the muscovite cooling ages yielded by the infraophiolitic metamorphic rocks in southern Quebec. It was thus concluded that the Saint-Daniel Mélange derives from the sedimentary recycling of uplifted ophiolites and related metamorphic rocks and represents the lower part of a syncollisional piggyback basin (Schroetter et al., 2006; Tremblay, Ruffet and Bédard, 2011).

The Magog Group comprises a ca. 10 km-thick Middle to Late Ordovician flysch succession that unconformably overlies the Saint-Daniel Mélange (fig. 4.2, 4.3; Cousineau and St-Julien, 1994; Schroetter et al., 2006). Volcaniclastic rocks are common toward the base, in the Frontière, Etchemin and Beauceville formations, but more than 70% of the group is composed of turbiditic slate and sandstone attributed to the Saint-Victor Formation (Cousineau and St-Julien, 1994; St-Julien, 1987).

The Ascot Complex represents the remnants of a peri-Laurentian volcanic arc (fig. 4.2; Tremblay, Hébert and Bergeron, 1989; Tremblay, 1992a). It is composed of felsic and mafic volcanic rocks and syn-volcanic intrusives that are overlain by, and in fault contact with, pebbly phyllites that correlate with those of the Saint-Daniel Mélange (Tremblay, Hébert and Bergeron, 1989; Gauthier et al., 1989; Tremblay and St-Julien, 1990). Felsic volcanic rocks and granite belonging to the Ascot Complex have yielded U-Pb zircon ages of 460 ± 3 Ma and 441 +7/-12 Ma (David and Marquis, 1994), and an ^{40}Ar-^{39}Ar muscovite age of 462 ± 2 Ma (Tremblay, Ruffet and Castonguay, 2000). The abundance of inherited Precambrian zircons, and the trace element and isotopic geochemical signature of the Ascot Complex felsic rocks indicates a strong contribution of Laurentian continental crust and/or sediments in its genesis (Tremblay, Hébert and Bergeron, 1989; Tremblay et al., 1994; David and Marquis, 1994).

4.1.1.2 The northern Vermont Rowe-Hawley belt

Dunnage and Humber zone rocks of southern Quebec can be mapped continuously into the northern Vermont Rowe-Hawley belt (figs. 4.1, 4.2; Doolan et al., 1982). The main stratigraphic correlations across the Quebec – Vermont international border are summarized and illustrated in the geologic map of figure 4.2. Well-preserved ophiolites are not preserved in northern Vermont, but amphibolites, serpentinites, felsic gneisses and granitoid rocks discontinuously exposed from central Vermont to Connecticut are believed to represent dismembered Early Ordovician forearc-arc-backarc remnants (e.g. the Shelburne Falls arc; Stanley and Ratcliffe, 1985; Kim and Jacobi, 1996, 2002; Karabinos et al., 1998). The Saint-Daniel Mélange is mapped into Vermont as the Cram Hill Formation (Doolan et al., 1982 and references therein). The contact between the Cram Hill Formation and the underlying Stowe Formation, which correlates with the Caldwell Group of the southern Quebec Humber zone, is marked by the Umbrella Hill Formation conglomerate and has been interpreted either as an unconformity (Doolan et al., 1982; De Souza and Tremblay, 2010b) or a fault (Stanley and Ratcliffe, 1985; Kim et al., 1999). Felsic volcanic rocks and dykes occurring in the Cram Hill Formation were tentatively correlated with part of the Ascot Complex (Doolan et al., 1982; Stanley and Ratcliffe, 1985). The Moretown Formation of Vermont does not have any known counterpart in southern Quebec. It consists of quartz arenites, feldspathic sandstone and phyllites (Cady et al., 1963; Doolan et al., 1982; Kim et al., 1999) and has been recently divided into western and eastern informal members (Kim et al., 1999), that either represent part of the Humber or Dunnage zone, or both (Rowley and Kidd, 1981; Doolan et al., 1982; Stanley and Ratcliffe, 1985; Kim et al., 1999; Kim, 2006; De Souza and Tremblay, 2010b). The Bolton Igneous Group has been correlated with the Coburn Hill Volcanics and Mount Norris Intrusive Suite (Kim et al., 2003).

4.1.1.3 The Gaspé Belt

The Gaspé Belt in southern Quebec is made up of two stratigraphic assemblages of Silurian to Early Devonian age (Lavoie and Asselin, 2004). The first assemblage is distributed in a series of synformal outliers that unconformably overlie Dunnage and Humber zone rocks to the northwest of the La Guadeloupe fault (fig. 4.1). It consists of siliciclastic rocks and limestone units that belong to the Glenbrooke Group and the Lac Aylmer, Cranbourne and St-Luc formations. The second assemblage underlies the Connecticut Valley – Gaspé trough and is represented by the Saint-Francis Group and Frontenac Formation (fig. 4.1). The Saint-Francis Group is dominated by siliciclastic rocks that are divided into the Lac Lambton, Ayers Cliff and Compton formations, and can be laterally correlated, at least in part, with the volcano-sedimentary package of the Frontenac Formation (fig. 4.1; Lavoie and Asselin, 2004). Although the thickness of the Saint-Francis Group remains unknown, seismic reflection data indicate that it forms a ca. 3 km thick seismostratigraphic unit (Spencer et al., 1989). In New England, the Ayers Cliff and Compton formations are mapped into and correlate with the Vermont Waits River and Gile Mountain formations, respectively (Hueber et al., 1990; McWilliams, Walsh and Wintsch, 2010). The maximum age of the Waits River Formation is constrained to Wenlockian time based on U-Pb dating of a crosscutting felsic dyke (423 ± 3 Ma; Aleinikoff and Karabinos, 1990; Hueber et al., 1990), whereas the upper age limit of the Gile Mountain Formation is believed to be Pragian to Emsian, as suggested by U-Pb dating of detrital zircons and felsic volcanic rocks (Rankin and Tucker, 2009; McWilliams, Walsh and Wintsch, 2010). Numerous tectonic settings have been proposed for the formation of the Connecticut Valley-Gaspé trough (see Tremblay and Pinet

(2005) and Rankin et al. (2007) for a review). Nevertheless, there seems to be a growing consensus on relating the trough to Silurian-Devonian delamination-related extension and basin formation (van Staal and de Roo, 1995; Robinson et al., 1998; Tremblay and Pinet, 2005; Rankin et al., 2007).

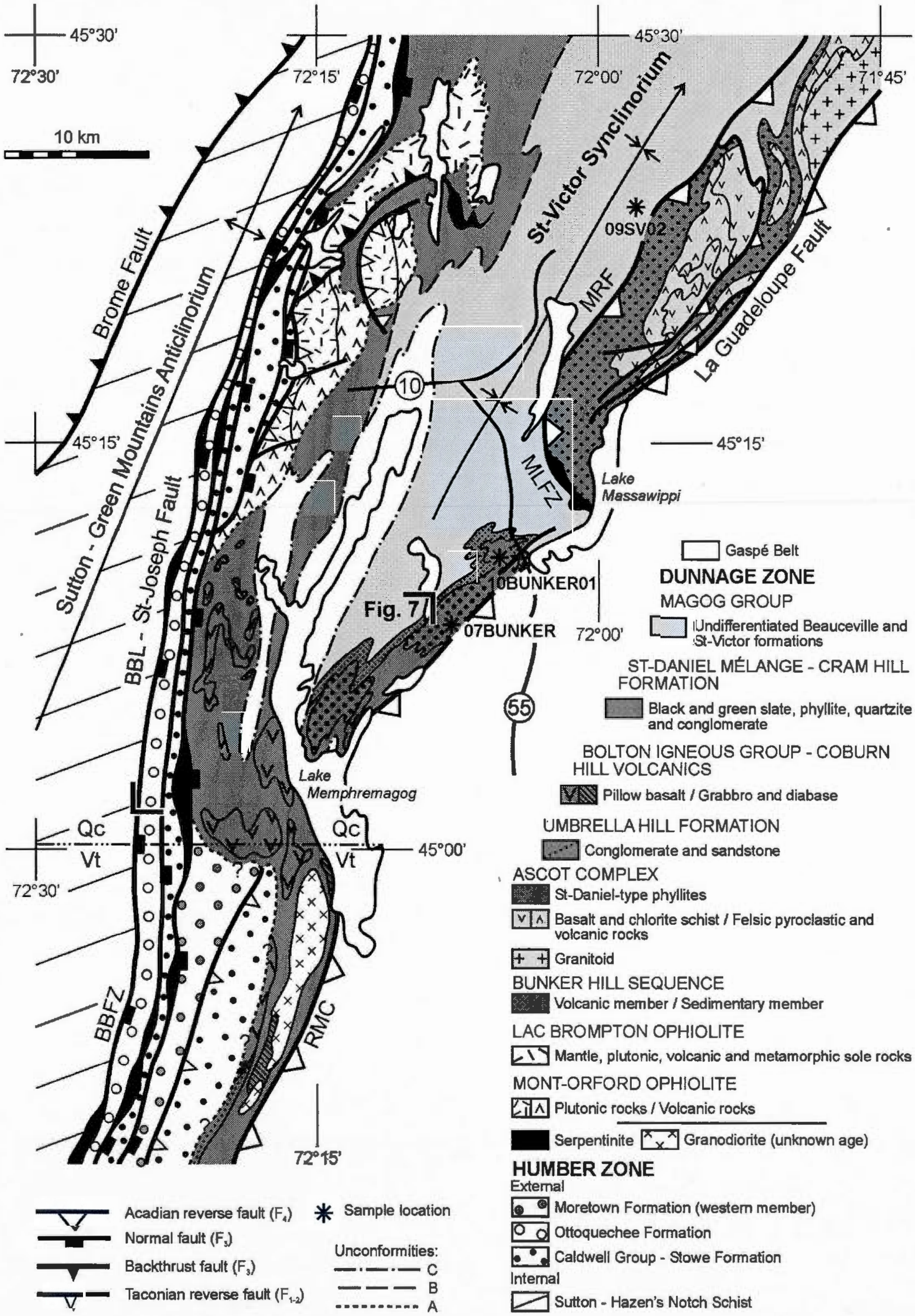


Figure 4.2 : Generalized geological map of the Lake Memphremagog area (after Gale, 1980; Lamothe, 1979, 1981a, 1981b; De Romer, 1980; Doolan et al., 1982; Slivitzky and St-Julien, 1987; Rickard, 1991; Tremblay, 1990, 1992b; Huot, 1997; Kim et al., 1999; Castonguay and Tremblay, 2003; De Souza et al., 2008). BBFZ – Burgess Branch fault zone; MLFZ – Massawippi Lake fault zone; MRF – Magog River fault; other symbols as in figure 4.1.

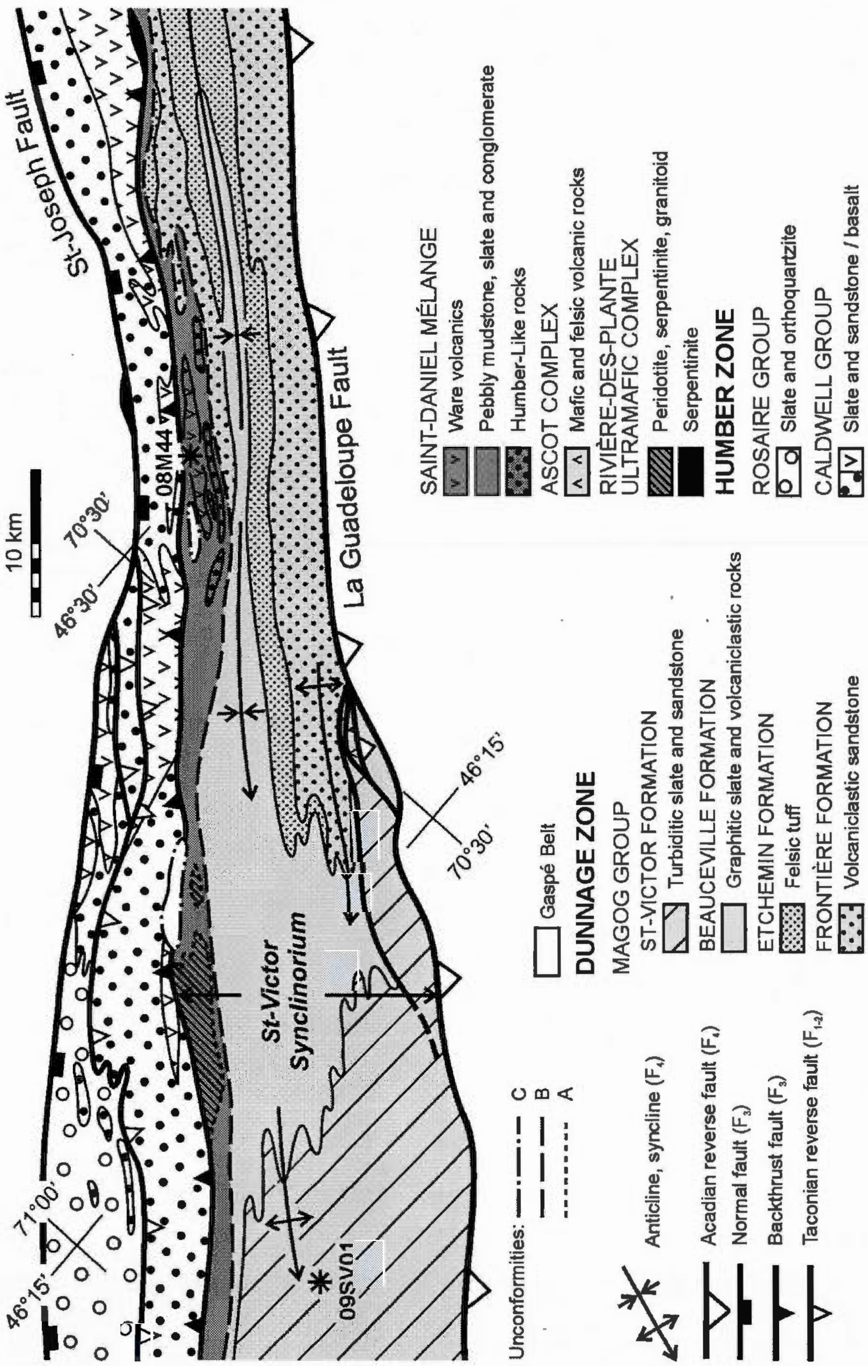


Figure 4.3 : Geological map of the Beauce area (after Cousineau, 1986; 1990; St-Julien, 1987; De Souza and Tremblay, 2010a). Asterisks indicate sample locations.

4.1.1.4 Tectonic synthesis

The southern Quebec Dunnage and Humber zones and the northern Vermont Rowe-Hawley belt have undergone a threefold tectonic evolution, with evidence of deformation and metamorphism during the Ordovician, the Silurian to Early Devonian and the Middle Devonian (Tremblay and Castonguay, 2002; Castonguay et al., 2012). The oldest structural fabric is a S_{1-2} composite foliation that is related to ophiolite emplacement and northwest-directed thrusting of the Humber zone allochthonous nappes during the taconic orogeny, and has been isotopically dated at 471 to 450 Ma (Tremblay and Pinet, 1994; Castonguay et al., 2001, 2012; Castonguay, Ruffet and Tremblay, 2007; Sasseville et al., 2008; Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012). Hinterland-directed (i.e. southeast) deformation attributed to post-Taconian backthrusting has resulted in the formation of a S_3 foliation and northwest-dipping thrust faults (e.g. the Brome – Bennett fault; fig. 4.1), and has culminated with normal faulting along the Saint-Joseph – Baie Verte-Brompton line and Burgess Branch fault zones in southern Quebec and northern Vermont, respectively (Tremblay and Castonguay, 2002; Castonguay and Tremblay, 2003; Schroetter, Tremblay and Bédard, 2005; Castonguay et al., 2012). This latter episode of regional deformation has been attributed a Silurian-Early Devonian age (ca. 435 to 405 Ma) based on $^{40}Ar/^{39}Ar$ dating of Humber zone rocks (Castonguay et al., 2001, 2012; Castonguay, Ruffet and Tremblay, 2007; Sasseville et al., 2008). Finally, Middle Devonian regional metamorphism, between ca. 390-360 Ma, has been attributed to the Acadian orogeny and is related to the formation of a penetrative S_4 foliation that is overprinted, in the Lake Memphremagog area, by a northwest-dipping crenulation cleavage (Laird, Lanphere and Albee, 1984; Tremblay, Ruffet and Castonguay, 2000; Castonguay et al., 2001, 2012; Castonguay, Ruffet and Tremblay, 2007). S_4 corresponds to the earliest structural fabric in the Gaspé Belt and correlative units of northern Vermont (Cousineau and Tremblay, 1993; Tremblay and Pinet, 1994; Castonguay et al., 2012).

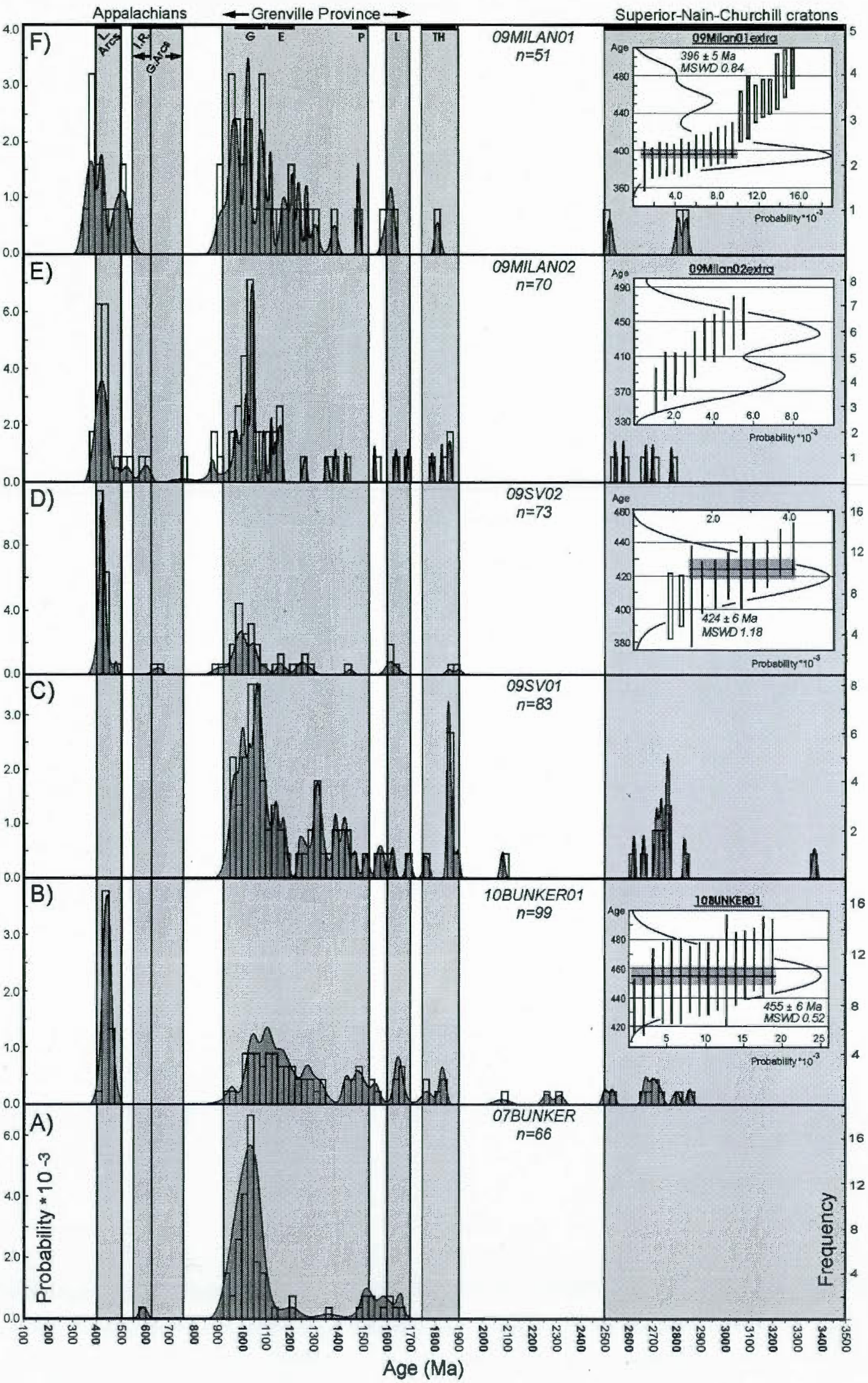
4.2 Geochronologic and stratigraphic data

Field work for this study was mostly conducted in the Beauce and Lake Memphremagog areas, which correspond, respectively, to the northeastern and southwestern periclinal terminations of the Saint-Victor synclinorium (fig. 4.1). Both areas are underlain by Humber and Dunnage zone rocks; the main structural and stratigraphic features that have been recognized in both areas are illustrated in figures 4.2 and 4.3. Rocks of the Beauce area have undergone low-grade metamorphism during the Acadian orogeny, whereas those of the Lake Memphremagog area attain a greenschist metamorphic grade.

Rock units that were studied and sampled for detrital zircon U-Pb and/or muscovite $^{40}Ar/^{39}Ar$ dating are the Bolton Igneous Group, the Bunker Hill sequence and the Ware Volcanics, as well as the Saint-Victor and Compton formations. The stratigraphic setting of these units and of the various sampling sites is presented in the following sections with the corresponding isotopic age results.

4.2.1 Analytic procedures

Zircon mineral concentrates were prepared from samples of crushed rock by using standard density and magnetic separation techniques. Zircons were extracted from concentrates by handpicking under a binocular microscope, mounted in epoxy and polished for analysis by LA-MC-ICP-MS (laser ablation-magnetic sector-inductively coupled plasma-mass spectrometry). The analyses were conducted at the University of Alberta, Edmonton, Alberta, by using the analytic protocol presented by Simonetti et al. (2005). The analyses that are less than 30% discordant are shown in Appendix F and illustrated as probability density distribution diagrams in figure 4.4. Unless mentioned, the $^{206}Pb/^{238}U$ ages are used for zircons younger than ca. 1000 Ma, whereas the $^{207}Pb/^{206}Pb$ ages are used for zircons older than ca. 1000 Ma. However, although less precise for zircons younger than ca. 1000 Ma, $^{207}Pb/^{206}Pb$ ages are generally more accurate and less sensitive to lead loss and instrumental fractionation relative to the $^{206}Pb/^{238}U$ ratio (Gehrels, 2011), and are preferably considered for determining youngest zircons age clusters whenever possible. Age clusters used for age determinations and the calculation of weighted mean ages are defined as a series of at least six $^{207}Pb/^{206}Pb$ or $^{206}Pb/^{238}U$ ages that overlap with 2σ error bars and have a 2σ uncertainty of less that 10%. Error bars of 1σ are used in the probability density distribution diagrams, whereas all ages are reported in text at the 2σ level. Cathodoluminescence imaging was conducted at the Microscopy and Microanalysis Facility, University of New Brunswick, Fredericton, New Brunswick, the results of which are summarized in fig. 4.5.



Appalachians
Grenville Province
Superior-Nain-Churchill cratons
L Arcs
I.R.
G-Arcs
G
E
P
L
TH
F)
09MILAN01
n=51
09Milan01extra
396 ± 5 Ma
MSWD 0.84
E)
09MILAN02
n=70
09Milan02extra
D)
09SV02
n=73
424 ± 6 Ma
MSWD 1.18
C)
09SV01
n=83
B)
10BUNKER01
n=99
10BUNKER01
455 ± 6 Ma
MSWD 0.52
A)
07BUNKER
n=66
Age
Probability*10-3
Probability *10 -3
Frequency
Age (Ma)

Figure 4.4 : Probability density distribution – frequency diagrams of detrital zircon ages for samples a) 07BUNKER, b) 10BUNKER01, c) 09SV01, d) 09SV02, e) 09MILAN02, f) 09MILAN01. The age – probability density distribution plots shown in the insets of b), d) and f) are for the youngest age clusters from which were calculated weighted mean ages; ages illustrated by empty boxes were excluded from mean calculations. Insets show ages that are ca. 90% to 105% concordant. Note that in inset of f), the ages do not form an age cluster (see text for explanation). All the probability density distribution plots were generated using AgeDisplay (Sircombe, 2004), whereas weighted mean ages were calculated with Isoplot (Ludwig, 2008). Light grey shaded areas in the background highlight main episodes of continental crust formation and magmatism in the northeastern Canadian Shield (St-Onges, Scott and Wodicka, 2002; Wardle and Hall, 2002; Gower and Krogh, 2002; Clark and Wares, 2004; Tollo et al., 2004; Ansdell et al., 2005; Corrigan, Galley and Pehrsson, 2007; Percival, 2007) and Appalachians (Cawood, McCausland and Dunning, 2001; Hatcher, 2010; Hibbard, van Staal and Rankin, 2010). E – Elzevirian Orogeny; G – Grenvillian Orogeny; G.Arcs – peri-Gondwanan arcs; IR Iapetan rift magmatism; L.Arcs – peri-Laurentian arcs; P – Pinwarian Orogeny; L – Labradorian Orogeny; TH – Trans-Hudson, New Quebec, Torngat and Makkovik orogens. n – represents the number of analyses presented in the corresponding diagram.

Figure 4.5 : Cathodoluminescences images of dated zircons from the a) Bunker Hill sequence volcanic member – 10BUNKER01; b) Ware Volcanics – 08M44; c) Saint-Victor Formation – 09SV02; d) extra grain sets of the Milan member, Compton Formation – 09MILAN01extra and e) – 09MILAN02extra. Dashed lines highlight laser ablation pits and dotted lines delimit inherited zircon cores. The numbers in italic correspond to the age and uncertainty of the dated grains, whereas the numbers in parentheses correspond to the analytical results presented in Appendix F. Scale bars are all 50 μm and zircons from a), b) and e) are all at the same scale.

For the $^{40}Ar/^{39}Ar$ analyses, muscovite flakes were handpicked from the same samples of crushed rock that were processed for the preparation of zircon concentrates. Single muscovite grains were then analyzed by laser step-heating at the CNRS-Université de Rennes 1, Rennes, France, following the analytical procedures that are given by Ruffet, Féraud and Amouric (1991), Ruffet et al. (1995), Castonguay et al. (2001) and Castonguay, Ruffet and Tremblay (2007). All of the muscovite ages are reported using 1σ error bars in fig. 4.6.

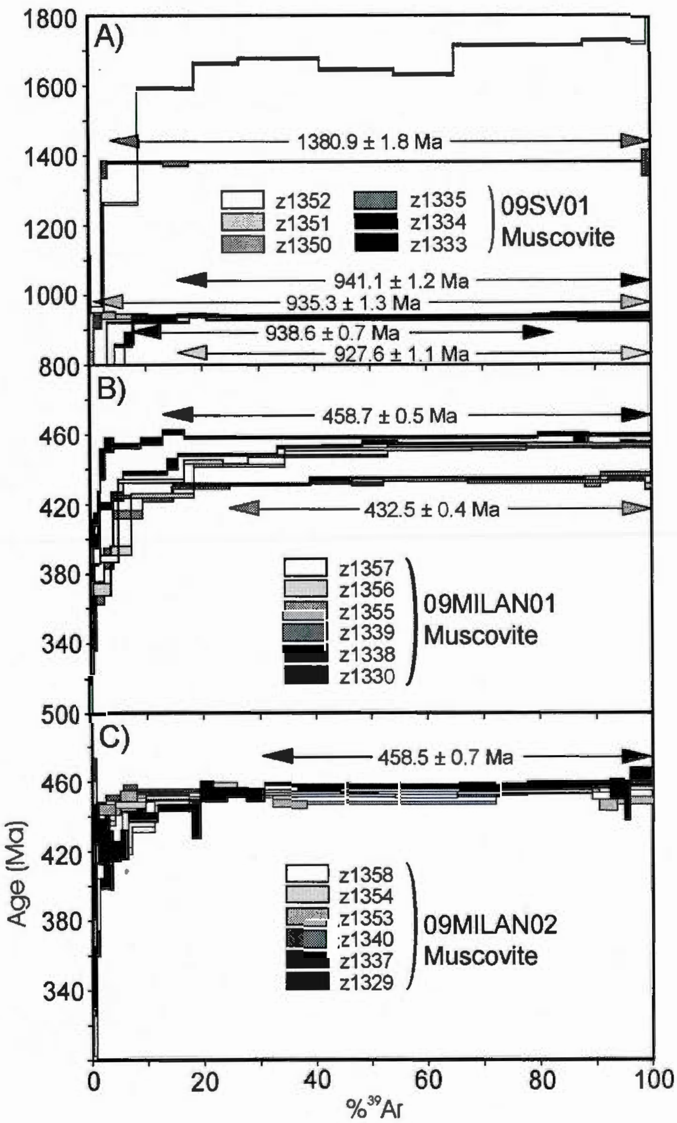


Figure 4.6 : $^{40}Ar/^{39}Ar$ age spectra for detrital muscovites from a) the St-Victor Formation – 09SV01, and the b) Milan member – 09MILAN01 c) – 09MILAN02.

4.2.2 The Bolton Igneous Group

The Bolton Igneous Group includes all mafic volcanic and intrusive rocks found in the Saint-Daniel Mélange west of Lake Memphremagog (fig. 4.7; Cooke, 1950; Ambrose, 1957; Lamothe, 1981a; Rickard, 1991). Basalt, gabbro and mafic schists have also been mapped as part of the Saint-Daniel Mélange to the east of Lake Memphremagog, in the Fitch Bay area, and probably correlate with those of the Bolton Igneous Group (fig. 4.7; De Romer, 1980). Gabbroic rocks have been observed in the Bunker Hill sequence sedimentary member and were also included with the Bolton Igneous Group in our map of fig. 4.7.

Complete petrographic descriptions and geochemical characterization of the Bolton Igneous Group are provided by Mélançon et al. (1997). It consists of well-preserved pillowed basaltic-to-andesitic volcanic rocks (fig. 4.8a), gabbro and diabase. Intermediate to felsic dykes also occur locally. The volcanic rocks form isolated to folded and almost continuous massifs of variable apparent thickness (0.3 to 2.0 km) that extend for distances of up to 20 km with consistent younging directions, whereas the intrusive facies form meter- to decametre-scale or larger mappable dykes and/or sills that crosscut the Saint-Daniel Mélange slates. Both the volcanic and intrusive facies show a greenschist facies mineral assemblage, with a metamorphic and deformation gradient increasing from north to south. The lower and upper contacts of the volcanic rocks with surrounding slates are exposed in many localities. The lower contact is often marked by the occurrence of diabase and gabbro dykes and sills, some of which can be mapped into the overlying volcanic rocks. Both the lower and upper contacts are locally marked by ca. 100-150 m thick greenish to purple slate and chert that contain manganese oxide laminations and nodules (figs. 4.7, 4.8b), and by the local occurrence of polymetallic sulphide mineralization (Trottier, Brown and Gauthier, 1991). Evidence for faulted contacts is lacking, indicating, as proposed by Doolan et al. (1982), Rickard (1991) and Trottier, Brown and Gauthier (1991), that the Bolton Igneous Group basalts form conformable horizons interlayered within the Saint-Daniel Mélange. However, in the vicinity of the Quebec – Vermont border, contacts are highly strained due to a southward increase in the intensity of deformation and strain localization along slate-basalt contacts. Our work and previous mapping by Lamothe (1981a) and Rickard (1991) clearly shows that the Bolton basalts, as well as the Saint-Daniel Mélange slates, were folded prior to the deposition of the overlying Glenbrooke Group. We correlate this episode of folding with the F_3 phase of Tremblay and Pinet (1994), to which is also probably related the presence of overturned bedding in the volcanic rocks (fig. 4.7). Two samples of quartz-bearing dykes of intermediate composition were processed for U-Pb zircon dating; these have however exclusively yielded inherited zircon grains that were not analyzed.

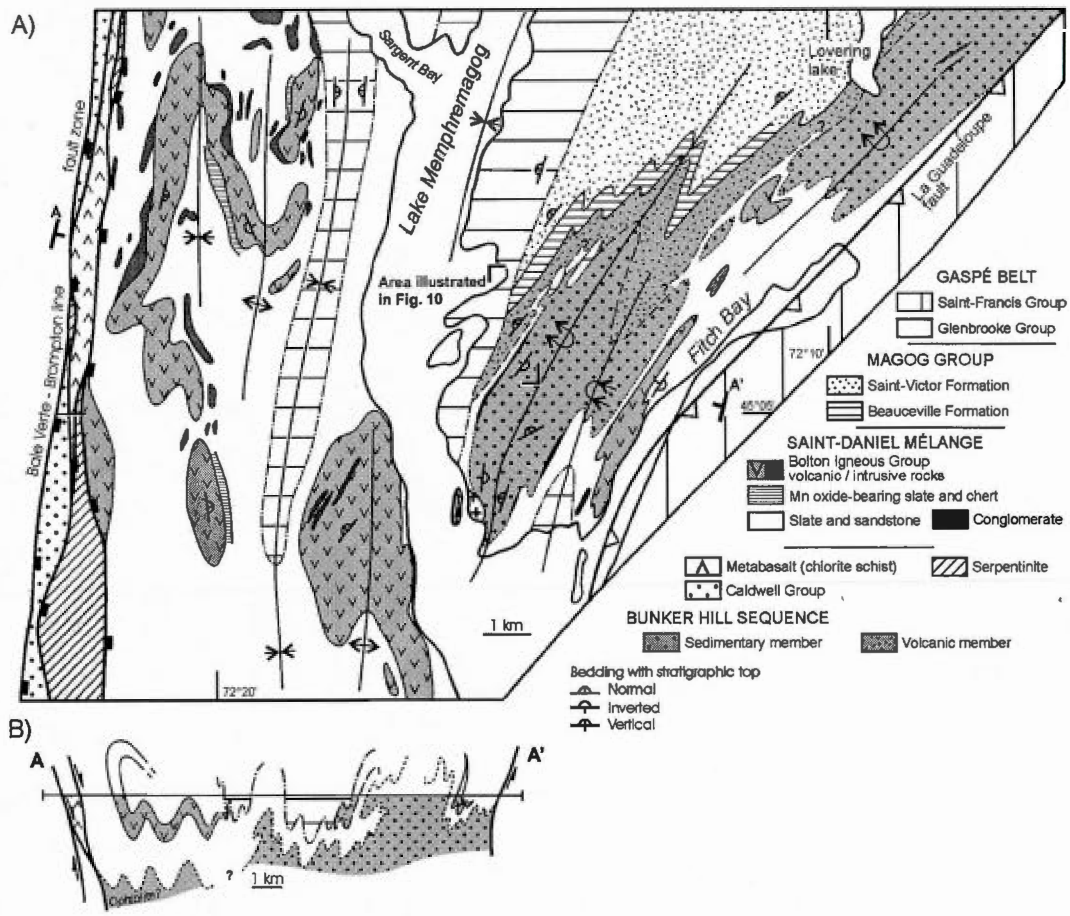


Figure 4.7 : a) Detailed geological map and b) cross-section of the Lake Memphremagog – Fitch Bay area showing the main stratigraphic relationships between the Saint-Daniel Mélange, Bolton Igneous Group, Bunker Hill sequence and Magog Group. No significant vertical exaggeration in section. Same legend for a) and b) and the unconformities are illustrated as in fig. 4.2. Modified from De Romer (1980), Tremblay (1990) and De Souza (2012).

4.2.3 The Bunker Hill sequence

The Bunker Hill sequence is a volcano-sedimentary rock assemblage of unknown age that occurs between the Memphremagog and Massawippi lakes. It is divided into two informal members: a volcanic member consisting of pyroclastic and volcaniclastic rocks and a Humber-like sandstone-dominated sedimentary member (fig. 4.2; De Romer 1980; Tremblay, 1990; Blais, 1991). Both members are exposed along prominent linear hills between Fitch Bay and Massawippi Lake (figs. 4.2 and 4.7). The sedimentary member consists of a monotonous succession of interstratified green to grey and poorly-sorted feldspathic greywacke with discontinuous lenses of quartz-pebble conglomerate, siltstone and mudslate. The major constituents of the greywacke are angular to sub-rounded grains of quartz, plagioclase, granitoid, k-feldspar and quartzite. Zircon, opaque minerals and bluish quartz are also abundant. The greywackes show strong petrographic similarities with those of the Caldwell Group (De Romer, 1980; Slivitsky and St-Julien, 1987; Tremblay, 1990), which is a known passive margin unit derived from the erosion of Laurentian crystalline basement (Tawadros, 1977; Cousineau, 1990). Tremblay (1992a) implicitly correlated both units. Younging directions measured in the sedimentary member are opposite to those found in the overlying Magog Group, which suggests, according to Tremblay (1990), the existence of a major discontinuity between both of these units.

The volcanic member occurs to the northwest of the sedimentary member and is best exposed in a quarry near Massawippi Lake (fig. 4.9). It is made up of cream- to light brown-weathering and aphanitic felsic tuff and chert interlayered with volcanic conglomerate, volcaniclastic sandstone and black phyllite. The felsic tuff is light grey to green, massive to laminated (fig. 4.8c) and shows a saccharoidal recystallization texture. A unique characteristic of the volcanic conglomerate and volcaniclastic sandstone is that they both contain bright green-colored clasts that are generally less than 5 mm in size and consist of fuchsite, chlorite and chromite with traces of serpentine. The volcanic conglomerate is however mostly made up of tuffaceous fragments identical to those of the surrounding tuffs, with minor amounts of granitic, volcaniclastic, dacitic? and pelitic fragments embedded in a fine-grained felsic matrix (fig. 4.8d). The conglomeratic horizons are generally less than 10 m-thick, are devoid of any internal stratification and locally show graded bedding over several meters. Variably carbonatized and brown-weathering gabbroic sills and/or dykes of transitional-alkaline geochemical affinity were mapped as part of the Bunker Hill sequence volcanic member and have been interpreted as a key feature for the tectonic interpretation of the sequence (Blais, 1991; Tremblay, 1990, 1992a). Similar dykes were however observed to cut the Saint-Daniel Mélange slates during our study, which makes their age and interpretation uncertain.

The relationships between the Bunker Hill sequence sedimentary and volcanic members, Saint-Daniel Mélange, Magog Group and Gaspé Belt were investigated on the eastern shore of Lake Memphremagog (fig. 4.10). There, sandstone beds of the Bunker Hill sequence form an inverted southeast-facing stratigraphic succession that is successively overlain by: 1) polymictic conglomerate (ca. 10 m) and slate included in the Saint-Daniel Mélange, 2) tuffaceous and volcaniclastic rocks belonging to the volcanic member of the Bunker Hill sequence, 3) graphitic pyritiferous black slate that correlate with the Beauceville Formation, 4) a turbiditic succession of well-bedded siltstone and slate belonging to the Saint-Victor Formation, and 5) the unconformity marking the base of the Glenbrooke Group. Graded bedding observed in sandstone of the Saint-Daniel Mélange, volcaniclastic rocks of the Bunker Hill sequence and the Saint-Victor Formation turbidites all suggest that they form part of a northwest-facing stratigraphic succession. Along-strike, the sedimentary and volcanic members of the Bunker Hill sequence are locally shown to be adjacent to one another, without the presence of Saint-Daniel Mélange slate (fig. 4.7). This may be attributed to poor outcrop quality in those areas and/or to the occurrence of a discontinuity at the base of the volcanic member. The conglomerate that is found toward the base of the Saint-Daniel Mélange comprises fragments of green to grey feldspathic greywacke, felsic tuff, green phyllite and siltstone that measure up to 50 cm and are supported by a black slaty matix. The sandstone fragments are petrographically identical to the underlying Bunker Hill greywackes. Similar conglomerates have also been mapped ca. 8 km to the northeast of the Lake Memphremagog area, along Highway 55 (fig. 4.2). At this latter locality, some of the greywacke fragments clearly show a penetrative pre-depositional foliation (figs. 4.8e, 4.8f). This, as well as conflicting younging directions between the Bunker Hill sequence sedimentary member and overlying rocks, suggests that the sedimentary member was deformed and overturned prior to the sedimentation of the Saint-Daniel Mélange. It is therefore inferred that the contact between the Bunker Hill sequence sedimentary member and the Saint-Daniel Mélange is probably an unconformity, and that both members of the Bunker Hill sequence should be considered as distinct stratigraphic units that have undergone different geologic evolutions.

U-Pb data

Two samples consisting of a coarse-grained feldspathic greywacke (07BUNKER) and a volcanic conglomerate (10BUNKER01) were selected from the sedimentary and volcanic members of the Bunker Hill sequence for U-Pb zircon geochronology. The sampling sites are located in figs. 4.1, 4.2 and 4.9. Sample 07BUNKER yielded abundant zircon grains showing a variety of morphologies that suggest igneous and metamorphic sources. A random selection from this detrital grain set was analyzed and has yielded six ages. The probability density distribution diagram for this dataset is rather simple, with a prominent peak at 1100-925 Ma that represents over 77% of the analyzed grains, and a more discrete one at 1675-1475 Ma (fig 4.4a). Most of the remaining analyses are distributed between these two age groups, but a single Ediacaran age of 594 ± 28 Ma was also obtained.

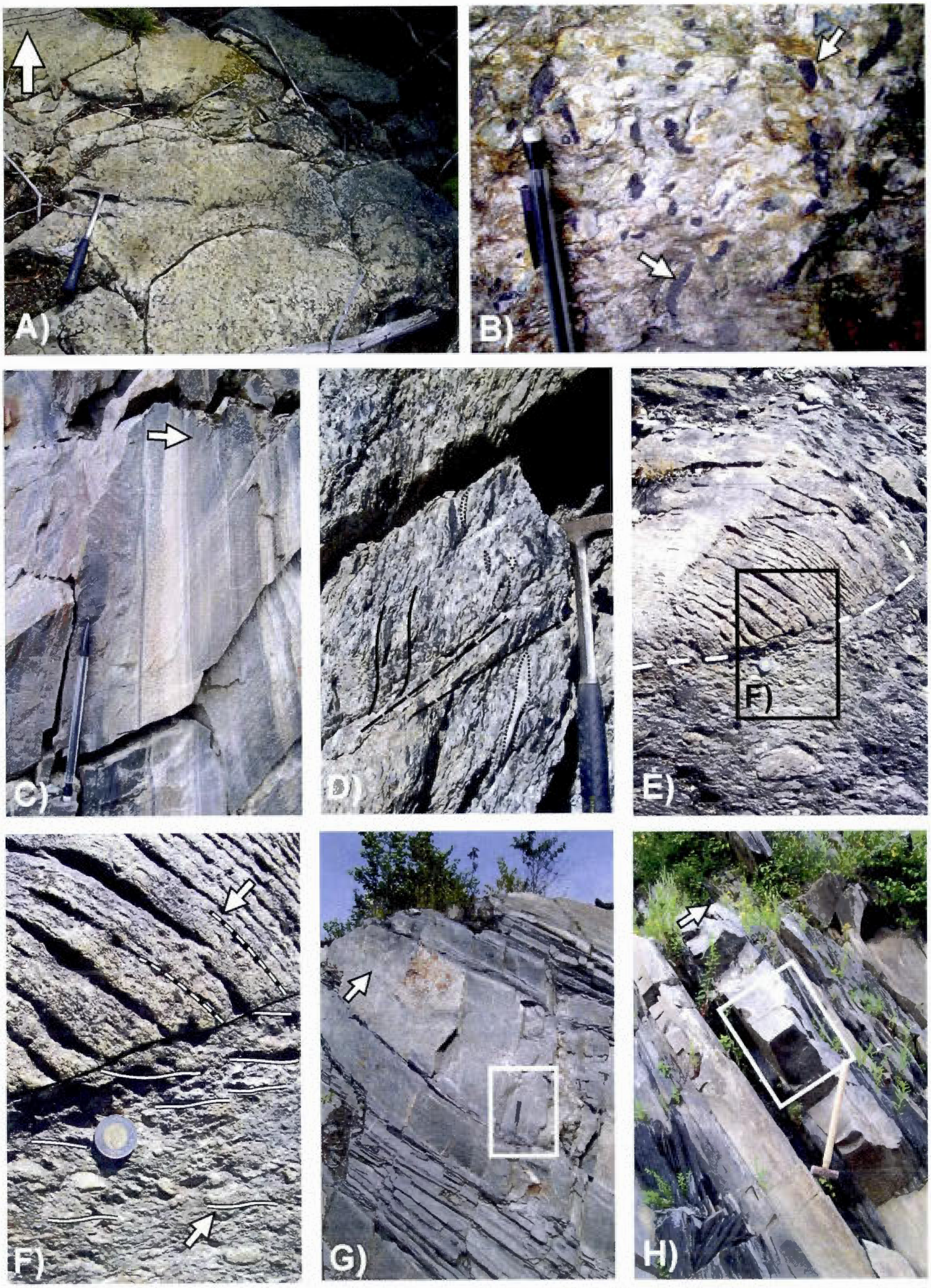


Figure 4.8 : Field photograph of a) well-preserved pillowed volcanic rocks of the Bolton Igneous Group located ca. 2 km to the west of Sargent Bay on Lake Memphremagog (fig. 4.7) indicating an eastward younging direction (white arrow). The photo is in plane view and was taken looking toward the east; b) manganese oxide nodules (white arrow) in a cherty horizon marking the contact between the Bolton Igneous Group basalts and Saint-Daniel Mélange slates; c)

bedded tuff of the Bunker Hill sequence volcanic member with an horizon of graded volcaniclastic sandstone (pinkish brown layer in center). Younging direction is toward the southeast at this locality (white arrow); d) volcanic conglomerate belonging to the Bunker Hill sequence volcanic member and that was sampled for U-Pb zircon dating (sample 10BUNKER01). Clasts are highlighted by dotted lines, whereas the full and dashed lines correspond to the main foliation (S_4) and a superposed crenulation cleavage, respectively; e) polymictic conglomerate belonging to the Saint-Daniel Mélange of the Fitch Bay area. The clast outlined by the white dashed line is composed of foliated feldspathic sandstone similar to the one of the Bunker Hill sequence sedimentary member; f) detail of area indicated in e) and showing the sharp contact between the foliated clast and the pebbly matrix. Note that the foliation in the clast (dashed white line) is at high angle to the main foliation (F_4) in the matrix (full white line); g) Saint-Victor Formation turbidites that were sampled for U-Pb dating of detrital zircons (sample 09SV02); the arrow indicates the stratigraphic top (toward the northwest), whereas the white box shows the area of the outcrop that was sampled; h) Compton Formation (Milan member) black slate and brown lithic sandstone; the stratigraphic top is toward the southeast (white arrow) and the white box indicates the area that was sampled for isotopic dating (sample 09MILAN02).

A wide variety of zircon grains were separated from sample 10BUNKER01. Many show anhedral, amoeboid to discoid or more-or-less spherical shapes and are most probably inherited from metamorphic sources, whereas others are euhedral with sharp to smooth edges and bipyramidal prismatic and equant to elongated crystal shapes that suggest an igneous origin. The relative probability plot for this sample shows a very broad Precambrian age distribution with a dominant composite peak in the 925-1250 Ma age interval that represents 34% of the analyzed grains (fig. 4.4b). It is however characterized by a very distinct Early Paleozoic age population represented by 24 single grain analyses (fig. 4.4b). These younger ages were all obtained from euhedral and elongated to prismatic biterminated zircon crystals and crystal tips that show well-defined sector and oscillatory zoning, without any evidence of inherited cores (fig. 4.5a). Sixteen of these analyses are concordant at >93% and form an age cluster from which was calculated a weighted mean $^{207}Pb/^{206}Pb$ age of 455 ± 6 Ma (inset of fig. 4.4b).

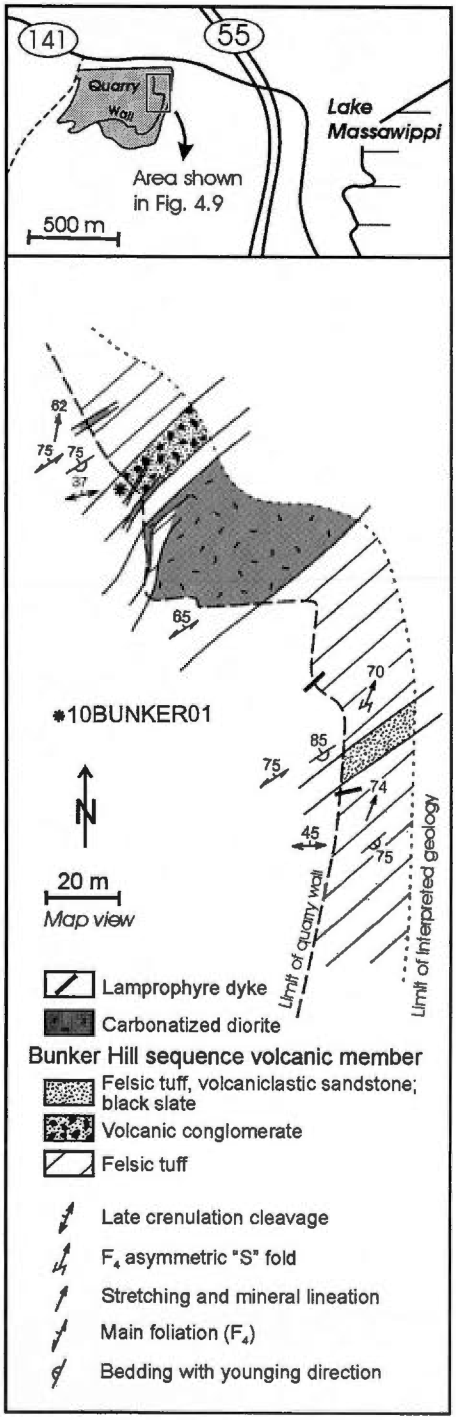


Figure 4.9 : Detailed sketch map of a section of the quarry where the Bunker Hill sequence volcanic member sample 10BUNKER01 was collected. The inset shows the location of the quarry.

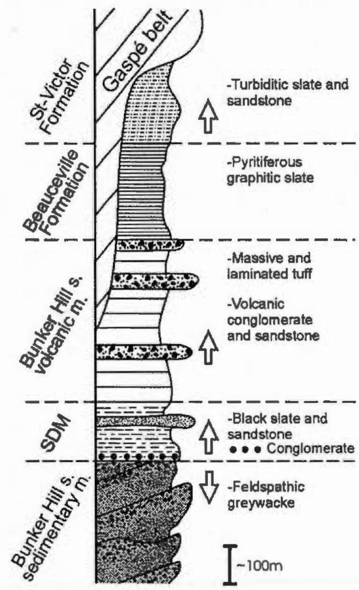


Figure 4.10 : Schematic stratigraphic column of the contact zone between the Bunker Hill sequence sedimentary and volcanic members, the Saint-Daniel Mélange, the Magog Group and the Gaspé Belt on the east shore of Lake Memphremagog. White arrows indicate younging directions determined from graded bedding. See fig. 4.7 for the area represented.

4.2.4 The Ware Volcanics

The Ware Volcanics (Cousineau, 1990) are made up of dacitic to rhyolitic volcanic rocks occurring as lens-shaped bodies within green and black slates and siltstones of the Saint-Daniel Mélange. These volcanic rocks are white to tan weathering, aphyric to feldspar-, amphibole- and quartz-phyric, locally show a vesicular texture, flow-banding and well-developed perlitic fractures up to 1cm-wide. Brecciated facies are abundant. The preservation of jigsaw fit textures, flow-banding and fragments showing contorted jagged edges suggest that most of these are autoclastic flow breccias (Fisher, 1960). Pyroclastic facies may also be present (Cousineau, 1990). The volcanics underlie siliciclastic rocks of the Saint-Luc Formation, which have been attributed to the Gaspé Belt (Cousineau, 1990; Lavoie and Asselin, 2004). Where observed, the contact with the Saint-Daniel Mélange slates is sharp and marked by the occurrence of green siliceous mudstone. The lack of fault rocks, shear-related features and no apparent increase in strain at or toward the contact with the Saint-Daniel Mélange suggests that the Ware Volcanics represent conformable volcanic flows within the Saint-Daniel slates. These volcanic rocks are calc-alkaline, peraluminous and show highly-fractionated rare earth element spectra with elevated chondrite-normalized La/Yb and Th/Yb ratios (5-11 and 16-19; see geochemical data in

Appendix E). Such Th and La enrichment may suggest a strong crustal component in the genesis of these volcanic rocks, whereas the lack of voluminous mafic rocks precludes an origin by the differentiation of a basaltic magma.

U-Pb data

A sample (08M44) from the Ware Volcanics that was selected for U-Pb consisted of massive plagioclase- and quartz-phyric rhyolite. Zircon grains that were separated from this sample vary from subrounded frosty and brownish zircon grains, to equant, elongated and doubly-terminated prisms and needle-like acicular crystals. The LA- ICP-MS analyses and cathodoluminescence images revealed that most of these grains are inherited Neo- to Mesoproterozoic xenocrysts (fig. 4.5b) with ages varying between ca. 1400 and 800 Ma (Appendix F). Only two analyses of acicular and oscillatory-zoned zircon fragments have yielded Ordovician-Silurian ages of 438 ± 28 Ma and 440 ± 17 Ma, which makes further age determinations based on this dataset highly speculative.

4.2.5 The Magog Group

Our work on the Magog Group is synthesized in figures 4.2 and 4.3 and is essentially based on reconnaissance mapping in various areas of southern Quebec and the compilation of previous work by St-Julien (1965, 1975, 1987), Slivitsky and St-Julien (1987), Cousineau (1990), Tremblay (1991, 1992b) and Schroetter (2004). Complete petrographic descriptions and structural data for the Magog Group are provided in Cousineau (1990), Cousineau and St-Julien (1994) and Tremblay (1992b). In summary, the lower part of the Magog Group is composed of undifferentiated slates and volcaniclastic rocks attributed to the Beauceville Formation (Slivitzky and St-Julien, 1987), which are further divided into the Frontière, Etchemin and Beauceville formations in the Beauce area (fig. 4.3; Cousineau and St-Julien, 1994). There, the Frontière and Etchemin formations consist of interlayered chromite-bearing volcaniclastic sandstone and mudslate, and felsic tuff, respectively, whereas the Beauceville Formation is dominated by graphitic slate and volcaniclastic rocks (Cousineau, 1990; Cousineau and St-Julien, 1994). The overlying St-Victor Formation is the most extensive unit of the Magog Group and consists of quartz- and lithoclast-rich turbidites (Cousineau and St-Julien, 1994; St-Julien, 1987).

Magog-type rock units, almost identical to those of the Frontière and Etchemin formations, have been mapped as part of the Saint-Daniel Mélange along the northwestern limb of the St-Victor synclinorium (Cousineau, 1990), but are included in the Magog Group in our map of figure 4.3. The unconformity marking the base of the Magog Group in the Lake Memphremagog area (St-Julien and Hubert, 1975; Doolan et al., 1982; Schroetter et al., 2006) was extrapolated to the Beauce area.

U-Pb and $^{40}Ar/^{39}Ar$ data

Two samples of coarse-grained lithic sandstone from the Saint-Victor Formation in the Beauce (09SV01) and Lake Memphremagog areas (09SV02; fig. 4.8g) were selected for U-Pb and $^{40}Ar/^{39}Ar$ detrital zircon and muscovite geochronology, respectively. These sampling sites are located in the lower and upper parts of the Saint-Victor Formation, along the northwestern limb and hinge zone of the Saint-Victor synclinorium, respectively (figs. 4.1, 4.2, 4.3). Sample 09SV01 has yielded abundant zircon grains of various shapes that suggest igneous and metamorphic sources. Most grains are amber to pinkish and reddish-brown colored, show rounded edges and elliptical to discoid and stubby prismatic shapes. Muscovite grains with a diameter of 0.5-1.0 mm were also separated from this sample for ^{40}Ar-^{39}Ar analysis. Only a minor amount of zircon grains was recovered from sample 09SV02, which was however devoid of detrital muscovite. Separated zircons consisted mostly of clear to amber euhedral grains with smooth to sharp edges and equant to elongated and bipyramidal crystal shapes, but brownish and anhedral to well-rounded grains and crystal fragments were also present. The probability density distribution diagram for the zircon analyses of sample 09SV01 is characterized by a lack of Paleozoic age populations (fig. 4.4c). It shows a prominent and broad age group in the 900 to 1400 Ma interval and distinct peaks in the late Paleoproterozoic (1900-1850 Ma) and Neoarchean (2775-2700 Ma). This contrasts with the dataset acquired from sample 09SV02, for which 41% of the dated zircons yielded two Ordovician (480 ± 12 Ma; 459 ± 23 Ma) and 31 Silurian to Early Devonian ages (fig. 4.4d). All of these Paleozoic ages were obtained from euhedral zircons showing well-defined oscillatory and sector zoning, and lacking inherited cores (fig. 4.5c). The remaining analyses from this latter sample form Neo- to Mesoproterozoic age groups in the 1100 to 900 Ma and 1250 to 1150 Ma intervals, with single grain Neoproterozoic ages at 675-625 Ma, and Paleoproterozoic ages as old as 1650-1600 Ma and 1900-1850 Ma. Nine of the analyses yielding Silurian-Early Devonian ages are concordant at ca. 90% to 105%

and form an age cluster from which was calculated a weighted mean $^{238}U/^{206}Pb$ age of 424 ± 6 Ma (inset of fig. 4.4d).

The ^{40}Ar-^{39}Ar analyses that were performed on muscovite grains from sample 09SV01 yielded five well-defined plateau ages, four in the ca. 928-941 Ma age interval and one at 1380.9 ± 1.8 Ma (muscovite Z1335; fig. 4.6a). A highly-disturbed age spectrum was rather obtained for muscovite Z1352, with high- and low-temperature degassing steps suggesting a minimum crystallization age in the Paleoproterozoic (> ca. 1730 Ma) and a strong overprint in the Meso- to Neoproterozoic, respectively (fig. 4.6a).

4.2.6 The Compton Formation

The Compton Formation is the most widely represented unit of the Saint-Francis Group and has been interpreted as a succession of alternating sandstone and mudstone comprising three vertically-stacked informal members: the Milan, Lac Drolet and Saint-Ludger members (Lebel and Tremblay, 1993; Lafrance, 1995; Lavoie, 2004). The Milan member is characterized by the presence of massive, laminar and laminated sandstone with subordinate mudslate interbeds (fig. 4.8h), whereas turbiditic sandstone and mudslate dominate in the Lac Drolet member (Lavoie, 2004). A sedimentological study of the Compton Formation has shown that it formed in a shallow marine deltaic setting that evolved from a delta-front environment in the Milan member to a pro-delta environment in the Lac Drolet member (Lavoie, 2004).

U-Pb and $^{40}Ar/^{39}Ar$ data

Two samples, 09MILAN01 and 09MILAN02, were collected from individual beds of Milan member lithic sandstone (see fig. 4.1 for the location of sampling sites). Zircons that were separated from these two samples are of various shapes and colors, varying from well-rounded to prismatic and amber to reddish-brown, and with most grains generally showing rounded edges. However, a minor amount of grains form a distinct morphological population composed of euhedral and colorless to light amber colored prismatic equant to acicular crystals and crystal fragments with sharp edges. Zircon grains were first randomly-selected from samples 09MILAN01 and 09MILAN02, whereas supplemental grains were carefully handpicked from the euhedral zircon populations, and which form the 09MILAN01b and 09MILAN02b extra grain sets. Cathodoluminescence imaging of these extra grains has revealed that most show well-defined oscillatory and/or sector zoning, whereas others are characterized by the presence of inherited cores that are surrounded by oscillatory-zoned overgrowths (figs. 4.5d, 4.5e). The detrital zircon probability density distribution plots for samples 09MILAN01 and 09MILAN02 are more-or-less similar, with a very broad Precambrian population and Paleozoic ages extending into the Devonian, with ca. 30% and 26 % of the dated grains corresponding, respectively, to the 950-1100 Ma and 550-375 Ma age intervals for both samples (figs. 4.4e, 4.4f). The 09MILAN01b and 09MILAN02b grain sets have also yielded Precambrian to Devonian ages (Appendix F). All of the Paleoproterozoic to Early Neoproterozoic ages for these extra grain sets were however obtained from inherited zircon cores (figs. 4.5d and 4.5e). The youngest ages from the 09MILAN01b grain set form a well-defined age cluster in the Early to Middle Devonian that is composed of 13 overlapping concordant (95% to 101%) analyses yielding a weighted mean $^{207}Pb/^{206}Pb$ age of 396 ± 5 Ma (inset of fig. 4.4f). Only few Devonian ages were obtained from the 09MILAN02b grain set and, unfortunately, do not form an age cluster (inset of fig. 4.4e). As for samples 09MILAN01 and 09MILAN02, both extra grain sets contain varying amounts of Ordovician to Silurian ages, whereas four Ediacaran ages of 596 ± 31, 614 ± 32 Ma and 552 ± 17 Ma, 562 ± 17 Ma were obtained from sample 09MILAN02 and the 09MILAN02b grain set, respectively.

Samples 09MILAN01 and 09MILAN02 have also yielded abundant detrital muscovite flakes that we analyzed for $^{40}Ar/^{39}Ar$ dating. Sample 09MILAN01 shows a bimodal age distribution, with four analyses in the Late Ordovician and two in the Silurian. Unfortunately, all of the dated grains show evidence of Ar loss in the low- to medium-temperature steps of the age spectra at ca. 375-380 Ma, and only two of the analyses have yielded well-defined plateau ages at 458.7 ± 0.5 Ma and 432.5 ± 0.4 Ma (fig. 4.6b). The remaining age spectra for the Late Ordovician and Silurian muscovite age groups show high-temperature steps up to ca. 450-455 Ma and 435-437 Ma, which represent minimum age estimates for the crystallization and/or cooling of these grains below the closure temperature of muscovite (fig. 4.6b). For sample 09MILAN02, only one muscovite grain has yielded a non-equivocal plateau age at 458.5 ± 0.7 Ma (fig. 4.6c; muscovite z1329). Although this is unclear for muscovite z1337, the age spectra for the remaining analyses show evidence of thermal resetting in the low-temperature steps at ca. 420-430 Ma rather than during the Devonian. Notwithstanding this overprint, the medium- to high-temperature steps for the latter age spectra correspond to a ca. 450-460 Ma age interval, which is consistent with the plateau age that was obtained for muscovite z1329 (fig. 4.6c).

4.3 Discussion

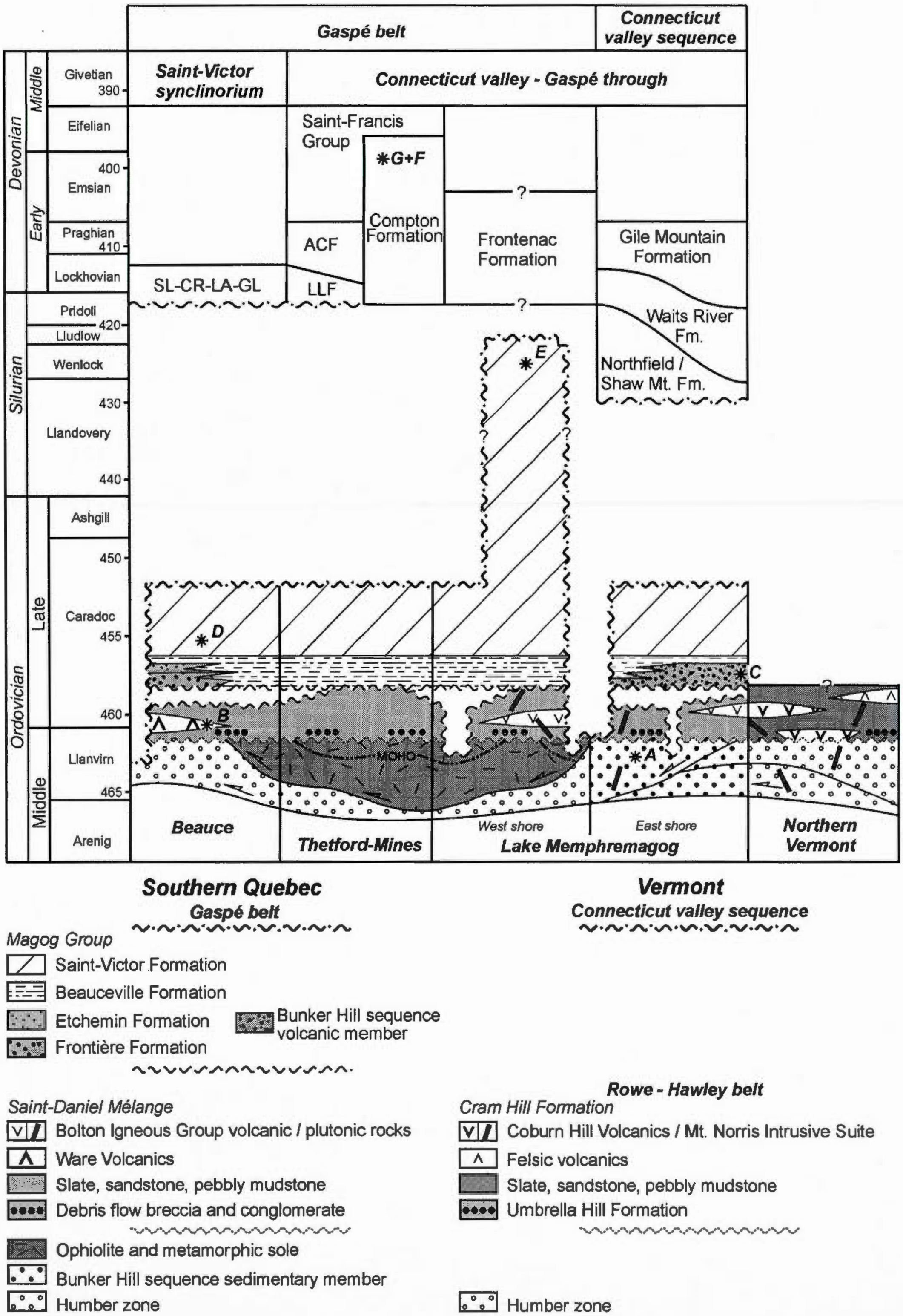
Most of the U-Pb detrital zircon ages that were obtained in our study correspond to major episodes of Appalachian magmatism and Laurentian crust formation and amalgamation, and can be separated into three main age groups: 1) Archean ($\geq$ 2500 Ma), 2) Paleoproterozoic to Neoproterozoic (ca. 1900 to 900 Ma) and 3) Cambrian to Devonian (ca. 500 to 400 Ma). Precambrian age groups are variably represented and prevail in all of the dated samples, clearly indicating significant recycling of continental crystalline basement, whereas the presence of Paleozoic zircons rather depends on the stratigraphic setting and provenance of the dated rock units. The detrital muscovite $^{40}Ar/^{39}Ar$ results are less heterogeneous, but still suggest the erosion of Precambrian (ca. 1730-930 Ma) and Ordovician to Silurian rocks (ca. 460-420 Ma). More specifically, >2.5 Ga high-grade gneiss terranes and granite-greenstone belts are common in the Superior, Nain and Churchill provinces of the Canadian Shield and provide viable sources of Archean zircons (fig. 4.4; Wardle and Hall, 2002; Percival, 2007). Proterozoic ages between ca. 1700-900 Ma and 1900-1750 Ma rather correspond to magmatism and/or high-grade metamorphism in the Grenville Province of eastern North America and main episodes of continental crust formation and amalgamation of the northeastern Canadian Shield, respectively (fig. 4.4; St-Onges, Scott and Wodicka, 2002; Wardle and Hall, 2002; Gower and Krogh, 2002; Clark and Wares, 2004; Tollo et al., 2004; Ansdell et al., 2005; Corrigan, Galley and Pehrsson, 2007). Only few Neoproterozoic ages between ca. 550 and 675 Ma were obtained (figs. 4.4d, 4.4e), yet these overlap with magmatism related to the opening of the Iapetus Ocean in the Northern Appalachians (ca. 620-550 Ma; Cawood, McCausland and Dunning, 2001) and peri-Gondwanan Avalonian-Ganderian arcs (ca. 760-545 Ma; O'Brien et al., 1996). Cambrian-Ordovician igneous rocks are preserved as part of the southern Quebec and western Maine ophiolite belts (ca. 504-470 Ma; David and Marquis, 1994; Kusky, Chow and Bowring, 1997; Moench and Aleinikoff, 2002; Whitehead, Dunning and Spray, 2000; Gerbi et al., 2006), and the Shelburne Falls arc of western New England (ca. 487-471 Ma; Karabinos et al., 1998), whereas Ordovician to Silurian volcanic and plutonic rocks are abundant in the Ascot Complex (ca. 462-441 Ma; David and Marquis, 1994), as well as the Rowe-Hawley and Bronson Hill belts (ca. 470-443 Ma; Lyons, Aleinikoff and Zartman, 1986; Tucker and Robinson, 1990; Moench and Aleinikoff, 2002). Silurian to Early Devonian volcanic and volcaniclastic rocks locally occur in the Connecticut Valley – Gaspé trough, from southern Vermont to the Gaspé Peninsula (for a review see Tremblay and Pinet, 2005). The detrital muscovite $^{40}Ar/^{39}Ar$ ages obtained in this study form Precambrian and Ordovician-Silurian single grain ages and/or populations at ca. >1730 Ma, 1381 Ma, 940-930 Ma and 459-433 Ma, with evidence of thermal overprinting in the Silurian (ca. 430-420 Ma) and Late Devonian (ca. 375-380 Ma). The Humber zone contains abundant muscovite-bearing metamorphic rocks that have undergone greenschist-grade regional metamorphism during the Middle to Late Ordovician and the Silurian to Early Devonian (Castonguay et al., 2001, 2012; Castonguay, Ruffet and Tremblay, 2007), and represents, with the Grenville Province, potential sources for the detrital muscovite record of the studied rock units. The Late Devonian overprint recognized in sample 09MILAN01 is however attributed to post-depositional Acadian regional metamorphism.

4.3.1 Age and provenance interpretations

In order to better constrain the age and provenance of the studied rock units, the stratigraphic and geochronological data that are presented above are examined in the following sections according to the revised stratigraphic framework of the St-Victor synclinorium and Connecticut Valley – Gaspé trough, which is synthesized in figure 4.11.

4.3.1.1 Bolton Igneous Group and Ware Volcanics

The Saint-Daniel Mélange has been attributed a Llanvirn to earliest Caradocian age based on newly-acquired $^{40}Ar/^{39}Ar$ muscovite and amphibole ages on its underlying ophiolitic-metamorphic basement, and biostratigraphic age constraints for overlying Magog Group slates (De Souza et al., 2012; Tremblay, Ruffet and Bédard, 2011). Consequently, although isotopic dating of the Bolton Igneous Group and Ware Volcanics has been unsuccessful, these rock units are dominated by volcanic flows that are interlayered within the Saint-Daniel Mélange slates and can thus be attributed to a similar age interval.



Gaspé belt
Connecticut valley sequence
Saint-Victor synclinorium
Connecticut valley - Gaspé through
Devonian
Middle
Early
Givetian
390
Eifelian
400
Emsian
Praghian
410
Lockhovian
Silurian
Pridoli
420
Lludlow
Wenlock
430
Llandovery
440
Ordovician
Late
Middle
Ashgill
450
Caradoc
455
460
Llanvirn
465
Arenig
Saint-Francis Group
*G+F
Compton Formation
ACF
LLF
SL-CR-LA-GL
Frontenac Formation
Gile Mountain Formation
Waits River Fm.
Northfield / Shaw Mt. Fm.
*E
*D
*C
*B
*A
MOHO
Beauce
Thetford-Mines
West shore
East shore
Lake Memphremagog
Northern Vermont
Southern Quebec
Gaspé belt
Vermont
Connecticut valley sequence
Magog Group
Saint-Victor Formation
Beauceville Formation
Etchemin Formation
Bunker Hill sequence volcanic member
Frontière Formation
Saint-Daniel Mélange
Bolton Igneous Group volcanic / plutonic rocks
Ware Volcanics
Slate, sandstone, pebbly mudstone
Debris flow breccia and conglomerate
Ophiolite and metamorphic sole
Bunker Hill sequence sedimentary member
Humber zone
Rowe - Hawley belt
Cram Hill Formation
Coburn Hill Volcanics / Mt. Norris Intrusive Suite
Felsic volcanics
Slate, sandstone, pebbly mudstone
Umbrella Hill Formation
Humber zone

Figure 4.11 : Schematic synthesis of the main stratigraphic relationships between the Humber and Dunnage zones of the Beauce, Thetford-Mines, and Lake Memphremagog areas, and the various units of the Gaspé Belt of southern Quebec, as well as the inferred correlations with the northern Vermont Rowe – Hawley belt and Connecticut Valley sequence, as proposed in this article. See text for further discussions. ACF – Ayers Cliff Formation; CR – Cranbourne Formation; GL – Glenbrooke Group; LA – Lac Aylmer Formation; LLF – Lac Lambton Formation; SL – Saint-Luc Formation. The samples presented in the text are symbolized by asterisks: *A* – 07BUNKER; *B* – 08M44; *C* – 10BUNKER01; *D* – 09SV01; *E* – 09SV02; *F* – 09MILAN02; *G* – 09MILAN01. Wavy lines indicate the presence of unconformities and/or paraconformities (see fig. 4.2 for line symbols). The stratigraphy of the Gaspé and Rowe-Hawley belts are modified from Lavoie and Asselin (2004) and Doolan et al. (1982), whereas the one of the Connecticut Valley sequence is shown as compiled in McWilliams, Walsh and Wintsch (2010). The Ordovician-Silurian and Devonian time scales are from Sadler, Cooper and Melchin (2009) and Gradstein, Ogg and Smith (2004), respectively. Note change in time scale at 450 Ma.

4.3.1.2 Bunker Hill sequence

Although the depositional age of the sedimentary member of the Bunker Hill sequence remains uncertain, upper and lower age limits can be suggested based on stratigraphic relationships with adjacent rock units and the detrital zircon age data presented herein. If our interpretation that the Saint-Daniel Mélange unconformably overlies and is in part derived from the erosion of the Bunker Hill greywackes, it can be inferred that the latter rocks were deposited prior to sedimentation of the Saint-Daniel. This is supported by the lack of Ordovician or younger detrital zircons in sample 07BUNKER. On the other hand, the youngest single grain zircon age measured in the sedimentary member, at 594 ± 29 Ma, grossly constrains the maximum depositional age of the formation to the Ediacaran period. The predominance of Mesoproterozoic zircons and the lack of Archean, early Paleozoic and Paleoproterozoic age populations in the Bunker Hill sequence greywackes, combined with their sialic composition and absence of felsic volcanic and ophiolitic detritus, strongly suggest that this unit derives mainly from Grenvillian crustal sources.

On the other hand, the available age constraints for the Saint-Daniel Mélange and the Magog Group slates set the age of the Bunker Hill sequence volcanic member to earliest Caradocian. The 455 ± 6 Ma age yielded by sample 10BUNKER01 is considered as the best absolute age estimate for its volcaniclastic rocks. The presence of detrital chromite grains, together with the felsic and calc-alkaline composition of the member indicates that it was formed in the proximity of a volcanic arc, and that it derives in part from the erosion of ultramafic rocks. The randomly-selected detrital zircon grain set of sample 10BUNKER01 also supports considerable recycling of Proterozoic to Archean basement and shows a much more broadly defined Precambrian zircon population than for sample 07BUNKER. Although the contribution of a western source component is obviously not excluded, a dominant eastern provenance for the Bunker Hill sequence volcanic member better accounts for its age and composition. This is consistent with the present coordinates of the Ascot Complex, its strong Laurentian inheritance (Tremblay, Hébert and Bergeron, 1989; Tremblay et al., 1994) and association with mafic-ultramafic rocks (Labbé, 1991; Hébert and Labbé, 1997).

4.3.1.3 Magog Group

The lower age limit of the Magog Group is considered as latest Llanvirnian to Caradocian based on its graptolite fauna (Cousineau and St-Julien, 1994), the maximum age attributed to the Saint-Daniel Mélange, and a 462 +5/-4 Ma zircon age for a felsic tuff of the Beauceville Formation (Marquis et al., 2001). Although the upper age limit of the Magog Group remains unclear, it is necessarily older than overlying Pridolian rocks of the Glenbrooke Group and Lac Aylmer Formation. In a paleogeographic study, Cousineau and St-Julien (1994) have concluded that the Frontière and Etchemin formations (i.e. the lowermost part of the Magog Group) derive in part from the erosion of a magmatic arc situated to the southeast of the basin, and that the Beauceville Formation marks a transition toward the northwest-derived terrigenous succession of the Saint-Victor Formation.

Our $^{40}Ar/^{39}Ar$ and U-Pb detrital muscovite and zircon age data for the Saint-Victor Formation bring significant new constraints on the age and provenance interpretation of the uppermost part of the Magog Group. The zircon age distribution for sample 09SV01 is exclusively characterized by Precambrian ages that can be compared with major episodes of Laurentian crust formation and assembly, such as those of the Grenville and Superior provinces and Paleoproterozoic orogens of the northeastern Canadian Shield (fig. 4.4c). The $^{40}Ar/^{39}Ar$ muscovite ages for this same sample indicate the predominance of a Grenvillian component, which may be derived either from Grenvillian basement, but more probably from low-grade Taconian nappes of the Humber zone as second-cycle sediment, thus confirming a northwestern source. On the other hand, the detrital zircon age distribution for sample 09SV02 shows a much less dominant Precambrian age population and delineates a strong input from Ordovician-Silurian Appalachian arcs (fig. 4.4d). However, the 424 ± 6 Ma weighted mean age that was calculated for the younger grains does not correspond to any known volcanic-plutonic unit in southern Quebec, but possible distal zircon sources include Silurian volcanic rocks of the Piermont Allochthon in western Maine (ca. 430-418 Ma; Moench et al., 1995), as well as late Llandoverian andesites and felsic volcaniclastic rocks of the Lac-Raymond and Pointe-aux-Trembles formations exposed in the western part of the Gaspé Peninsula (David and Gariépy, 1986; Bourque, Malo and Kirkwood, 2000). This Silurian age for the Magog Group is preliminary and should be validated by further work, but it suggests that these turbidites may be as young as Llandoverian-Ludlovian. If this age is correct, it can be concluded that the Magog Group represents a 25-35 Ma period of more-or-less continuous sedimentation that may have been interrupted by periods of erosion and/or of non-deposition, as it is in the Gaspé Peninsula (Tremblay, Malo and St-Julien, 1995).

4.3.1.4 Saint-Francis Group, Compton Formation

The age of the Saint-Francis Group is only loosely constrained. The Lac Lambton Formation is Pridolian-Early Devonian, whereas the overlying Ayer's Cliff and Compton formations have yielded Early Devonian and Pridolian to Early Devonian chitinozoan and plant remains (Hueber et al., 1990; Lavoie, 2004; Lavoie and Asselin, 2004). The upper age limit of the Saint-Francis Group is not defined, but it necessarily predates the Late Devonian Acadian low- to greenschist-grade regional metamorphic imprint in the area (380-375 Ma; Tremblay, Ruffet and Castonguay, 2000) and granitoid plutons that crosscut it (384-374 Ma; Simonetti and Doig, 1990). Further age and provenance interpretations can however be proposed by examining our U-Pb and $^{40}Ar/^{39}Ar$ data for the Compton Formation. Samples 09MILAN01 and 09MILAN02 show detrital zircon age distributions that suggest contributions from Archean and Proterozoic Laurentian sources, and Cambrian to Devonian magmatic rocks of the Northern Appalachians (figs. 4.4e, 4.4 f). However, Neoproterozoic ages that were obtained from sample 09MILAN02 and its extra grain set, between ca. 620-550 Ma, overlap with the age interval attributed to magmatism related to the opening of the Iapetus Ocean and peri-Gondwanan Avalonian arcs (figs. 4.4e, 4.4f). Conversely, the ca. 396 ± 6 Ma weighted mean age that was calculated for the 09MILAN02b grain set suggests that sedimentation of the Compton Formation has persisted at least into latest Emsian time.

The $^{40}Ar/^{39}Ar$ age data for samples 09MILAN01 and 09MILAN02 can be readily compared with the tectonic-metamorphic evolution of the Humber zone (Castonguay et al., 2001, 2012; Tremblay and Castonguay, 2002; Castonguay, Ruffet and Tremblay, 2007; Tremblay, Ruffet and

Bédard, 2011) : 1) the ca. 460-450 Ma high-temperature steps and ca. 459 Ma plateaus that characterize the age spectra of both samples correspond to obduction-related regional metamorphism and nappe emplacement, and 2) the ca. 433 Ma plateau age and ca. 437-435 Ma high-temperature steps in the age spectra of sample 09MILAN01, as well as the partial thermal overprinting recognized in the low-temperature steps of sample 09MILAN02 at ca. 430-420 Ma, can be attributed to hinterland-directed superposed deformation and related Silurian to Early Devonian metamorphism. The Silurian metamorphic imprint is only recorded in the low-temperature steps of the muscovite age spectra of muscovite sample 09MILAN02, whereas it persists into the medium- to high-temperature steps 09MILAN01, possibly suggesting the deeper exhumation and erosion of the Taconian orogenic wedge in the latter case. Also, the correlation of detrital muscovite ages from the Compton Formation with the tectonothermal evolution of the internal Humber zone suggests, preferably, that Ediacaran zircon ages obtained from sample 09MILAN02 derive from the erosion of rift-related magmatic rocks such as those preserved in Humber zone nappes of southern Quebec (Kumapareli et al., 1989; Hodych and Cox, 2007). Southeastern-derived sources composed of peri-Gondwanan arc-related rocks can not however be excluded.

4.3.2 Correlations and regional implications

Stratigraphic relationships, facies distribution and geochrological data compiled and presented herein can be used to propose a revised stratigraphic framework for the Dunnage zone. Key issues discussed in this section concern 1) volcaniclastic and Humber-like rocks of the Bunker Hill sequence, and 2) the Saint-Daniel Mélange and related volcanic-plutonic rocks of the Bolton Igneous Group, Ware Volcanics, and Ascot Complex. The correlations and interpretations that are discussed in the following section are illustrated and synthesized with the geology of the Connecticut Valley-Gaspé trough in figure 4.11. Although the southern Quebec ophiolites represent salient features of the study area, regional correlations regarding ophiolite stratigraphy and relationships with adjacent rock units of the Dunnage and Humber zones have been already published recently and will not be discussed further (Schroetter, Tremblay and Bédard, 2005; Schroetter et al., 2006; De Souza et al., 2008, 2012; De Souza and Tremblay, 2010a).

4.3.2.1 Bunker Hill sequence

As proposed above, the volcanic and sedimentary members of the Bunker Hill sequence should be treated as two distinct lithostratigraphic units. Our mapping and detrital zircon U-Pb data for sample 07BUNKER suggest that the sedimentary member derives mostly from the erosion of Grenvillian crystalline basement, and that it was deposited sometime between the late Neoproterozoic and Middle Ordovician, prior to the emplacement, uplift, and erosion of ophiolites onto Laurentia. Moreover, the lithostratigraphic and tectonic history of these Humber-like rocks can be reconciled with the Humber zone *s.s.*, more specifically with the Caldwell Group, which mostly consists of quartzo-feldspathic sandstone belonging to the Neoproterozoic to Early Cambrian rift sequence of Laurentia in southern Quebec (Tawadros, 1977; Cousineau, 1990). Feldspathic sandstones petrographically similar to the Caldwell Group also occur in the Beauce area to the southeast of the Baie Verte – Brompton line (fig. 4.3; Cousineau, 1990; Cousineau and St-Julien, 1992) and possibly correlate with the Bunker Hill sequence sedimentary member. We believe that Humber-like rocks of the Bunker Hill sequence sedimentary member, and possibly those of the Beauce area, correlate with the Caldwell Group and represent antiformal inliers through the Dunnage zone. Such inliers have been documented in the Thetford-Mines area and in the northern Vermont Rowe-Hawley belt, where greenschist-grade quartzo-feldspathic sandstone and phyllite attributed to the Caldwell Group and Stowe Formation, occur structurally beneath ophiolitic rocks and the Cram Hill Formation, respectively (figs. 4.1, 4.2, 4.11; Doolan et al., 1982; Tremblay and Pinet, 1994; Tremblay and Castonguay, 2002; Schroetter, Tremblay and Bédard, 2005; Tremblay, Ruffet and Bédard, 2011). The Chain Lakes massif, a quartzofeldspathic gneiss and schist unit exposed in western Maine and southern Quebec (see fig. 4.1 for location; Boone and Boudette, 1989; Gerbi, Johnson and Aleinikoff, 2006), is another rock unit that has been correlated with the Humber zone of southern Quebec (Cousineau, 1991; Pinet and Tremblay, 1995b; De Souza and Tremblay, 2010a). Recent U-Pb dating of detrital zircons from the Chain Lakes massif clearly indicates the predominance of a Laurentian source, with significant Archean, Paleoproterozoic and Meso- to early Neoproterozoic age clusters and a minor peak in late Neoproterozoic (Gerbi, Johnson and Aleinikoff, 2006), an age spectrum that is compatible with the one obtained for sample 07BUNKER and suggests a similar provenance for both units (fig. 4.12). According to our data and correlations presented herein, it can be interpreted that low- to greenschist-grade Humber zone rocks underlie much of the Saint-Victor synclinorium Dunnage zone, and possibly extend into highly deformed and metamorphosed rocks of the Chain Lakes massif.

The Bunker Hill sequence volcanic member can rather be compared with the stratigraphic record of the lower Magog Group in the Beauce area (fig. 4.11). As for the Frontière and Etchemin formations, the most distinguishing features of the volcanic member are its inferred stratigraphic position between the Saint-Daniel Mélange and the Saint-Victor Formation turbidites, the presence of detrital chromite grains and the abundance of inherited Precambrian zircons. The overall composition and detrital zircon age populations of the volcanic member, as well as its relationships with the Saint-Daniel Mélange and Magog Group, is thus in compliance with the regional stratigraphic setting of the Saint-Victor synclinorium (fig. 4.11). Moreover, the 455 ± 6 Ma weighted mean age obtained for sample 10BUNKER01 overlaps with the inferred Llanvirnian to Caradocian age of the Frontière and Etchemin formations (Cousineau and St-Julien, 1994) and the 462 +5/-4 Ma U-Pb zircon age yielded by felsic tuff of the Beauceville Formation (Marquis et al., 2001). The predominance of felsic tuffs and volcaniclastic rocks, together with the lack of granitic rocks and extensive volcanic flows in the Bunker Hill sequence rules out any direct correlation with the Ascot Complex. The latter may however represent a viable source of felsic detritus and volcanic ejecta for the Bunker Hill sequence and correlative units of the lower Magog Group.

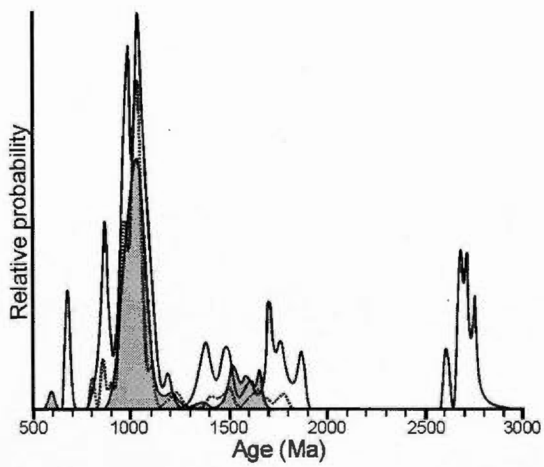


Figure 4.12 : Relative probability plot comparing U-Pb detrital zircon ages for sample 07BUNKER (shaded area) with those for metasedimentary rocks of the Chain Lakes massif (full and dotted lines; Gerbi, Johnson and Aleinikoff, 2006).

4.3.2.2 Saint-Daniel Mélange and related volcanic rocks

The most common and widely represented rock-types in the Saint-Daniel Mélange are interlayered black and green slate, lithic sandstone, dolomitic siltstone and quartzite, and a diagnostic intraformational pebbly mudstone breccia (Lavoie, 1989; Cousineau, 1990; Schroetter et al., 2006). Ophiolitic and metamorphic clast-bearing polymictic debris flow breccias and conglomerates found at the base of the Saint-Daniel Mélange, have been documented from various areas of southern Quebec and mark an unconformity at the top of ophiolites and related metamorphic rocks (Schroetter et al., 2006; De Souza et al., 2008, 2012; De Souza and Tremblay, 2010a). In northern Vermont, the Umbrella Hill Formation consists of debris flow conglomerates occurring at the base of the Cram Hill Formation (Doolan et al., 1982). It contains abundant vein quartz, metasedimentary and sedimentary rock fragments (Badger, 1980; Doolan et al., 1982), and unconformably overlies previously metamorphosed and deformed rocks assigned to the Humber zone (Doolan et al., 1982). The Umbrella Hill Formation is located at the same stratigraphic position as the Saint-Daniel Mélange polymictic debris flows, and both units contain abundant continent-derived clasts (Doolan et al., 1982; Schroetter et al., 2006). Therefore, we interpret both units as time-correlative marker horizons of a Taconian unconformity. However, the lack of ophiolites and related rocks beneath the unconformity in northern Vermont, as well as the absence of ophiolite-derived detritus in the Umbrella Hill Formation, suggests that ophiolitic nappes, now represented by the southern Quebec ophiolite belt, either did not originally extend into northern Vermont, or were locally completely eroded.

Mafic volcanic rocks and dykes of the Bolton Igneous Group can be geochemically and petrographically correlated with the Coburn Hill Volcanics and Mount Norris Intrusive Suite of the Rowe-Hawley belt, respectively (fig. 4.2; Doolan et al., 1982; Kim et al., 2003). Nevertheless, in addition to the Cram Hill Formation, mafic dykes of the Mount Norris Intrusive Suite are known to crosscut previously-foliated rocks of the Moretown and Stowe formations, and can be interpreted as the result of syn- to post-tectonic magmatism (Kim et al., 2003). Altogether, these units represent an extensive and significant igneous event in the Quebec-Vermont Appalachians, with field relationships indicating that the related magmas crosscut and were locally erupted onto deformed Humber zone rocks, and are interlayered with and overlain by Saint-Daniel Mélange-Cram Hill Formation slates.

Although the Bolton Igneous Group and Ware volcanics do not have similar compositions, these units share similar relationships with the Saint-Daniel Mélange and were thus formed in the same tectonic setting. The Ware Volcanics form the only extensive felsic volcanic flows found within the Saint-Daniel Mélange, but felsic tuffs are locally very abundant (Schroetter et al., 2006). Felsic volcanic rocks also occur in the Cram Hill Formation (Gale, 1980) and possibly correlate with those of the Ware Volcanics.

Llanvirn to early Caradocian (ca. 460-463 Ma) volcanic rocks are also found as part of the Ascot Complex in association with Saint-Daniel-type phyllites, and appear to be more-or-less coeval with the Bolton Igneous Group and Ware Volcanics. However, the Ascot Complex can be distinguished from the latter volcanic rocks on the basis of several grounds. First, Ascot Complex basalts show a wide range of geochemical affinities, varying from mid-ocean ridge-type basalts and island arc tholeiites to calc-alkaline and boninitic lavas (Tremblay, Hébert and Bergeron, 1989; Hébert and Labbé, 1997). The bimodal composition of the Ascot Complex and the widespread occurrence of syn-volcanic granitoids, as well as its association with ultramafic rocks preclude any further correlation with the Bolton Igneous Group and Ware Volcanics. The Ascot Complex volcanic and plutonic rocks have been interpreted as tectonic slivers that were thrust into the Saint-Daniel Mélange (Tremblay and St-Julien, 1990; Tremblay, 1992a), but the volcanic rock-phyllite contact is locally depositional (Gauthier et al., 1989), which suggests that at least part of the Ascot Complex forms a volcanic arc basement that is overlain by and/or interdigitated with Saint-Daniel-type phyllites.

4.3.3 Paleotectonic interpretations

The isotopic age data and stratigraphic relationships presented in this article highlight significant differences in the age, provenance and origin of Neoproterozoic? to Devonian units of the Saint-Victor synclinorium and Connecticut Valley – Gaspé trough. Nevertheless, these discrepancies can be reconciled with existing geological data for the Quebec-Vermont Appalachians in order to speculate on the paleotectonic evolution of the Laurentian margin. In southern Quebec and western New England, Taconian accretionary events include the Early to Late Ordovician emplacement of the southern Quebec ophiolites, as well as the formation and accretion of the Shelburne Falls, Ascot Complex and Bronson Hill volcanic arc terranes in Ordovician to Silurian times (Doolan et al., 1982; Stanley and Ratcliffe, 1985; Tremblay, 1992a; Pinet and Tremblay, 1995b; Karabinos et al., 1998; Hollocher, Bull and Robinson, 2002; Tremblay and Castonguay, 2002). Structural characteristics of the Laurentian margin and the current disposition of ophiolites, mélanges, flysch units and arc volcanics all suggest that, outboard of Laurentia, plate convergence was accommodated by an east-dipping (present coordinates) subduction zone throughout Middle Cambrian to Late Ordovician time (Osberg, 1978; Stanley and Ratcliffe, 1985; Pinet and Tremblay, 1995b; Robinson et al., 1998; Moench and Aleinikoff, 2002; Hollocher, Bull and Robinson, 2002; Rankin et al., 2007). It is however uncertain if the final closure of the Iapetus Ocean, as a result of the accretion of peri-Gondwanan Gander margin rocks, was accomplished above a subduction zone dipping beneath (van Staal et al., 1998; Rankin et al., 2007) or away from Laurentia (Tremblay and Pinet, 2005). The inferred Ordovician to Devonian paleogeographic and tectonic evolution of the Laurentian margin, as deduced from our observations and geochronological data, and previous work by numerous authors in the southern Quebec and New England Appalachians, is illustrated in fig. 4.13 and can be interpreted as follows.

Taconian obduction and slab breakoff

Uprooting and thrusting of oceanic lithosphere toward and onto Laurentia was initiated at ca. 479-472 Ma, and continued for a period of approximately 12-to-22 Myr, as ophiolitic nappes were translated over the margin and uplifted with underlying metamorphic rocks at ca. 460 Ma (Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012). By this time ophiolitic and continental metasedimentary rocks formed an orogenic wedge that was eroded and exhumed, probably as a forearc ridge(s) from which detritus was recycled oceanward onto the forearc oceanic basement to form olistostromal and conglomeratic units of the Saint-Daniel Mélange (fig. 4.13a; Schroetter et al., 2006; Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012). Uplift and erosion of the amalgamated forearc block were however irregular along the strike of the collision zone, with the erosion level reaching ophiolitic crustal rocks in Thetford-Mines, but cutting deeper into the exhumed continental margin in the vicinity of the Quebec-Vermont border.

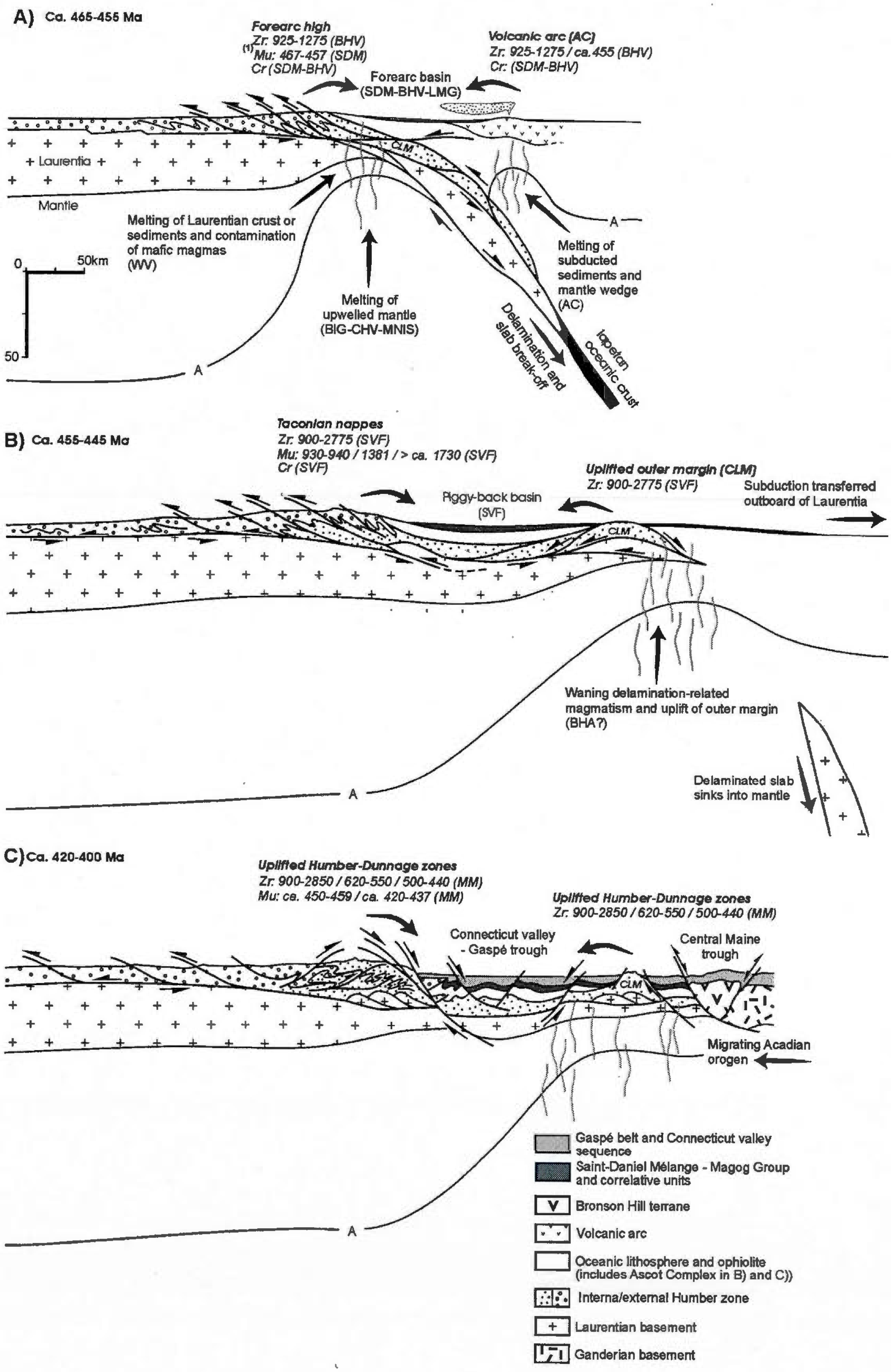

A) Ca. 465-455 Ma
Forearc high
(1) Zr: 925-1275 (BHV)
Mu: 467-457 (SDM)
Cr (SDM-BHV)
Forearc basin
(SDM-BHV-LMG)
Volcanic arc (AC)
Zr: 925-1275 / ca. 455 (BHV)
Cr: (SDM-BHV)
Laurentia
CLM
Mantle
Melting of Laurentian crust or sediments and contamination of mafic magmas (WV)
0
50km
50
A
Melting of upwelled mantle (BIG-CHV-MNIS)
Melting of subducted sediments and mantle wedge (AC)
Delamination and slab break-off
Iapetan oceanic crust
B) Ca. 455-445 Ma
Taconian nappes
Zr: 900-2775 (SVF)
Mu: 930-940 / 1361 / > ca. 1730 (SVF)
Cr (SVF)
Piggy-back basin
(SVF)
Uplifted outer margin (CLM)
Zr: 900-2775 (SVF)
Subduction transferred outboard of Laurentia
CLM
Waning delamination-related magmatism and uplift of outer margin (BHA?)
Delaminated slab sinks into mantle
A
C) Ca. 420-400 Ma
Uplifted Humber-Dunnage zones
Zr: 900-2850 / 620-550 / 500-440 (MM)
Mu: ca. 450-459 / ca. 420-437 (MM)
Connecticut valley - Gaspé trough
Uplifted Humber-Dunnage zones
Zr: 900-2850 / 620-550 / 500-440 (MM)
Central Maine trough
CLM
Migrating Acadian orogen
A
Gaspé belt and Connecticut valley sequence
Saint-Daniel Mélange - Magog Group and correlative units
Bronson Hill terrane
Volcanic arc
Oceanic lithosphere and ophiolite (includes Ascot Complex in B) and C))
Interna/external Humber zone
Laurentian basement
Ganderian basement

Figure 4.13 : Schematic model for the tectonostratigraphic evolution of the southern Quebec and northwestern New England Appalachians in Middle Ordovician to Middle Devonian. The detrital zircon, muscovite and chromite mineral record as presented and synthesized in this article are indicated in italic with the inferred provenance (bold lettering) and host rock unit(s) (acronyms in parentheses). a) Delamination and breakoff of subducted Laurentian lithosphere and formation of the Saint-Daniel Mélange, Bolton Igneous Group, Ware Volcanics and lowermost stratigraphic units of the Magog Group in a syn-obduction forearc basin; b) Delamination of the subducted lithosphere is complete and deformation migrates toward the foreland during the emplacement of the Taconian nappes; c) extensional collapse of the orogen and formation of the Connecticut Valley – Gaspé trough. Note that only a) is scaled with a vertical exaggeration of ~ 1.5. Dark grey wavy lines underline the presence of migrating magma. A – asthenosphere-lithoshere boundary; AC – Ascot Complex; BHA – Bronson Hill arc; BIG – Bolton Igneous Group; CHV – Hill volcanics; CLM – Chain Lakes massif; Cr – chromite; BHV Bunker Hill sequence volcanic member; FEF – Frontière and Etchemin formations; LMG – lower units of Magog Group (Frontière, Etchemin and Beauceville formations); MM – Milan member; MNIS – Mount Norris Intrusive Suite; Mu – muscovite; SDM – Saint-Daniel Mélange; SVF – Saint-Victor Formation; WV – Ware Volcanics; Zr – zircon. (1) refers to detrital muscovite $^{40}Ar/^{39}Ar$ age data presented in previous work by Schroetter et al. (2006) and Tremblay, Ruffet and Bédard (2011).

As recorded in the Ascot Complex (Tremblay, Ruffet and Castonguay, 2000), renewed/incipient arc magmatism along Laurentia was penecontemporaneous with this period of Taconian orogenesis, as well as with the formation of the Bolton Igneous Group, Mount Norris Intrusive Suite and Ware Volcanics. We think that this magmatism can be reconciled when considering the introduction of continental material into the subduction/collision zone. The subduction (or attempted subduction) of buoyant continental lithosphere is an inherent part of arc-continent collision zones and ophiolite emplacement, that frequently leads to the crustal contamination of subduction zone magmas, delamination-related breakoff of the downgoing slab, thus providing a favorable setting for accelerated uplift of the orogenic wedge and the generation of mafic magmatism due to asthenospheric upwelling (Cloos et al., 2005; Dewey, 2005; Alfonzo and Zlotnik, 2011; Brown et al., 2011). Thermochronological studies (Tremblay, Ruffet and Bédard, 2011; De Souza et al., 2012) and paleotectonic data (Pinet and Tremblay, 1995b; De Souza and Tremblay, 2010a) suggest that a minimum distance of ca. 100 km of Laurentian margin was underthrusted and possibly subducted during the emplacement of the southern Quebec ophiolites. Consequently, the introduction of subducted sediments into the subduction melting zone may have triggered the formation of abundant felsic magmas (Johnson and Plank, 1999; Shimoda and Tatsumi, 1999; Shimoda, Tatsumi and Morishita, 2003), which would then account for the bimodal composition, the inherited crustal signature and strong geochemical variations of supra-subduction zone magmas that characterize the Ascot Complex igneous rocks (fig. 4.13a; Tremblay, Hébert and Bergeron, 1989; Tremblay et al., 1994; David and Marquis, 1994). Subducting slab breakoff has been proposed to account for the Mount Norris Intrusive Suite (Kim et al., 2003; Coish et al., 2010) and may well be applied to both the Bolton Igneous Group and Ware Volcanics. Collisional delamination resulting in subducting slab breakoff has been documented in New Guinea, where the northern margin of the Australian continent has subducted northward beneath the Pacific Plate during emplacement of the Irian Ophiolite Belt (Cloos et al., 2005). There, the subducting plate made up of continental and oceanic lithosphere, delaminated and foundered into the asthenosphere, which has resulted in the adiabatic partial melting of the upwelled mantle or continental crust and in the formation of mafic and felsic magmas that intruded the orogenic wedge built up along the Australian Plate margin, ca. 10 Myr after the initial impingement of continental material into the subduction zone (Cloos et al., 2005). This model fits well with stratigraphic relationships and age data presented herein for southern Quebec and northern Vermont (fig. 4.13a).

As collision proceeded, continued magmatism and erosion of the volcanic arc resulted in the dispersal of felsic pyroclastic and volcaniclastic rocks throughout the forearc region to form the Frontière and Etchemin formations, and the Bunker Hill sequence volcanic member, above the unconformity marking the base of the Magog Group. The Beauceville Formation was subsequently deposited as the basin subsided and the volcanic arc was progressively shutting down.

Emplacement of Taconian nappes

By 460-457 Ma, the obduction of the ophiolites *s.s.* was more-or-less complete and thrusting was transferred to the Humber zone continental interior via a foreland-propagating piggy-back thrust system (Tremblay, Ruffet and Bédard, 2011), which was active during most of the Late Ordovician, until ca. 445 Ma (St-Julien and Hubert, 1975; Tremblay and Castonguay, 2002; Sasseville et al., 2008). During this time period, detritus shed from low-grade Taconian nappes of the Humber zone, as well as ophiolitic and/or chromite-bearing sedimentary rocks were transported into the basin, accounting for the detrital zircon and muscovite record of the lower part of the Saint-Victor Formation turbidites (fig. 4.13b). It can also be inferred that isostatic rebound, which is expected following subducting slab breakoff (Cloos et al., 2005; Alfonso and Zotlik, 2011), has resulted in the subaerial exposure and erosion of the outer Laurentian margin, part of which is possibly preserved in the Chain Lakes massif (fig. 4.13b; Pinet and Tremblay, 1995b; De Souza and Tremblay, 2010a).

As convergence continued following slab breakoff, subduction along Laurentia has either undergone a polarity flip (continentward subduction) or transferred oceanward to a new or existing subduction zone. Although subduction beneath Laurentia has been suggested as of the Late Ordovician-Silurian (Karabinos et al., 1998; Rankin et al., 2007; van Staal, 2007; Dorais, Workman and Aggarwal, 2008; van Staal et al., 2008; Dorais et al., 2012), the lack of stratigraphic evidence for the development of an Andean-type margin and formation of a consequent accretionary complex along the outer margin of Laurentia does not favor the polarity flip model (Pinet and Tremblay, 1995b; Hollocher, Bull and Robinson, 2002; Tremblay and Pinet, 2005).

By late Llandoverian-Wenlockian time, the Gander margin collided with composite Laurentia, a tectonic event that is generally referred to as the Salinic orogeny in maritime Canada (Dunning et al., 1990; van Staal, 2007; van Staal et al., 2008), but the polarity of the subduction zone(s) leading to this accretionary event in New England and Quebec remains speculative (Tremblay and Pinet, 2005; Aleinikoff et al., 2007; Rankin et al., 2007; Wintsch et al. 2007; van Staal et al., 2008). If our inferred Silurian age for the uppermost part of the Saint-Victor Formation is correct, it would suggest that the Magog Group had evolved from a forearc basin in the Late Ordovician, to an intra- or peri-continental basin during the Silurian, as emplacement and deformation of the Taconian nappes culminated with the formation of hinterland-directed folds and faults (Tremblay and Castonguay, 2002; Castonguay and Tremblay, 2003; Castonguay, Ruffet and Tremblay, 2007; Castonguay et al., 2012).

Post-Taconian basin formation

Extensional collapse of the Taconian orogen and formation of the Connecticut Valley – Gaspé trough cover sequence was initiated in Pridolian time in southern Quebec (Tremblay and Castonguay, 2002; Tremblay and Pinet, 2005). Silurian to Early Devonian (ca. 417-405 Ma; Castonguay et al., 2001, 2012; Castonguay, Ruffet and Tremblay, 2007) normal faulting along the Baie Verte – Brompton line and Saint-Joseph fault in southern Quebec and the Burgess Branch fault in northern Vermont (Kim et al., 1999; Castonguay and Tremblay, 2003; Castonguay et al., 2012), provides a mechanism for basin formation, as well as for the uplift and erosion of the Laurentian margin and accreted oceanic and pericontinental igneous and sedimentary rocks of the Humber and Dunnage zones, which account for much of the detrital muscovite and zircon populations in the Compton Formation (fig. 4.13c). Sedimentation in the trough has persisted throughout the Early Devonian and came to an end with the arrival of the Acadian deformation front in the Middle Devonian (Bradley et al., 2000; Bradley and Tucker, 2002; Tremblay, Ruffet and Castonguay, 2000).

4.4 Conclusion

This article presents and synthesizes statigraphic and isotopic age data regarding the tectonic evolution of syn- to post-Taconian stratigraphic units of the southern Quebec Appalachians and their correlation with adjacent units of northern Vermont. The proposed stratigraphic framework suggests that Cambrian-Ordovician rocks of the Saint-Victor synclinorium and the northern Vermont part of the Rowe-Hawley belt have undergone a geologic history that can be synthesized into a tectonic model that accounts for the along-strike lithologic and stratigraphic variations, and the geochronological record of both areas.

The final stages of oceanic lithosphere emplacement onto Laurentia during the Taconian orogeny gave rise to the formation of a Llanvirn to Caradocian forearc basin. The Saint-Daniel Mélange, correlative units of the Rowe-Hawley belt, and overlying rocks of the lower Magog Group and Bunker Hill sequence volcanic member were successively deposited onto a basement comprising obducted ophiolites and continental rocks belonging to the Humber zone (e.g. the Stowe Formation and Bunker Hill sequence sedimentary member). The basin accumulated debris shed from uplifted forearc highs composed of Cambrian-Ordovician ophiolites and metamorphic rocks, and volcanic debris from an outboard volcanic arc represented by the Ascot Complex. This period of syncollisional uplifting has been synchronous with the interruption of southeast-directed subduction (away from Laurentia), as the subducted margin delaminated and foundered into the mantle. Conformable volcanic flows and their intrusive counterparts that occur within the Saint-Daniel Mélange, and which are mostly represented by the Bolton Igneous Group and Ware Volcanics in southern Quebec, and the Coburn Hill Volcanics and Mount Norris Intrusive Suite in northern Vermont, were formed as a result of delamination and related asthenospheric upwelling beneath the collision zone. Tuffs and volcaniclastic rocks of the Bunker Hill sequence volcanic member and lower Magog Group were deposited unconformably onto the Saint-Daniel Mélange as the outboard arc, represented by the Ascot Complex, was still active but progressively shutting down. Onlapping turbidites of the Saint-Victor Formation record the erosion of the external nappe domain as Taconian deformation progressed toward the foreland region. Collapse of the Taconian orogen has resulted in the formation of the Connecticut Valley – Gaspé trough, which accumulated sediment eroded away from uplifted and previously assembled oceanic and continental rocks of Laurentian affinity throughout the Early Devonian. A possible contribution from accreted peri-Gondwanan terrains into the basin remains uncertain, but the lack of clear evidence for such an input suggests that the Iapetan suture is probably located southeast of the Chain Lakes massif.

4.5 Acknowledgements

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CONCLUSIONS GÉNÉRALES

Cette thèse comporte trois chapitres dont l'objectif principal vise à déterminer l'évolution tectonostratigraphique du domaine océanique des Appalaches du sud du Québec, ainsi que de d'apporter une contribution scientifique plus fondamentale sur notre compréhension des chaînes d'obduction en général. Une approche méthodologique combinant des travaux de cartographie géologique et de géochronologie $^{40}Ar/^{39}Ar$ et U-Pb a été employée afin de résoudre des problématiques spécifiques à chacune des unités et régions à l'étude. Ces travaux ont principalement été concentrés dans les régions du lac Memphrémagog au sud-ouest, et de la Beauce au nord-est, permettant ainsi d'avoir une vision globale des terrains étudiés et de leur variation le long de la chaîne taconienne. Les ophiolites et le Mélange de Saint-Daniel ont toutefois représenté le cœur de cette étude en raison de leur étendue géographique, leur histoire géologique complexe et des variations lithologiques et/ou compositionnelles qui peuvent y être observées d'une région à l'autre.

L'article présenté dans le second chapitre de la thèse porte essentiellement sur la géologie du Complexe ultramafique de la Rivière-des-Plante. L'étude détaillée d'affleurements-clé combinée à une cartographie plus systématique des unités ophiolitiques et sédimentaires qui lui sont adjacentes a permis de mettre en évidence son origine et son évolution géologique.

Il a d'abord été démontré que le Complexe ultramafique de la Rivière-des-Plante représente un massif ophiolitique profondément érodé qui est corrélé avec les complexes ophiolitiques de Thetford-Mines, Asbestos et Lac-Brompton. Ses contacts sud-est et nord-ouest avec le Mélange de Saint-Daniel et le Groupe de Caldwell correspondent, respectivement, à une discordance angulaire marquée par la présence de brèche sédimentaire polygénique, et à une faille de rétro-chevauchement dont le tracé correspond à celui de la ligne Baie Verte – Brompton dans cette région. L'unité felsique fragmentaire qui caractérise le Complexe ultramafique de la Rivière-des-Plante n'est ni exotique, ni d'origine métasédimentaire. Elle est plutôt constituée de roches granitiques qui sont intrusives dans les roches ultramafiques. La texture fragmentaire de ces granitoïdes est due à l'abondance de xénolithes, tandis que la présence de faciès déformés et mylonitiques témoigne de leur origine syn-tectonique. Les brèches du Mélange de Saint-Daniel contiennent une forte proportion de blocs constitués de granitoïde folié et non-folié riche en xénolithes, indiquant clairement que la déformation ductile des roches granitiques du Complexe ultramafique de la Rivière-des-Plante a précédé la sédimentation du Mélange de Saint-Daniel. Les caractéristiques pétrographiques ainsi que la composition minéralogique et géochimique des granitoïdes du Complexe ultramafique de la Rivière-des-Plante suggèrent un lien génétique avec les granitoïdes syn-obduction qui recoupent l'unité mantellique de l'ophiolite de Thetford-Mines. Ceux-ci ont été datées à 470 $^{+5}/_{-3}$ Ma et 469 ± 4 Ma et sont attribués à la fusion partielle de la marge laurentienne pendant la mise en place de l'ophiolite (Whitehead, Dunning et Spray, 2000; Tremblay, Ruffet et Bédard, 2011).

La corrélation de l'unité felsique fragmentaire du Complexe ultramafique de la Rivière-des-Plante avec les roches métamorphiques du massif de Chain Lakes est ici infirmée. Toutefois, l'âge de crystallisation des granitoïdes d'obduction correspond à l'âge du métamorphisme et de la migmatization qui ont affecté le massif de Chain Lakes à ca. 468-469 Ma (Dunning et Cousineau, 1990; Gerbi, Johnson et Aleinikoff, 2006). Ces contraintes géochronologiques, ainsi que l'affinité continentale laurentienne du massif de Chain Lakes, suggèrent que ce dernier représente une source viable pour les magmas felsiques qui ont contribué à la formation des granitoïdes d'obduction préservés au sein de la ceinture ophiolitique du sud du Québec. Les caractéristiques structurales et pétrologiques du massif de Chain Lakes sont comparables à celles des roches métamorphiques de la zone de Humber interne, qui sont localement exposées à la faveur de fenêtres structurales, sous l'ophiolite de Thetford-Mines (i.e. le dôme de Bécancour). Il est donc proposé que les ophiolites s'enracinent au sud-est du massif de Chain Lakes et ont été transportées jusqu'à leur position actuelle durant l'orogénie taconienne.

Le second chapitre de la thèse fait suite aux travaux présentés sur le Complexe ultramafique de la Rivière-des-Plante. Il inclut notamment des données cartographiques et géochronologiques $^{40}Ar/^{39}Ar$ sur l'ophiolite du Lac-Brompton, le Complexe ultramafique de la Rivière-des-Plante, ainsi que le Mélange ophiolitique de Nadeau dans la péninsule gaspésienne. Les résultats de cette étude permettent de mettre en évidence la corrélation des massifs ophiolitiques situés le long de la ligne Baie Verte – Brompton, entre le sud du Québec et la Gaspésie, ainsi que la chronologie et la nature des processus associés à leur obduction sur la marge laurentienne, et à la formation de certains mélanges qui leur sont associés.

Les échantillons qui ont été datés pour cette partie de la thèse proviennent de gabbro ophiolitique et de roches métasédimentaires et métabasiques de la semelle métamorphique de l'ophiolite du Lac-Brompton, des granitoïdes d'obduction du Complexe ultramafique de la Rivière-des-Plante, ainsi que de roches granitiques et métasédimentaires du Mélange ophiolitique de Nadeau.

Les âges $^{40}Ar/^{39}Ar$ sur amphibole et muscovite pour l'ophiolite du Lac-Brompton sont comparables à ceux obtenus pour des faciès semblables de l'ophiolite de Thetford-Mines. Il est donc suggéré que ces massifs ophiolitiques ont connu une évolution thermochronologique similaire. D'abord, un âge sur amphibole de 476.7 ± 6.2 Ma pour un gabbro de l'ophiolite du Lac-Brompton est concordant avec les âges moyens de 479.2 ± 1.6 Ma et de 477.6 ± 3.5 Ma obtenus par datation $^{40}Ar/^{39}Ar$ sur les amphiboles et U-Pb sur les zircons provenant de gabbro et de plagiogranite de l'ophiolite de Thetford-Mines (Whitehead, Dunning et Spray, 2000; Tremblay, Ruffet et Bédard, 2011). Cette concordance suggère que la cristallisation et le refroidissement des roches crustales de ces deux ophiolites ont été synchrones. D'autre part, nos données sur la semelle métamorphique de l'ophiolite du Lac-Brompton indiquent qu'elles ont été exhumées et ont refroidi sous la température de fermeture de la muscovite à ca. 467 Ma, et que les amphiboles ont été recristallisées à ca. 463 Ma durant le charriage de l'ophiolite sur la marge laurentienne. Cette évolution se compare à celle de la semelle de l'ophiolite de Thetford-Mines, avec des âges sur amphibole et muscovite qui suggèrent un refroidissement progressif sous les températures de fermeture de ces minéraux à ca. 471 et 466 Ma, respectivement, et une recristallisation tardive jusqu'à ca. 460-457 Ma (Tremblay, Ruffet et Bédard, 2011). Les âges obtenus sur les granitoïdes d'obduction du Complexe ultramafique de la Rivière-des-Plante sont identiques à ceux de l'ophiolite de Thetford-Mines et indiquent un refroidissement initial sous la température de fermeture de la muscovite à 466 Ma, et un événement de recristallisation à 461 Ma qui semble associé à un épisode de déformation ductile. Nos données géochronologiques confirment l'hypothèse selon laquelle l'ensemble des ophiolites du sud du Québec peuvent être corrélées stratigraphiquement et forment les vestiges d'un même segment de lihosphère océanique obductée. Des âges $^{40}Ar/^{39}Ar$ sur muscovite de ca. 475 et 470 ont été obtenus pour des granites d'obduction et les schistes micacés du Mélange ophiolitique de Nadeau. Ceux-ci sont plus anciens que dans le sud du Québec, ce qui suggère que le Mélange ophiolitique de Nadeau a connu une histoire thermochronologique différente.

Lorsque combinées aux relations stratigraphiques observées sur le terrain, ces contraintes géochronologiques ont également des incidences sur l'âge et l'origine du Mélange de Saint-Daniel. La présence de fragments dérivés des roches ophiolitiques, des granitoïdes d'obduction et des roches métamorphiques infraophiolitques dans le mélange, indique clairement que l'exhumation et l'érosion de reliefs composés d'ophiolites et de roches métamorphiques durant les stades finaux d'obduction, ont conduit à la sédimentation d'unités chaotiques sur la nappe ophiolitique du Llanvirnien tardif au Caradocien précoce (ca. 463-457 Ma).

En comparant nos données $^{40}Ar/^{39}Ar$ aux données géochronologiques régionales existantes, ainsi que celles pour les principaux massifs ophiolitiques qui affleurent le long de la ligne Baie Verte – Brompton, i.e les complexes de Thetford-Mines et de Mont-Albert, il est possible de reconstituer l'histoire géologique de leur mise en place. Suite à la cristallisation de la lithosphère océanique, l'obduction des ophiolites a débuté avec la formation des roches métamorphiques infraophiolitiques entre ca. 479 et 472 Ma, et s'est terminée vers 460-457 Ma, un âge qui correspond aux derniers stades de recristallisation et de déformation ductile dans la nappe ophiolitique et à la sédimentation de mélanges olistostromaux et d'unités flyschiques.

Toutefois, il a pu être démontré que l'obduction s'est effectuée de façon diachrone le long de la chaîne taconienne, avec une collision précoce enregistrée au sein du Mélange ophiolitique de Nadeau (avant 475 Ma) et une mise en place tardive des complexes ophiolitiques du sud du Québec et de celui du Mont-Albert. Ce diachronisme peut être expliqué par la géométrie irrégulière de la paléomarge laurentienne, avec une collision précoce en marge du promontoire du Saint-Laurent et la préservation de déformations plus tardives dans le cœur du réentrant de Québec.

Le quatrième chapitre de la thèse présente une synthèse de l'évolution tectono-stratigraphique de la chaîne taconienne dans le sud du Québec et le nord du Vermont, avec un accent particulier sur le Mélange de Saint-Daniel. Des données géochronologiques U-Pb et $^{40}Ar/^{39}Ar$ sur zircon et muscovite détritiques sont présentées afin de contraindre l'âge et la provenance de certaines unités, dont la séquence de la Colline Bunker, le Groupe de Magog dans la zone de Dunnage et la Formation de Compton dans la Ceinture de Gaspé.

Les unités volcaniques et sédimentaires précédemment interprétées comme des écailles tectoniques au sein du Mélange de Saint-Daniel sont revues selon le contexte établi dans les chapitres précédents. Les roches volcaniques du Groupe de Bolton et de la Volcanite de Ware forment des horizons stratigraphiques concordants dans le Mélange de Saint-Daniel. Contrairement à Tremblay (1990), la séquence de la colline Bunker n'est pas incluse dans le Mélange de Saint-Daniel. Cette unité est plutôt divisée en un membre sédimentaire et un membre volcanique qui ne font pas partie d'une même séquence stratigraphique. La provenance, l'âge, la composition et les relations stratigraphiques du membre sédimentaire avec le Mélange de Saint-Daniel suggèrent qu'il peut être corrélé avec les roches métasédimentaires de la zone de Humber et forme partie du substrat sur lequel a été déposé le Mélange de Saint-Daniel. Le membre volcanique possède plutôt des affinités avec les unités inférieures du Groupe de Magog, ce qui est concordant avec sa position stratigraphique et l'âge U-Pb de 455 ± 6 Ma qui a été obtenu pour ses roches volcanoclastiques. La corrélation des roches sédimentaires et volcaniques incluses dans le Mélange de Saint-Daniel du sud du Québec avec certaines unités stratigraphiques de la ceinture Rowe-Hawley du nord du Vermont, suggère que ces deux régions ont évolué selon une histoire géologique complémentaire. Il en ressort que le Mélange de Saint-Daniel représente la base d'un bassin avant-arc formé en bordure de la chaîne taconienne composée d'ophiolites obductées et de roches métamorphiques laurentiennes. La présence de coulées volcaniques dans le bassin (i.e. le Groupe de Bolton et la Volcanite de Ware) est interprétée comme étant due à un épisode de délamination de la plaque subductée, et ne résulte pas du volcanisme d'arc préservé au sein du Complexe d'Ascot.

Le Groupe de Magog repose en discordance sur le Mélange de Saint-Daniel. Sa stratigraphie, ainsi que l'âge des zircons et muscovites détritiques échantillonnées vers la base de la Formation de Saint-Victor, indiquent que la provenance des sédiments a évolué d'une source principalement composée de roches volcaniques et intrusives felsiques (i.e. un arc volcanique), à une source terrigène possiblement représentée par les nappes taconiennes de la zone Humber externe. La présence de zircons siluriens dans la partie sommitale de cette séquence turbiditique, indique toutefois que la sédimentation du Groupe de Magog n'est possiblement pas restreinte à l'Ordovicien supérieur, tel que précédemment proposé, mais se serait échelonnée sur une période de 25-35 Ma, jusqu'au Silurien.

Les provenances suggérées par les zircons et muscovites détritiques du membre de Milan (Formation de Compton) peuvent être entièrement réconciliées avec sa formation au sein d'un bassin sédimentaire post-collisionnel développé en marge de la chaîne taconienne. L'âge moyen des zircons les plus jeunes de cette unité est Emsien à Eifelien (396 ± 6 Ma), et indique que la sédimentation de la Formation de Compton a persisté jusqu'à la fin du Dévonien inférieur et possiblement jusqu'au début du Dévonien moyen.

CONTRIBUTIONS SCIENTIFIQUES FONDAMENTALES

Les travaux présentés dans cette thèse ont des impacts fondamentaux sur notre compréhension de certains aspects des chaînes d'obduction, notamment sur la durée et l'évolution du processus d'obduction et de la formation des mélanges supraophiolitiques.

L'obduction réfère généralement à l'ensemble des phénomènes géologiques qui conduisent à la mise en place d'une ophiolite sur une marge continentale, incluant : 1) la formation de lithosphère océanique et d'une semelle métamorphique infraophiolitique dans un environnement de subduction intra-océanique, 2) l'exhumation de la semelle métamorphique, 3) le transport de la nappe ophiolitique sur une marge continentale, et 4) l'érosion subaérienne du prisme orogénique ainsi formé (Wakabayashi et Dilek, 2003). L'obduction *sensu stricto* peu donc être considérée comme se terminant avec le soulèvement et l'érosion de la nappe ophiolitique obductée et des roches métamorphiques sous-jacentes, conduisant ainsi à la formation de bassins sédimentaires synorogéniques et de mélanges (Tremblay, Ruffet and Bédard, 2011). Dans ce contexte, les observations cartographiques et datations isotopiques compilées et présentées dans cette thèse mettent en évidence certains processus géologiques associés à l'obduction, notamment sur l'évolution de la nappe ophiolitique pendant son transport sur la marge continentale, ainsi que sur la formation des mélanges supraophiolitiques qui lui sont associés.

En plus des roches ophiolitiques crustales et mantelliques, les principales unités qui ont été étudiées pendant ce doctorat sont les granitoïdes d'obduction, les roches métamorphiques infraophiolitiques et certains bassins sédimentaires syncollisionnels. Les granitoïdes sont généralement interprétés comme le produit de la fusion partielle de sédiments déposés sur, ou en périphérie de la marge continentale chevauchée par l'ophiolite et font donc partie intérgrante du processus d'obduction de certaines ophiolites (Pearce, 1989; Whitehead, Dunning et Spray, 2000). En combinant les données isotopiques $^{40}Ar/^{39}Ar$ et les relations tectonostratigraphiques présentées dans cette thèse pour le Complexe ultramafique de la Rivière-des-Plante, aux données géochronologiques existantes (U-Pb et $^{40}Ar/^{39}Ar$) sur les granitoïdes d'obduction de l'ophiolite de Thetford-Mines, il est possible de reconstituer l'évolution tectonique et chronologique de ces roches. Nous pouvons conclure qu'il existe un écart de 3 à 5 Ma entre la cristallisation des granitoïdes à ca. 469-470 Ma (U-Pb; Whitehead, Dunning et Spray, 2000) et leur passage sous la température de fermeture de la muscovite à ca. 466-465 Ma ($^{40}Ar/^{39}Ar$; cette étude; Tremblay, Ruffet et Bédard, 2011), tandis que des âges $^{40}Ar/^{39}Ar$ sur muscovite pour des faciès déformés et non-déformés (cette étude; Tremblay, Ruffet et Bédard, 2011), suggèrent qu'ils ont été déformés et recristallisés jusqu'à ca. 461-460 Ma. Nous avons également pu démontrer la présence de fragments de granitoïde déformé et non-déformé dans les mélanges olistostromaux (conglomérat polygénique) qui reposent en discordance sur les roches ophiolitiques et granitiques, indiquant que la sédimentation des mélanges provient de l'érosion du substratum ophiolitique et que la déformation ductile de la nappe ophiolitique a précédé cette période d'érosion et de sédimentation.

Ces conclusions sont également corroborées par nos travaux sur les roches métamorphiques infraophiolitiques. Les âges $^{40}Ar/^{39}Ar$ sur muscovite et amphibole qui ont été obtenus sur les micaschistes et amphibolites immédiatement sous-jacents à la discordance marquant la base des mélanges olistostromaux dans la région de Lac-Brompton, sont de ca. 467 à 464 Ma et ca. 462 Ma, respectivement. Ces âges indiquent également que les roches métamorphique de la semelle infraophiolitique ont été déformées et recrystallisées jusqu'à ca. 462 Ma, et ensuite rapidement érodées et exhumées pour former les mélanges.

Ceci nous indique que, suite à la cristallisation des granitoïdes d'obduction, la nappe ophiolitique a été transportée sur la marge continentale pendant une période d'environ 10 Ma avant d'être érodée et exhumée au sein de la chaîne orogénique taconienne, pour ensuite être recyclée dans les mélanges olistostromaux qui caractérisent les bassins sédimentaires supraophiolitiques. La formation de ces bassins sédimentaires coincide avec les dernières évidences de recristallisation dans les roches de la semelle métamorphique et les granitoïdes d'obduction à ca. 457 Ma dans la région de Thetford-Mines (Tremblay, Ruffet et Bédard, 2011) et le transfert de la déformation vers l'avant pays, avec la mise en place des nappes externes jusqu'à ca. 450-445 Ma (St-Julien et Hubert, 1975; Tremblay et Castonguay, 2002; Sasseville et al., 2008). Les reconstitutions et corrélations lithotectoniques que nous avons proposées suggèrent également que la source des granitoïdes d'obduction est possiblement localisée à ca. 75 à 100 km au sud-est de leur position actuelle, dans la marge laurentienne déformée et métamorphisée représentée par le massif de Chain Lakes.

En somme, à partir de l'exemple appalachien étudié dans cette thèse, il est clair que la formation des mélanges et des bassins supraophiolitiques appartient intégralement au processus d'obduction, et est une conséquence directe de l'érosion du prisme orogénique formé durant les stades finaux de mise en place et de soulèvement de la nappe ophiolitique au sein d'un orogène.

ORIENTATION DES TRAVAUX FUTURS

Bien que les travaux présentés dans cette thèse ont mis en évidence certaines caractéristiques lithotectoniques de la zone de Dunnage des Appalaches du sud du Québec, ils soulignent également des lacunes quant à notre compréhension concernant certains aspects de la géologie appalachienne. Ces travaux peuvent donc servir de balises pour la réalisation de projets futurs.

Le premier article de la thèse met en évidence une corrélation possible entre les roches métamorphiques du massif de Chain Lakes et celles de la zone de Humber interne. Une série d'analyse $^{40}Ar/^{39}Ar$ sur les muscovites et amphiboles du massif de Chain Lakes, ainsi qu'une description détaillée des éléments structuraux qui le caractérisent, permettraient de compléter les données existantes sur ce massif, d'en déterminer l'évolution tectono-métamorphique au cours du Paléozoïque inférieur, ainsi que de la comparer avec celle de la zone de Humber du sud du Québec.

Les observations et données géochronologiques présentées dans le troisième chapitre mettent en évidence des disparités chronologique dans l'histoire d'obduction et de sédimentation syn-orogénique des régions du sud du Québec et de la Gaspésie. Il demeure toutefois à vérifier si ces disparités sont cohérentes avec les déformations et le métamorphisme préservés au sein du Groupe de Maquereau, en Gaspésie. Une étude structurale de ce groupe, ainsi qu'une analyse cinématique de la faille de Port-Daniel sont donc suggérées et devraient idéalement être combinées, si possible, à des travaux de datation $^{40}Ar/^{39}Ar$. La datation de minéraux détritiques (zircon et muscovite) des formations sédimentaires du Groupe de Mictaw et du Mélange de Port-Daniel permettrait également de déterminer les provenances sédimentaires de ces unités et de les comparer avec celles que nous avons documentées dans le sud du Québec.

Les données géochronologiques sur minéraux détriques qui sont présentées dans le dernier chapitre de la thèse suggèrent que la sédimentation du Groupe de Magog s'est échelonnée de l'Ordovicien supérieur au Silurien et que la provenance des divers faciès sédimentaires qui le composent ont varié significativement au cours du temps. Des levés stratigraphiques détaillés de section-type dans chacune des formations du Groupe de Magog et une analyse pétrographique des faciès gréseux et conglomératiques pourraient être combinée à un échantillonnage ciblé pour des fins de datations U-Pb sur zircons détritiques. Une telle étude permettrait de compléter et de synthétiser les données existantes sur l'âge et la provenance du Groupe de Magog, ainsi que de préciser l'évolution tectono-sédimentaire du bassin. À une échelle plus régionale, cette méthodologie pourrait également être appliquée à l'étude des formations flyschiques ordoviciennes à siluriennes des régions de Témiscouata et de la Gaspésie. Enfin, en ce qui concerne le Groupe de Magog, la présence de gabbro-diorite dans les formations de Frontière, Etchemin et Beauceville ne demeure que peu documentée. Il est proposé d'en faire la caractérisation géochimique et, si possible, de dater ces intrusions, soit par la méthode U-Pb sur zircon ou $^{40}Ar/^{39}Ar$ sur amphibole.

Nos travaux sur la Formation de Compton suggèrent que la sédimentation de cette unité aurait persisté jusqu'à la fin du Dévonien moyen. Or, cette période est marquée par un changement majeur dans le contexte tectonique du bassin de la Ceinture de Gaspé, d'un environnement extensif intra-continental au Silurien supérieur-Dévonien inférieur (Tremblay et Pinet, 2005) à un bassin d'avant-pays au front de l'orogénie acadienne au Dévonien inférieur à supérieur (Bradley et Tucker, 2002). Les mêmes types de travaux que ceux proposés pour le Groupe de Magog et visant à comprendre le contexte stratigraphique et sédimentologique de la Formation de Compton sont essentiels pour déterminer la paléogéographie du bassin.

ANNEXE A

CARTE GÉOLOGIQUE DU COMPLEXE ULTRAMAFIQUE DE LA RIVIÈRE-DES-PLANTE, RÉGION DE BEAUCEVILLE

En pochette

ANNEXE B

CARTE GÉOLOGIQUE DE LA RÉGION DU LAC MEMPHRÉMAGOG – SUD DU QUÉBEC ET NORD DU VERMONT

En pochette

ANNEXE C

CARTE GÉOLOGIQUE DE LA RÉGION DE FITCH BAY, LAC MEMPHRÉMAGOG

En pochette

ANNEXE D

DONNÉES LITHOGÉOCHIMIQUES MULTIÉLÉMENTAIRES POUR LES ROCHES GRANITIQUES DU COMPLEXE ULTRAMAFIQUE DE LA RIVIÈRE-DES-PLANTE

Données publiées dans l'article du Chapitre II

Les analyses ont été effectuées à l'Institut national de recherche scientifique Eau-Terre-Environnement, Québec. L'ensemble des éléments en trace, ainsi que le Sc, V, Zn, Cu, Cr, Ba et le Sr ont été analysés par ICP-AES, tandis que tous les autres éléments ont été analysés par ICP-MS. Tous les échantillons viennent du Complexe ultramafique de la Rivière-des-Plante, sauf les échantillons 07M170 et 07M171, qui ont été prélevé au sein de l'ophiolite de Thetford-Mines. TMO – Thetford-Mines ophiolite; (M) – granite à texture mylonitique; n.d. – sous la limite de détection. *Fer total exprimé sous la forme Fe_2O_3. Localisation des échantillons en coordonnées UTM-NAD 1983 – ZONE 18.

Granitoïd Sample	Type 1 08M31B E 821615 N 5129654	Type 1 08M70 E 821342 N 5128743	Type 2 07M24 E 824354 N 5131065	Type 2 07M42 E 823713 N 5131405	Type 2 07M58 E 824652 N 5133009	Type 2 07M84 E 826092 N 5136228
SiO_2 wt%	70.49	71.12	66.01	70.47	69.79	68.03
TiO_2	0.32	0.37	0.9	0.8	0.66	0.82
Al_2O_3	14.26	14.43	14.2	13.34	13.61	15.33
Fe_2O_3*	2.54	2.72	5.02	5.17	3.95	5.59
MnO	0.07	0.05	0.05	0.1	0.09	0.07
MgO	1.03	1.83	1.31	1.29	1.6	1.13
CaO	0.86	0.93	0.91	0.89	0.82	1.05
Na_2O	2.6	4.24	2.05	1.65	2.18	1.46
K_2O	6.39	3.42	5.58	3.19	4.78	3.89
P_2O_5	0.29	0.21	0.18	0.08	0.13	0.14
Total	98.83	99.32	96.21	96.97	97.61	97.51
Rb ppm	166	65	160	106	111	139
Ba	832	888	2 161	758	1 393	997
Sr	210	234	245	182	290	153
Th	8.57	6.34	8.9	11.68	8.26	9.9
Nb	14.71	14.49	15.12	11.97	12.18	12.72
Ta	1.06	1.25	0.99	0.78	0.85	0.84
Zr	167	154	355	339	256	302
Hf	4.97	4.53	9.7	9.55	7.43	8.61
Y	30.45	21.27	32.52	31.05	25.88	26.53
La	35.31	26.4	40.19	42.2	33.73	42.96
Ce	75.15	56.07	82.38	86.77	68.46	86.09
Pr	9.15	6.83	9.51	9.49	7.87	9.7
Nd	33.82	25.45	35.94	35.06	31.35	36.94
Sm	7.02	5.32	6.49	6.07	5.29	6.25
Eu	1.26	1.04	1.29	1.33	1.32	1.41
Gd	6.08	4.43	5.43	4.81	4.3	5.07
Tb	0.98	0.69	0.85	0.76	0.69	0.77
Dy	4.09	5.98	5.24	4.91	4.37	4.61
Ho	0.75	1.12	1.03	0.99	0.84	0.87
Er	1.98	2.97	3.03	2.99	2.43	2.5
Tm	0.28	0.42	0.43	0.42	0.35	0.35
Yb	1.92	2.88	2.88	2.68	2.31	2.27
Lu	0.28	0.41	0.44	0.41	0.35	0.35
Cr	174	142	154	180	149	175
Cu	n.d.	n.d.	32.3	19.5	17.7	22.6
Zn	41.1	31.1	62.3	84.8	44.4	86.1
V	18,00	27.4	63.6	65.4	46.3	64.1
Sc	6.5	6.44	9.72	11.8	7.82	11.23

Granitoïd Sample	Type 2 07M144 E 828489 N 5136290	Type 2 (M) 07M25 E 823588 N 5132129	TMO 07M170 E 781623 N 5104558	TMO 07M171 E 786572 N 5107664
SiO_2 wt%	75.02	67.11	71.27	72.27
TiO_2	0.77	0.93	0.33	0.27
Al_2O_3	11.52	14.55	14.21	14.14
Fe_2O_3*	3.9	5.65	2.5	2.03
MnO	0.03	0.08	0.06	0.05
MgO	0.9	2.02	0.87	0.8
CaO	0.25	1.38	1.3	0.87
Na_2O	1.14	2.67	2.55	4.23
K_2O	3.9	2.62	4.82	2.99
P_2O_5	0.09	0.18	0.23	0.22
Total	97.52	97.18	98.14	97.86
Rb ppm	117	65	180	122
Ba	1002	947	711	523
Sr	130	255	184	190
Th	11.15	9.92	6.67	6
Nb	11.7	13.89	14.09	14.91
Ta	0.73	0.9	1.03	1.45
Zr	480	286	144	133
Hf	13.18	8.05	4.36	4.16
Y	18.77	30.48	28.63	23.38
La	33.7	42.17	28.99	24.24
Ce	70.17	85.33	59.65	48.92
Pr	7.7	9.76	6.96	5.52
Nd	28.73	36.5	26.11	20.25
Sm	4.71	6.53	5.18	3.9
Eu	0.99	1.58	1.13	0.84
Gd	3.57	5.39	4.54	3.39
Tb	0.54	0.83	0.75	0.58
Dy	3.2	5.06	4.69	3.64
Ho	0.63	0.97	0.89	0.72
Er	1.82	2.86	2.5	2.08
Tm	0.27	0.4	0.36	0.31
Yb	1.79	2.61	2.27	2.02
Lu	0.29	0.4	0.34	0.3
Cr	179	129	110	137
Cu	15.4	26.3	14.4	11.8
Zn	53.4	69.6	44.1	25.8
V	47.9	74.2	19.7	n.d.
Sc	9.08	10.79	6.29	5.2

ANNEXE E

DONNÉES LITHOGÉOCHIMIQUES MULTIÉLÉMENTAIRES NON-PUBLIÉES

L'ensemble des éléments en trace, ainsi que le Sc, V, Zn, Cu, Cr, Ba et le Sr ont été analysés par ICP-AES, tandis que tous les autres éléments ont été analysés par ICP-MS. VW – Volcanite de Ware; GB - roches volcaniques et intrusives du Groupe de Bolton; GB? – gabbro recoupant le membre sédimentaire de la séquence de la Colline Bunker; SCBV – membre volcanique de la séquence de la Colline Bunker; SCBS membre sédimentaire de la séquence de la Colline Bunker; GC – Groupe de Caldwell; FF – Formation de Frontière; MSD – Mélange de Saint-Daniel; n.d. – sous la limite de détection. *Fer total exprimé sous la forme Fe_2O_3.

%	07M114A	07M117	07M120	07M133	07M134	07M143	07M159
	E 850180	E 850489	E 853864	E 857093	E 857017	E 853374	E 852417
	N 5156613	N 5156378	N 5161036	N 5164498	N 5164666	N 5160180	N 5158572
	VW	VW	VW	VW	VW	VW	VW
	bréchique	bréchique	massif	Massif, quartz	Massif, vésiculaire	Massif, quartz	massif, perlitique
SiO_2	64.83	70.90	73.32	67.71	71.62	65.45	67.69
TiO_2	0.62	0.47	0.45	0.65	0.50	0.50	0.63
Al_2O_3	14.29	12.89	12.17	13.67	12.53	12.47	14.53
Fe_2O_3*	5.61	2.69	2.97	4.05	3.20	5.54	3.62
MgO	3.42	1.81	1.32	2.65	1.25	4.27	1.47
CaO	0.67	0.20	0.65	0.59	2.70	3.13	0.83
Na_2O	2.88	0.16	4.59	3.13	2.09	1.33	4.92
K_2O	2.52	7.96	1.95	3.29	2.82	3.40	3.07
MnO	0.31	0.08	0.05	0.08	0.04	0.08	0.08
P_2O_5	0.12	0.11	0.11	0.15	0.15	0.12	0.13
ppm							
Rb	124.40	200.71	64.89	98.73	101.32	148.09	73.54
Ba	274.20	1181.70	848.40	894.80	611.70	701.80	1005.00
Pb	13.04	< 20	36.80	20.28	20.05	27.11	22.09
Sr	79.60	47.10	138.60	62.00	93.50	398.90	180.50
Th	7.00	8.39	6.48	7.47	7.35	5.98	8.69
U	1.19	2.13	1.35	1.62	1.62	1.18	1.83
Nb	11.08	12.62	9.60	11.51	10.99	9.17	12.76
Ta	0.74	0.91	0.69	0.78	0.78	0.63	0.91
Zr	228.69	259.40	202.51	251.69	214.36	190.58	270.46
Hf	6.48	7.33	5.70	7.02	6.08	5.37	7.72
Y	26.19	28.18	21.84	23.59	25.44	24.60	28.65
La	34.184	34.595	19.604	15.275	24.377	27.027	21.802
Ce	70.831	70.516	41.994	41.961	55.636	55.216	55.965
Pr	8.551	8.280	4.934	6.183	6.795	6.431	7.083
Nd	31.835	31.119	19.249	26.138	26.295	24.158	28.479
Sm	5.391	5.756	3.870	5.102	5.010	4.441	5.183
Eu	1.504	1.096	0.679	1.144	1.016	0.915	0.999
Gd	4.329	4.749	3.404	4.168	4.280	3.826	4.464
Tb	0.678	0.759	0.545	0.641	0.674	0.613	0.731
Dy	4.187	4.657	3.462	3.875	4.153	3.845	4.697
Ho	0.832	0.916	0.700	0.773	0.822	0.756	0.929
Er	2.448	2.718	2.076	2.280	2.380	2.254	2.765
Tm	0.347	0.397	0.307	0.338	0.350	0.322	0.405
Yb	2.247	2.550	2.022	2.262	2.271	2.132	2.668
Lu	0.342	0.395	0.310	0.351	0.346	0.332	0.415
Cr	214	104	108	175	96	341	187
Ni	96	45	n.d.	n.d.	n.d.	79	n.d.
Co	17	n.d.	n.d.	n.d.	n.d.	24	n.d.
Cu	29	34	n.d.	16	13	38	35
Zn	60	103	23	64	46	54	40
V	94	38	63	85	38	106	82
Sc	17	12	11	14	10	20	19
total	98.31	99.33	98.7	98.32	98.89	98.98	99.06

%	07B32 E 709190 N 5004137 GB dyke, intermédiaire	07B60 E 710519 N 5007300 GB dyke grenu, mafique	07B97 E 709125 N 5001203 GB dyke grenu, intermédiaire	07B106 E 710126 N 4994180 GB dyke, quartz	07B210 E 718174 N 4995029 GB coussiné, schiste vert	07B211 E 718016 N 4995020 GB coussiné	07B217 E 719516 N 4997346 GB coussiné, schiste vert
SiO_2	51.75	48.42	56.77	46.56	49.15	47.41	57.22
TiO_2	1.15	1.24	1.55	0.76	0.97	1.29	1.06
Al_2O_3	14.24	14.18	14.14	14.42	14.92	15.50	14.53
Fe_2O_3*	9.60	12.06	10.41	8.50	9.04	10.52	10.17
MgO	5.10	6.68	2.48	6.53	6.17	6.82	3.71
CaO	5.44	5.54	4.75	9.15	6.60	11.40	5.59
Na_2O	2.06	2.33	4.61	3.14	4.45	3.02	4.14
K_2O	0.47	n.d.	0.20	0.04	n.d.	n.d.	0.60
MnO	0.19	0.18	0.15	0.18	0.18	0.17	0.17
P_2O_5	0.12	0.12	0.19	0.05	0.09	0.11	0.13
ppm							
Rb	26.99	2.44	13.23	7.19	0.67	0.39	18.55
Ba	77.60	17.70	163.50	61.50	8.80	40.60	125.90
Pb	45.04	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sr	71.10	155.50	267.90	242.70	119.60	238.00	92.80
Th	4.00	1.92	4.75	1.13	3.17	0.99	4.07
U	0.85	0.44	1.14	0.28	0.62	0.24	0.94
Nb	7.08	4.14	8.63	2.25	4.80	2.89	5.79
Ta	0.51	0.30	0.65	0.16	0.36	0.21	0.43
Zr	115.96	78.02	151.73	42.34	95.90	77.42	95.25
Hf	3.50	2.45	4.62	1.40	3.02	2.45	3.05
Y	34.95	30.25	47.69	19.59	32.53	32.99	30.35
La	16.296	8.457	18.205	5.161	11.801	5.692	15.478
Ce	34.454	18.919	39.529	11.353	25.456	13.954	32.161
Pr	4.196	2.511	4.914	1.481	3.224	2.037	3.844
Nd	17.529	11.394	20.991	6.737	13.762	10.138	15.809
Sm	4.174	3.104	5.308	1.872	3.551	3.111	3.712
Eu	1.120	1.046	1.522	0.643	1.026	1.144	1.033
Gd	4.542	3.836	6.078	2.381	4.156	4.103	4.023
Tb	0.799	0.695	1.072	0.434	0.747	0.738	0.696
Dy	5.350	4.713	7.291	2.984	4.977	5.073	4.684
Ho	1.113	0.988	1.519	0.636	1.044	1.075	0.981
Er	3.367	3.008	4.615	1.914	3.143	3.260	2.930
Tm	0.485	0.428	0.671	0.272	0.453	0.457	0.425
Yb	3.174	2.829	4.428	1.796	2.945	3.013	2.778
Lu	0.486	0.434	0.682	0.273	0.447	0.454	0.430
Cr	145	153	52	155	240	287	72
Ni	n.d.	n.d.	n.d.	58	79	91	n.d.
Co	27	39	23	21	39	40	29
Cu	21	72	12	21	68	34	51
Zn	58	84	77	58	70	72	84
V	239	312	239	228	213	272	283
Sc	33	43	27	37	35	39	33
total	98.08	97.95	100.72	99.1	98.81	98.63	99.34

%	08B157 E 725261 N 5003551 GB? métagabbro	08B161 E 725310 N 5003357 GB? métagabbro	07B134 E 729887 N 5008198 ? dyke carbonatisé	07B227 E 716224 N 4997837 SCBV grès v.clastique	09B548 E 719040 N 4999325 SCBV grès v.clastique	09BUNK E 729201 N 5007658 SCBV grès v.clastique	08B60 E 718415 N 4998350 SCBV grès v.clastique
SiO_2	60.2	46.3	50.2	67.46	60.36	60.96	73.22
TiO_2	.957	2.33	2.01	0.44	0.40	0.53	0.49
Al_2O_3	14.3	17.7	13.6	10.80	14.79	12.22	11.98
Fe_2O_3*	6.38	14.1	10.3	3.46	6.10	5.93	4.55
MgO	2.67	8.31	2.55	1.79	2.66	2.68	0.57
CaO	3.86	4.06	5.88	4.80	2.35	2.96	0.43
Na_2O	4.86	3.16	3.85	2.76	4.73	2.18	4.16
K_2O	.833	n.d.	1.20	1.86	1.71	2.76	1.45
MnO	.105	.178	.190	0.29	0.14	0.46	0.13
P_2O_5	.146	.266	.606	0.20	0.06	0.14	0.08
ppm							
Rb	31.6	3.8	40.5	58.20	58.36	118.46	44.70
Ba	100.	69.0	839.	592.27	591.92	701.23	1006.10
Pb	n.d.	n.d.	n.d.	19.56	5.49	9.49	n.d.
Sr	225.	237.	383.	416.89	192.46	197.27	81.80
Th	6.0	2.6	3.7	4.23	2.65	3.60	4.44
U	1.8	.3	1.5	1.62	1.03	1.45	1.15
Nb	7.8	7.6	10.1	5.62	3.14	5.57	5.63
Ta	.7	.5	0.84	0.45	0.28	0.43	0.33
Zr	158.1	182.5	383.6	112.30	75.11	110.70	131.91
Hf	4.7	5.6	9.41	2.71	2.04	2.47	3.92
Y	32.5	45.0	69.2	23.01	15.02	21.98	18.39
La	20.4	18.8	24.4	17.778	12.705	23.562	18.135
Ce	43.0	39.8	61.4	36.506	24.856	46.340	36.909
Pr	5.4	5.6	8.4	4.399	2.897	5.162	4.455
Nd	21.2	24.8	38.1	17.908	11.427	19.572	16.896
Sm	5.0	6.7	9.4	3.681	2.593	3.809	3.546
Eu	1.3	2.0	2.5	0.734	0.547	0.923	0.841
Gd	5.3	7.9	10.6	3.353	1.768	2.958	3.338
Tb	.9	1.3	1.75	0.583	0.346	0.562	0.524
Dy	5.8	8.6	11.2	3.613	2.781	3.849	3.252
Ho	1.2	1.7	2.27	0.732	0.579	0.728	0.653
Er	3.5	5.2	6.66	2.236	1.757	2.177	1.907
Tm	.5	.7	.93	0.289	0.262	0.345	0.280
Yb	3.4	4.5	5.88	2.085	1.807	2.108	2.063
Lu	.5	.7	0.88	0.342	0.285	0.331	0.328
Cr	184.	292.	39.8	234	127	255	187
Ni	n.d.	88.9	n.d.	35	33	81	n.d.
Co	20.4	50.7	28.5	10	17	19	n.d.
Cu	17.2	n.d.	29.7	10	19	26	n.d.
Zn	68.1	146.	103.	58	68	55	65
V	131.	301.	157.	81	110	105	65
Sc	20.4	48.3	20.4	13	20	18	14
total	100.56	101.64	99.73	101.57	101.32	99.13	99.6

%	08B150 E 716083 N 4997576 SCBV grès v.clastique	08B12 E 717521 N 4999325 SCBV tuf gris	08B10 E 717263 N 4999448 SCBV tuf gris	08B11 E 717453 N 4999390 SCBV tuf gris	08B80 E 716039 N 4997366 SCBV tuf gris	09BUNK55 E 729809 N 5007804 SCBV tuf gris/grès	08B83-1 E 716319 N 4997342 SCBS grès
SiO_2	76.99	77.22	79.22	77.49	75.25	59.88	74.87
TiO_2	0.36	0.16	0.19	0.29	0.18	0.66	0.89
Al_2O_3	11.95	12.49	7.05	10.67	11.59	16.32	11.12
Fe_2O_3*	3.23	1.42	4.11	4.11	1.89	8.54	4.43
MgO	0.42	0.36	0.54	0.37	0.78	5.57	0.44
CaO	0.04	0.05	0.29	0.08	1.09	0.37	1.33
Na_2O	2.90	4.59	2.99	3.06	5.23	2.87	3.08
K_2O	1.77	1.63	0.39	1.36	1.27	1.49	2.34
MnO	0.10	0.04	0.20	0.16	0.05	0.12	0.09
P_2O_5	n.d.	n.d.	0.04	0.05	n.d.	0.14	0.16
ppm							
Rb	37.17	47.08	16.62	36.21	30.68	50.81	69.59
Ba	453.20	483.70	195.70	405.90	649.70	4628.84	1333.50
Pb	n.d.	n.d.	12.81	n.d.	n.d.	19.69	13.29
Sr	102.60	45.10	68.70	49.80	183.70	81.64	116.00
Th	3.61	8.01	3.35	3.36	7.13	4.92	7.75
U	0.73	1.81	0.73	0.66	1.66	1.51	1.42
Nb	3.20	5.54	2.97	2.97	4.26	6.36	15.57
Ta	0.20	0.41	0.19	0.19	0.31	0.38	0.86
Zr	98.45	106.78	70.14	88.69	96.33	155.02	456.31
Hf	3.21	4.23	2.24	2.83	3.80	3.48	12.25
Y	18.54	50.20	17.93	24.79	43.23	27.53	21.94
La	12.753	28.156	12.315	13.113	23.896	25.764	34.047
Ce	26.654	58.993	26.392	27.789	52.139	48.709	71.604
Pr	3.286	7.423	3.181	3.458	6.308	5.624	8.586
Nd	12.488	28.697	12.434	13.777	24.044	21.906	32.177
Sm	2.654	6.651	2.892	3.295	5.536	4.350	5.986
Eu	0.643	0.606	0.613	0.743	0.551	0.996	1.309
Gd	2.702	6.822	2.965	3.701	5.900	3.859	4.792
Tb	0.461	1.172	0.482	0.626	1.045	0.652	0.714
Dy	3.123	8.087	3.204	4.097	7.119	4.520	4.241
Ho	0.687	1.754	0.667	0.864	1.527	0.868	0.818
Er	2.159	5.241	2.053	2.598	4.512	2.712	2.305
Tm	0.345	0.801	0.320	0.382	0.701	0.402	0.337
Yb	2.584	5.861	2.432	2.740	5.182	2.491	2.415
Lu	0.411	0.907	0.390	0.426	0.803	0.387	0.377
Cr	148	67	188	151	89	119	214
Ni	n.d.	n.d.	n.d.	n.d.	n.d.	27	n.d.
Co	n.d.	n.d.	n.d.	n.d.	n.d.	25	n.d.
Cu	n.d.	n.d.	n.d.	n.d.	n.d.	3	n.d.
Zn	46	22	36	48	35	22	50
V	44	n.d.	n.d.	39	n.d.	182	56
Sc	12	7	8	12	8	29	7
total	99.63	99.16	97.71	99.62	99.81	100.5	101.25

	07B189	07B190	09B454	09B512	09B537	09B654	caldwell 1
	E 715953	E 718190	E 719240	E 724369	E 715771	E 727450	E 823315
	N 4993925	N 4995910	N 4997419	N 5001478	N 4995238	N 5005280	N 5133679
	SCBS	SCBS	SCBS	SCBS	SCBS	SCBS	GC
%	grès	grès	grès	grès	grès	grès	grès
SiO_2	83.54	74.90	67.77	76.45	80.31	69.00	75.03
TiO_2	0.50	0.54	1.00	0.72	0.75	1.00	0.73
Al_2O_3	6.50	9.40	16.19	12.21	10.26	15.10	10.91
Fe_2O_3*	3.27	8.26	5.34	4.05	2.93	5.17	3.92
MgO	0.53	1.72	1.28	0.90	0.60	1.24	0.74
CaO	0.11	0.28	0.33	0.39	0.41	0.28	1.23
Na_2O	1.17	1.24	2.57	1.68	1.84	2.53	1.66
K_2O	0.70	1.09	4.20	2.35	2.56	3.22	2.73
MnO	0.03	0.06	0.03	0.05	0.05	0.06	0.04
P_2O_5	0.07	< 0.038	0.24	0.14	0.16	0.19	0.08
ppm							
Rb	32.49	38.29	150.40	104.08	91.09	146.85	83.89
Ba	223.50	322.70	1450.07	560.20	665.67	799.58	730.50
Pb	12.98	n.d.	11.49	27.88	6.26	11.89	n.d.
Sr	36.60	63.50	110.36	69.47	100.00	75.74	95.80
Th	6.85	7.98	8.28	8.74	5.52	9.25	9.63
U	0.76	1.44	1.89	1.62	1.21	1.41	1.52
Nb	7.75	7.52	18.46	12.26	12.47	17.22	11.37
Ta	0.46	0.48	0.82	0.62	0.59	0.72	0.69
Zr	296.58	312.63	413.47	436.31	411.23	461.51	378.62
Hf	8.18	8.74	9.73	10.28	8.68	10.61	10.66
Y	11.04	15.85	29.53	19.21	21.18	30.37	17.83
La	16.658	22.617	46.935	30.739	27.612	41.069	32.555
Ce	31.428	45.780	96.298	66.516	57.614	83.612	66.423
Pr	3.793	5.089	10.998	7.284	6.969	9.946	7.524
Nd	13.804	18.395	43.403	28.720	24.881	39.236	27.155
Sm	2.621	3.409	8.121	5.520	5.010	7.337	4.918
Eu	0.553	0.751	1.601	1.106	0.894	1.416	1.106
Gd	2.188	2.824	5.621	3.759	3.818	5.336	3.852
Tb	0.321	0.453	0.921	0.599	0.622	0.886	0.576
Dy	1.910	2.819	5.734	3.526	3.790	5.634	3.395
Ho	0.371	0.564	1.090	0.621	0.681	1.087	0.646
Er	1.089	1.669	3.198	1.876	2.051	3.151	1.851
Tm	0.163	0.244	0.435	0.254	0.279	0.428	0.269
Yb	1.146	1.708	2.762	1.781	1.869	2.945	1.837
Lu	0.188	0.260	0.422	0.280	0.288	0.395	0.286
Cr	214	155	113	165	169	132	281
Ni	n.d.	n.d.	22	33	9	21	n.d.
Co	n.d.	n.d.	15	8	10	12	n.d.
Cu	n.d.	49	18	17	6	13	10
Zn	30	112	72	85	34	84	48
V	29	38	74	49	43	62	50
Sc	3	5	11	7	6	10	7
total	97.43	99.94	101.168	101.09	102.21	99.94	99.52

	caldwell 2	09FR2	09FR3	09FR4	08B256	08B83-2
	E 823409	E 849413	E 849839	E 853916	E 717592	E 716319
	N 5133070	N 5146908	N 5146510	N 5153586	N 4999271	N 4997342
	GC	FF	FF	FF	MSD	MSD
%	grès	grès	grès	grès	grès noir	grès noir
SiO_2	75.15	69.16	66.87	68.84	74.21	67.68
TiO_2	0.75	0.57	0.89	0.65	0.72	0.68
Al_2O_3	12.50	12.34	12.41	14.09	10.10	15.16
Fe_2O_3*	4.21	5.00	6.79	5.76	3.36	6.49
MgO	0.96	2.53	3.58	3.03	0.72	1.76
CaO	0.37	2.21	1.34	1.31	1.70	0.14
Na_2O	2.26	2.49	2.50	2.88	2.40	1.14
K_2O	2.81	1.86	1.48	2.04	1.57	3.99
MnO	0.04	0.07	0.07	0.07	0.07	0.09
P_2O_5	0.09	0.11	0.15	0.13	0.08	0.11
ppm						
Rb	89.67	80.94	66.91	90.65	47.42	116.43
Ba	797.80	432.55	372.86	454.64	269.40	734.80
Pb	14.16	25.34	31.24	25.44	12.77	18.14
Sr	123.00	186.90	138.58	193.56	139.10	74.90
Th	9.32	7.29	9.64	8.44	7.91	11.57
U	1.49	1.69	2.24	1.82	1.92	2.12
Nb	12.32	9.69	13.88	9.80	11.48	12.22
Ta	0.74	0.46	0.58	0.48	0.68	0.78
Zr	419.52	262.25	422.21	231.03	390.57	271.19
Hf	11.34	5.77	8.86	5.53	10.75	7.81
Y	19.31	21.68	30.43	20.77	15.19	24.48
La	31.930	29.888	40.016	27.151	26.999	42.233
Ce	67.150	54.154	77.839	55.095	54.455	82.510
Pr	7.678	6.792	9.075	6.463	6.039	9.575
Nd	27.994	24.682	33.880	24.347	21.578	34.496
Sm	5.108	4.665	6.229	4.367	3.827	6.402
Eu	1.072	1.042	1.363	0.965	0.828	1.175
Gd	3.973	3.834	5.203	3.319	2.969	5.030
Tb	0.589	0.611	0.886	0.580	0.445	0.756
Dy	3.556	3.560	5.019	3.611	2.644	4.641
Ho	0.684	0.679	1.029	0.702	0.515	0.900
Er	1.946	1.999	2.983	2.206	1.513	2.621
Tm	0.291	0.316	0.427	0.317	0.227	0.377
Yb	2.068	1.926	2.783	2.021	1.612	2.545
Lu	0.326	0.308	0.408	0.322	0.259	0.387
Cr	200	250	576	286	230	168
Ni	n.d.	61	88	64	n.d.	n.d.
Co	n.d.	14	24	13	n.d.	n.d.
Cu	n.d.	16	26	19	19	37
Zn	56	73	93	81	58	76
V	54	94	125	107	42	70
Sc	8	14	17	15	6	12
total	101.28	99.75	99.7	101.37	99.2	100.43

ANNEXE F

DONNÉES GÉOCHRONOLOGIQUES U-Pb (LA-ICP-MS)

Les analyses ont été effectuées à l'Université d'Alberta, Radiogenic isotope facility, Edmonton.
La localisation des échantillons est donnée en coordonnées UTM-NAD 1983 – ZONE 18.

Ware Volcanics E 854325

Sample 08M44 N 5161625

Grain #	^{206}Pb (cps)	^{204}Pb (cps)	$^{207}Pb/^{206}Pb$	2 σ	$^{207}Pb/^{235}U$	2 σ	$^{206}Pb/^{238}U$	2 σ	age (Ma) $^{207}Pb/^{206}Pb$	2 σ	$^{206}Pb/^{238}U$	2 σ	disc.
2	52496	97	0.07602	0.00092	2.10752	0.09427	0.20107	0.00866	1096	12	1181	46	-8.5
6	38004	121	0.05775	0.00355	0.56136	0.05094	0.07049	0.00471	520	67	439	28	16.2
8	58231	111	0.05407	0.00188	0.52722	0.02759	0.07072	0.00277	374	39	440	17	-18.4
9	52264	77	0.06737	0.00154	1.37250	0.08339	0.14775	0.00832	849	24	888	47	-4.9
11	44126	86	0.06521	0.00134	1.37901	0.08238	0.15338	0.00860	781	22	920	48	-19.1
12	58220	92	0.06818	0.00086	1.44336	0.05385	0.15353	0.00539	874	13	921	30	-5.7
13	130259	98	0.07171	0.00061	1.51105	0.06377	0.15282	0.00632	978	9	917	35	6.7
16	34261	97	0.06349	0.00185	1.33099	0.07306	0.15204	0.00707	725	31	912	39	-27.8
17	82019	119	0.07112	0.00094	1.49510	0.09281	0.15247	0.00925	961	14	915	52	5.1
24	93815	28	0.06772	0.00063	1.27177	0.05898	0.13621	0.00619	860	10	823	35	4.6
29	109446	43	0.08977	0.00136	2.60019	0.12855	0.21007	0.00988	1421	14	1229	52	14.8
32	135576	49	0.09025	0.00075	2.58533	0.16458	0.20777	0.01311	1431	8	1217	70	16.4
33	142949	44	0.09879	0.00075	3.41858	0.14896	0.25096	0.01077	1601	7	1443	55	11.0
34	201780	85	0.07253	0.00057	1.48536	0.06436	0.14853	0.00633	1001	8	893	35	11.6
35	45717	76	0.07414	0.00220	1.97235	0.11170	0.19295	0.00931	1045	30	1137	50	-9.6
36	71036	83	0.06796	0.00105	1.39619	0.07878	0.14900	0.00809	867	16	895	45	-3.5
38	125785	81	0.07149	0.00051	1.44714	0.05538	0.14680	0.00552	972	7	883	31	9.8
39	67221	54	0.07556	0.00091	1.86283	0.08306	0.17880	0.00768	1084	12	1060	42	2.3
40	84942	92	0.06847	0.00115	1.44885	0.05852	0.15346	0.00564	883	17	920	31	-4.6
41	90181	90	0.06715	0.00070	1.37228	0.05353	0.14822	0.00557	842	11	891	31	-6.2
42	116498	109	0.06855	0.00078	1.35895	0.07208	0.14377	0.00745	885	12	866	42	2.3
43	38461	139	0.06320	0.00279	1.31451	0.08670	0.15084	0.00739	715	47	906	41	-28.6
48	63186	117	0.06865	0.00094	1.52888	0.06735	0.16152	0.00676	888	14	965	37	-9.3
51	31817	155	0.08627	0.00162	3.23504	0.14649	0.27198	0.01121	1344	18	1551	57	-17.3
52	144373	152	0.05723	0.00052	0.72321	0.03592	0.09165	0.00448	500	10	565	26	-13.5

Bunker Hill sequence - sedimentary member E 726 682
Sample 07BUNKER N 5003869

Grain #	^{206}Pb (cps)	^{204}Pb (cps)	$^{207}Pb/^{206}Pb$	2 σ	$^{207}Pb/^{235}U$	2 σ	$^{206}Pb/^{238}U$	2 σ	age (Ma) $^{207}Pb/^{206}Pb$	2 σ	$^{206}Pb/^{238}U$	2 σ	disc.
2	99704	24	0.06628	0.00267	1.58618	0.09892	0.17357	0.00827	815	42	1032	45	-28.8
3	124729	22	0.06864	0.00278	1.74466	0.10610	0.18434	0.00836	888	42	1091	45	-24.8
4	322275	34	0.07302	0.00287	1.94204	0.11962	0.19289	0.00914	1014	40	1137	49	-13.2
9	109696	16	0.06949	0.00286	1.74434	0.12658	0.18206	0.01088	913	42	1078	59	-19.6
10	206360	14	0.07046	0.00281	1.70463	0.10168	0.17546	0.00779	942	41	1042	43	-11.5
11	68189	111	0.07283	0.00293	1.87212	0.12260	0.18642	0.00963	1009	41	1102	52	-10.0
12	358057	92	0.10026	0.00393	4.03051	0.25775	0.29155	0.01473	1629	36	1649	73	-1.4
13	98418	69	0.07196	0.00284	1.72040	0.09766	0.17340	0.00707	985	40	1031	39	-5.1
15	122013	83	0.07406	0.00292	1.88111	0.10817	0.18422	0.00771	1043	40	1090	42	-4.9
16	135728	60	0.07294	0.00287	1.78741	0.10790	0.17772	0.00813	1012	40	1055	44	-4.5
17	22703	17	0.07252	0.00296	1.94373	0.12269	0.19440	0.00936	1001	41	1145	50	-15.8
18	245700	39	0.07431	0.00292	1.84225	0.10581	0.17980	0.00752	1050	40	1066	41	-1.7
19	22072	31	0.09548	0.00387	3.81096	0.26479	0.28950	0.01634	1537	38	1639	81	-7.5
20	194897	37	0.07504	0.00295	1.89670	0.11206	0.18331	0.00808	1070	40	1085	44	-1.6
20	30189	113	0.06975	0.00287	1.72318	0.10466	0.17919	0.00801	921	42	1063	44	-16.7
22	48907	103	0.06799	0.00272	1.47054	0.08851	0.15686	0.00706	868	41	939	39	-8.8
23	74877	104	0.05751	0.00230	0.76517	0.04898	0.09649	0.00482	511	44	594	28	-16.9
24	495141	117	0.09329	0.00366	3.30010	0.20918	0.25657	0.01278	1494	37	1472	65	1.6
25	103911	104	0.08039	0.00318	2.32639	0.14859	0.20988	0.01053	1207	39	1228	56	-2.0
26	137849	94	0.07169	0.00283	1.65517	0.10418	0.16744	0.00821	977	40	998	45	-2.3
27	50101	72	0.07077	0.00289	1.73268	0.10163	0.17756	0.00748	951	42	1054	41	-11.7
28	186909	92	0.07118	0.00280	1.58577	0.09903	0.16158	0.00783	963	40	966	43	-0.3
29	133576	88	0.07360	0.00290	1.87156	0.11059	0.18442	0.00811	1031	40	1091	44	-6.4
30	187226	94	0.07012	0.00277	1.58599	0.08991	0.16405	0.00668	932	40	979	37	-5.5
31	149808	107	0.09931	0.00391	3.87399	0.23034	0.28291	0.01261	1611	37	1606	63	0.4
32	32004	106	0.07107	0.00297	1.70976	0.10787	0.17447	0.00824	960	43	1037	45	-8.7
33	38176	86	0.06691	0.00281	1.46343	0.08583	0.15863	0.00650	835	44	949	36	-14.7
34	1563568	81	0.09483	0.00372	3.49485	0.20725	0.26728	0.01190	1525	37	1527	60	-0.2
37	51508	80	0.07234	0.00288	1.79266	0.10294	0.17973	0.00744	995	40	1066	41	-7.6
38	784114	91	0.10176	0.00399	4.08197	0.24738	0.29094	0.01345	1656	36	1646	67	0.7
39	401423	79	0.07290	0.00286	1.71216	0.10009	0.17034	0.00737	1011	40	1014	40	-0.3
40	538790	86	0.07262	0.00285	1.66712	0.09592	0.16650	0.00701	1003	40	993	39	1.1
41	223186	41	0.09985	0.00393	3.84670	0.21842	0.27940	0.01145	1621	37	1588	57	2.3

Bunker Hill sequence - sedimentary member E 726 682
Sample 07BUNKER N 5003869

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
44	36722	69	0.06652	0.00278	1.52898	0.09133	0.16670	0.00712	823	44	994	39	-22.4
45	100937	62	0.09793	0.00389	3.76865	0.23523	0.27910	0.01344	1585	37	1587	67	-0.1
46	92153	0	0.07210	0.00285	1.80411	0.09866	0.18148	0.00685	989	40	1075	37	-9.5
47	102635	0	0.07358	0.00291	1.91839	0.11339	0.18909	0.00831	1030	40	1116	45	-9.1
48	150009	0	0.07306	0.00289	1.81036	0.10618	0.17972	0.00777	1016	40	1065	42	-5.3
51	322463	11	0.07346	0.00262	1.88341	0.13551	0.18595	0.01162	1027	36	1099	63	-7.7
52	68390	4	0.06765	0.00258	1.67250	0.12586	0.17932	0.01163	858	40	1063	63	-26.0
53	78836	5	0.09563	0.00349	3.87451	0.28559	0.29386	0.01881	1540	34	1661	93	-8.9
54	283254	2	0.08696	0.00311	2.89586	0.21093	0.24151	0.01532	1360	34	1395	79	-2.9
55	132133	2	0.07121	0.00258	1.81773	0.14373	0.18513	0.01301	963	37	1095	70	-14.8
56	113929	4	0.07121	0.00259	1.73321	0.12831	0.17654	0.01139	963	37	1048	62	-9.5
57	82560	5	0.06611	0.00254	1.53266	0.12400	0.16813	0.01197	810	40	1002	66	-25.6
59	101127	4	0.06760	0.00246	1.54055	0.11328	0.16528	0.01056	856	38	986	58	-16.3
60	175788	7	0.07235	0.00260	1.75529	0.13067	0.17595	0.01147	996	37	1045	63	-5.3
61	152116	40	0.07061	0.00257	1.70667	0.12708	0.17529	0.01139	946	37	1041	62	-10.9
62	196262	33	0.07181	0.00257	1.74035	0.12636	0.17576	0.01110	981	36	1044	61	-7.0
63	774670	34	0.07324	0.00261	1.80165	0.13643	0.17841	0.01192	1021	36	1058	65	-4.0
64	244025	159	0.07758	0.00395	1.98416	0.16851	0.18550	0.01262	1136	51	1097	68	3.7
65	239115	14	0.07349	0.00263	1.76788	0.13772	0.17447	0.01207	1028	36	1037	66	-1.0
68	229769	8	0.08049	0.00288	2.30399	0.17017	0.20761	0.01341	1209	35	1216	71	-0.6
69	328778	1	0.07253	0.00259	1.68981	0.11939	0.16897	0.01030	1001	36	1006	57	-0.6
70	271656	5	0.07147	0.00257	1.62751	0.11704	0.16516	0.01028	971	37	985	57	-1.6
71	85121	85	0.06581	0.00246	1.42143	0.10329	0.15665	0.00976	800	39	938	54	-18.5
73	104082	47	0.06687	0.00249	1.46505	0.10720	0.15889	0.01001	834	39	951	55	-15.1
75	152947	72	0.07188	0.00263	1.63730	0.12418	0.16521	0.01098	982	37	986	60	-0.4
76	146049	60	0.07244	0.00262	1.75297	0.12979	0.17550	0.01134	998	37	1042	62	-4.8
78	61251	51	0.06575	0.00265	1.53737	0.11493	0.16959	0.01067	798	42	1010	59	-28.6
79	154058	76	0.06797	0.00248	1.46927	0.11083	0.15679	0.01036	867	38	939	57	-8.8
80	67017	67	0.06729	0.00249	1.60364	0.11371	0.17285	0.01046	847	38	1028	57	-23.2
83	716296	83	0.07375	0.00263	1.71064	0.12506	0.16823	0.01074	1035	36	1002	59	3.4
84	196199	75	0.06974	0.00252	1.59746	0.11311	0.16614	0.01011	921	37	991	56	-8.2
85	433166	99	0.07343	0.00262	1.81122	0.12588	0.17890	0.01067	1026	36	1061	58	-3.7
86	433307	100	0.09415	0.00336	3.52816	0.26159	0.27179	0.01766	1511	34	1550	89	-2.9

Bunker Hill sequence - volcanic member E 729201

Sample 10BUNKER01 N 5007658

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
1	99493	198	0.05698	0.00235	0.56998	0.03447	0.07255	0.00321	491	91	452	19	8.3
2	56279	41	0.05930	0.00101	0.57787	0.04647	0.07068	0.00556	578	37	440	33	24.6
3	78807	21	0.05619	0.00064	0.56614	0.02628	0.07307	0.00329	460	25	455	20	1.2
4	36488	47	0.07757	0.00163	1.64337	0.12563	0.15366	0.01129	1136	21	921	63	20.2
5	24795	46	0.08080	0.00222	1.83804	0.14608	0.16498	0.01231	1217	27	984	68	20.6
6	28701	83	0.08150	0.00188	1.95648	0.13412	0.17412	0.01124	1233	23	1035	61	17.4
7	135961	48	0.05640	0.00071	0.55233	0.02320	0.07102	0.00285	468	28	442	17	5.7
8	54834	57	0.05615	0.00098	0.55480	0.02457	0.07166	0.00292	458	39	446	18	2.8
9	66949	29	0.05602	0.00064	0.54477	0.02229	0.07053	0.00277	453	25	439	17	3.2
10	14374	51	0.09330	0.00485	2.48285	0.20981	0.19300	0.01285	1494	49	1138	69	26.0
11	124688	66	0.05910	0.00129	0.57095	0.02876	0.07006	0.00318	571	47	437	19	24.3
12	40287	125	0.06083	0.00117	0.63037	0.04016	0.07516	0.00457	633	21	467	27	27.2
13	132180	39	0.05635	0.00056	0.55761	0.02316	0.07177	0.00290	466	22	447	17	4.3
14	119354	27	0.05594	0.00060	0.54403	0.02145	0.07054	0.00268	450	24	439	16	2.4
15	115299	52	0.05626	0.00060	0.53869	0.02213	0.06945	0.00275	462	24	433	17	6.6
16	86256	159	0.07592	0.00127	1.74749	0.15173	0.16695	0.01422	1093	17	995	78	9.6
17	80057	86	0.06035	0.00199	0.60395	0.03053	0.07259	0.00278	616	71	452	17	27.6
18	156912	168	0.18209	0.00264	12.70663	0.95792	0.50611	0.03744	2672	12	2640	158	1.5
20	124807	51	0.05601	0.00064	0.54070	0.02102	0.07001	0.00260	453	25	436	16	3.8
21	92635	56	0.05608	0.00061	0.54269	0.02291	0.07019	0.00286	455	24	437	17	4.1
22	128849	39	0.05568	0.00061	0.53583	0.02105	0.06980	0.00263	439	24	435	16	1.1
23	133266	59	0.05714	0.00088	0.55853	0.02409	0.07090	0.00286	497	34	442	17	11.5
24	104891	38	0.05566	0.00057	0.53661	0.02306	0.06992	0.00292	439	23	436	18	0.8
25	154277	53	0.05640	0.00065	0.54963	0.02275	0.07067	0.00281	468	26	440	17	6.2
26	174459	93	0.05596	0.00072	0.53077	0.02182	0.06879	0.00269	451	29	429	16	5.0
27	115874	87	0.05714	0.00091	0.56157	0.02343	0.07128	0.00275	497	35	444	17	11.1
28	63863	54	0.05597	0.00075	0.53300	0.02184	0.06907	0.00268	451	30	431	16	4.8
29	39738	5	0.08611	0.00223	1.99796	0.13819	0.16829	0.01079	1341	25	1003	59	27.2
31	82248	0	0.07764	0.00132	1.76770	0.11223	0.16513	0.01010	1138	17	985	56	14.4
33	57170	1	0.07749	0.00150	1.55280	0.10238	0.14532	0.00916	1134	19	875	51	24.4
34	60956	1	0.18684	0.00280	12.21891	0.77421	0.47432	0.02920	2715	12	2502	126	9.4
36	153071	1	0.09016	0.00136	2.85038	0.18722	0.22930	0.01466	1429	14	1331	76	7.6

Bunker Hill sequence - volcanic member E 729201
Sample 10BUNKER01 N 5007658

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	$^{207}Pb/^{206}Pb$	2 σ	$^{207}Pb/^{235}U$	2 σ	$^{206}Pb/^{238}U$	2 σ	$^{207}Pb/^{206}Pb$	2 σ	$^{206}Pb/^{238}U$	2 σ	disc.
37	178208	75	0.05600	0.00058	0.55899	0.02213	0.07239	0.00277	453	23	451	17	0.5
38	80563	69	0.05721	0.00097	0.57284	0.02674	0.07262	0.00316	500	37	452	19	9.9
40	175711	0	0.07428	0.00115	1.47951	0.11682	0.14446	0.01119	1049	16	870	63	18.3
41	39701	176	0.08301	0.00153	1.81761	0.12828	0.15880	0.01082	1269	18	950	60	27.0
42	71474	198	0.07833	0.00132	1.68128	0.13418	0.15566	0.01214	1155	17	933	67	20.7
43	219264	145	0.07397	0.00112	1.59492	0.11082	0.15638	0.01060	1041	15	937	59	10.7
44	216658	149	0.09632	0.00148	3.13815	0.20345	0.23630	0.01489	1554	14	1367	77	13.3
45	81213	144	0.07721	0.00138	1.58462	0.09468	0.14886	0.00848	1127	18	895	47	22.0
46	79910	144	0.07917	0.00133	1.77596	0.12314	0.16269	0.01094	1176	17	972	60	18.7
47	52136	66	0.05594	0.00071	0.51963	0.02107	0.06737	0.00260	450	28	420	16	6.8
48	102968	169	0.10236	0.00160	3.70937	0.25383	0.26282	0.01751	1667	14	1504	89	11.0
49	43018	137	0.09255	0.00208	2.52148	0.18487	0.19760	0.01379	1479	21	1162	74	23.4
51	40722	153	0.08573	0.00189	1.93671	0.12817	0.16384	0.01023	1332	21	978	56	28.6
52	104235	183	0.07652	0.00131	1.75146	0.13554	0.16601	0.01253	1109	17	990	69	11.5
53	29839	161	0.10835	0.00244	3.73492	0.28779	0.25001	0.01842	1772	21	1439	94	21.0
54	76966	179	0.09359	0.00154	2.93601	0.20794	0.22753	0.01567	1500	16	1322	82	13.1
55	17897	150	0.10723	0.00315	3.35864	0.26186	0.22717	0.01641	1753	27	1320	86	27.3
58	138618	84	0.09052	0.00136	2.73943	0.16626	0.21949	0.01291	1437	14	1279	68	12.1
60	57661	99	0.08041	0.00154	1.81690	0.12300	0.16387	0.01064	1207	19	978	59	20.4
61	107731	161	0.07600	0.00167	1.64308	0.10724	0.15681	0.00964	1095	22	939	53	15.3
62	138808	87	0.07471	0.00117	1.64501	0.10359	0.15969	0.00974	1061	16	955	54	10.7
63	35311	95	0.08393	0.00203	1.84968	0.11942	0.15984	0.00957	1291	24	956	53	27.9
64	141389	98	0.10148	0.00153	3.68955	0.23223	0.26368	0.01611	1651	14	1509	82	9.7
65	42198	90	0.07991	0.00155	1.64040	0.11030	0.14889	0.00959	1195	19	895	54	26.9
67	23267	109	0.14732	0.00294	7.35248	0.53305	0.36196	0.02523	2315	17	1991	118	16.2
68	66870	140	0.07869	0.00143	1.75535	0.10939	0.16178	0.00965	1164	18	967	53	18.3
69	324907	132	0.07519	0.00110	1.75054	0.10926	0.16886	0.01025	1074	15	1006	56	6.8
72	36471	44	0.08383	0.00213	1.79236	0.13171	0.15507	0.01069	1289	25	929	59	29.9
74	14676	53	0.12838	0.00493	4.93332	0.35721	0.27871	0.01711	2076	34	1585	86	26.6
78	38988	40	0.08477	0.00176	1.88215	0.13582	0.16104	0.01113	1310	20	963	62	28.5
79	86252	44	0.07614	0.00137	1.59125	0.10399	0.15157	0.00952	1099	18	910	53	18.4
80	143416	66	0.18894	0.00273	12.30662	0.93979	0.47241	0.03542	2733	12	2494	153	10.5
82	114198	68	0.07646	0.00133	1.65052	0.13130	0.15655	0.01215	1107	17	938	67	16.5

Bunker Hill sequence - volcanic member E 729201

Sample 10BUNKER01 N 5007658

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
83	80508	85	0.19694	0.00296	12.46495	0.87478	0.45905	0.03147	2801	12	2435	138	15.6
85	194771	128	0.07515	0.00113	1.64858	0.12863	0.15910	0.01218	1073	15	952	67	12.1
87	126828	53	0.10161	0.00151	3.70568	0.24222	0.26451	0.01683	1654	14	1513	85	9.6
88	59997	54	0.07943	0.00150	1.72088	0.10538	0.15712	0.00915	1183	19	941	51	22.0
90	107390	50	0.07626	0.00136	1.59238	0.10645	0.15145	0.00976	1102	18	909	54	18.7
92	243046	69	0.18264	0.00263	11.85198	0.81979	0.47064	0.03184	2677	12	2486	138	8.6
93	10419	63	0.10736	0.00337	2.40617	0.16620	0.16255	0.01000	1755	29	971	55	48.1
95	617652	51	0.11195	0.00161	4.62471	0.32772	0.29960	0.02079	1831	13	1689	102	8.8
97	26700	41	0.14286	0.00265	7.16060	0.47108	0.36353	0.02294	2262	16	1999	108	13.5
101	148658	28	0.07647	0.00115	1.67101	0.10408	0.15848	0.00958	1107	15	948	53	15.4
103	197894	32	0.07368	0.00113	1.63734	0.11843	0.16118	0.01139	1033	15	963	63	7.2
106	139917	10	0.10074	0.00131	3.54382	0.22387	0.25514	0.01577	1638	12	1465	81	11.8
107	583945	12	0.11154	0.00142	4.85861	0.36329	0.31591	0.02328	1825	12	1770	113	3.4
108	11826	13	0.09154	0.00308	2.33558	0.14724	0.18504	0.00986	1458	32	1094	53	27.1
109	37562	8	0.07877	0.00149	1.62015	0.12444	0.14918	0.01111	1166	19	896	62	24.8
110	10378	7	0.09487	0.00329	2.51522	0.16321	0.19228	0.01055	1526	33	1134	57	28.0
111	21208	6	0.08180	0.00218	1.69359	0.11497	0.15016	0.00937	1241	26	902	52	29.2
112	203062	5	0.07440	0.00097	1.71282	0.08689	0.16697	0.00818	1052	13	995	45	5.8
113	322931	4	0.18529	0.00236	12.91283	0.70512	0.50545	0.02684	2701	11	2637	114	2.9
114	82128	4	0.10088	0.00137	3.63171	0.17982	0.26109	0.01243	1640	13	1495	63	9.9
115	9717	7	0.09577	0.00352	2.51753	0.17828	0.19066	0.01154	1543	35	1125	62	29.5
116	16924	48	0.07314	0.00158	1.83281	0.09700	0.18174	0.00878	1018	22	1076	48	-6.2
117	113114	52	0.07094	0.00100	1.47517	0.07629	0.15082	0.00750	956	14	906	42	5.6
118	13993	45	0.09620	0.00209	3.76412	0.21633	0.28379	0.01510	1552	20	1610	75	-4.3
119	436312	48	0.16767	0.00213	9.73390	0.56446	0.42104	0.02382	2535	11	2265	107	12.6
120	48193	38	0.07319	0.00111	1.73203	0.10583	0.17164	0.01016	1019	15	1021	56	-0.2
121	643564	42	0.16475	0.00215	8.10778	0.59116	0.35692	0.02560	2505	11	1968	120	24.8
122	104745	16	0.20363	0.00262	14.10875	0.85271	0.50252	0.02968	2855	10	2625	126	9.8
124	152724	23	0.09911	0.00132	3.57305	0.20905	0.26146	0.01489	1608	12	1497	76	7.7
125	24196	23	0.07198	0.00118	1.70258	0.09846	0.17155	0.00952	985	17	1021	52	-3.9
126	52872	10	0.08317	0.00121	2.34707	0.16271	0.20468	0.01387	1273	14	1200	74	6.3
127	28286	29	0.06864	0.00131	1.47704	0.08212	0.15606	0.00815	888	20	935	45	-5.7
128	512182	12	0.11228	0.00142	4.98176	0.34522	0.32179	0.02192	1837	11	1798	106	2.4

Bunker Hill sequence - volcanic member E 729201

Sample 10BUNKER01 N 5007658

									age (Ma)				
Grain #	**^{206}Pb (cps)**	**^{204}Pb (cps)**	**$^{207}Pb/^{206}Pb$**	**2 σ**	**$^{207}Pb/^{235}U$**	**2 σ**	**$^{206}Pb/^{238}U$**	**2 σ**	**$^{207}Pb/^{206}Pb$**	**2 σ**	**$^{206}Pb/^{238}U$**	**2 σ**	**disc.**
129	48875	16	0.07374	0.00105	1.69448	0.08804	0.16667	0.00833	1034	14	994	46	4.2
130	92647	5	0.09235	0.00127	3.02502	0.15716	0.23757	0.01190	1475	13	1374	62	7.6

Saint-Victor Formation E 818868

Sample 09SV01 N 5115491

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
1	59193	35	0.08655	0.00169	2.54752	0.15751	0.21347	0.01253	1351	38	1247	66	8.4
2	23090	29	0.06876	0.00184	1.63054	0.13242	0.17199	0.01318	892	55	1023	72	-16.0
3	29344	36	0.07356	0.00355	1.71492	0.14935	0.16909	0.01225	1029	98	1007	67	2.3
4	34357	14	0.09675	0.00122	3.81820	0.22835	0.28624	0.01673	1562	24	1623	83	-4.4
5	165848	17	0.07493	0.00063	1.79540	0.11579	0.17378	0.01111	1067	17	1033	61	3.4
6	241400	25	0.18652	0.00132	13.11043	0.93625	0.50979	0.03623	2712	12	2656	153	2.5
7	45402	13	0.07308	0.00124	1.68656	0.10498	0.16737	0.01002	1016	34	998	55	2.0
8	17785	8	0.06859	0.00168	1.70016	0.10624	0.17977	0.01034	886	51	1066	56	-21.9
9	44198	14	0.08329	0.00138	2.41414	0.19805	0.21023	0.01689	1276	32	1230	89	4.0
10	353013	12	0.18923	0.00163	12.56189	0.82870	0.48146	0.03149	2736	14	2534	136	8.9
11	134151	21	0.28094	0.00295	24.64147	1.40962	0.63615	0.03577	3368	16	3174	139	7.3
12	15348	7	0.06720	0.00181	1.67629	0.12253	0.18091	0.01230	844	56	1072	67	-29.3
14	40651	13	0.09761	0.00121	3.55784	0.19896	0.26435	0.01441	1579	23	1512	73	4.8
15	155586	15	0.07256	0.00066	1.62668	0.09785	0.16258	0.00967	1002	18	971	53	3.3
16	17241	21	0.07827	0.00200	2.64750	0.16189	0.24532	0.01363	1154	51	1414	70	-25.2
17	24028	14	0.07380	0.00159	1.89077	0.13149	0.18581	0.01229	1036	43	1099	66	-6.6
18	17405	22	0.07345	0.00322	1.80389	0.15223	0.17811	0.01284	1027	89	1057	70	-3.2
19	9167	13	0.07424	0.00558	1.86724	0.20109	0.18241	0.01408	1048	151	1080	76	-3.3
20	1182326	26	0.17659	0.00123	11.37709	0.81571	0.46726	0.03334	2621	12	2472	145	6.9
21	22022	49	0.07013	0.00292	1.53693	0.10729	0.15895	0.00890	932	86	951	49	-2.2
22	54064	46	0.08579	0.00125	2.59932	0.14363	0.21976	0.01171	1333	28	1281	62	4.4
23	150238	43	0.20052	0.00148	14.99191	0.86809	0.54225	0.03114	2830	12	2793	129	1.6
24	976571	84	0.11332	0.00076	5.18280	0.30655	0.33170	0.01949	1853	12	1847	94	0.4
25	111980	63	0.07770	0.00068	2.04708	0.11110	0.19109	0.01023	1139	18	1127	55	1.1
26	27221	59	0.08409	0.00160	2.55731	0.14737	0.22056	0.01200	1295	37	1285	63	0.8
27	55836	65	0.07915	0.00101	2.09680	0.12612	0.19214	0.01129	1176	25	1133	61	4.0
28	261129	61	0.08827	0.00070	2.83093	0.15961	0.23260	0.01298	1388	15	1348	68	3.2
29	53808	80	0.07408	0.00117	1.74566	0.11272	0.17091	0.01070	1044	32	1017	59	2.8
30	158044	67	0.10776	0.00127	4.30896	0.25146	0.29002	0.01658	1762	22	1642	82	7.7
31	45484	76	0.07474	0.00105	1.76355	0.10498	0.17113	0.00990	1062	28	1018	54	4.4
33	30889	74	0.07371	0.00134	1.73567	0.10365	0.17078	0.00971	1034	37	1016	53	1.8
34	102827	82	0.07477	0.00071	1.80501	0.09941	0.17509	0.00950	1062	19	1040	52	2.3
35	34151	71	0.07653	0.00106	2.01717	0.11654	0.19117	0.01072	1109	28	1128	58	-1.8

Saint-Victor Formation E 818868
Sample 09SV01 N 5115491

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
36	57536	85	0.07438	0.00095	1.75402	0.09911	0.17103	0.00941	1052	26	1018	52	3.5
37	31783	60	0.07257	0.00092	1.66371	0.10287	0.16628	0.01006	1002	26	992	55	1.1
38	74422	65	0.10361	0.00105	3.97868	0.23398	0.27851	0.01613	1690	19	1584	81	7.1
39	67351	76	0.08252	0.00100	2.34338	0.15271	0.20597	0.01319	1258	24	1207	70	4.4
40	52924	71	0.09046	0.00103	2.95989	0.19238	0.23731	0.01519	1435	22	1373	79	4.8
41	115703	77	0.11320	0.00097	4.98715	0.30609	0.31952	0.01942	1851	15	1787	94	4.0
42	11835	64	0.07712	0.00323	2.04391	0.14151	0.19223	0.01059	1124	84	1133	57	-0.9
43	121166	80	0.07533	0.00069	1.79255	0.10342	0.17259	0.00983	1077	18	1026	54	5.1
44	651023	75	0.11343	0.00077	5.02819	0.35433	0.32149	0.02255	1855	12	1797	109	3.6
45	165881	83	0.11453	0.00082	5.11032	0.29683	0.32361	0.01865	1873	13	1807	90	4.0
46	15338	101	0.08914	0.00378	2.40210	0.17807	0.19545	0.01188	1407	81	1151	64	19.9
47	191919	120	0.11584	0.00114	5.10203	0.27571	0.31943	0.01697	1893	18	1787	82	6.4
48	82841	93	0.08504	0.00084	2.43923	0.14080	0.20804	0.01183	1316	19	1218	63	8.2
49	75164	127	0.09427	0.00108	3.24703	0.21803	0.24981	0.01653	1514	22	1438	85	5.6
50	213192	105	0.19233	0.00138	13.82129	0.78787	0.52119	0.02947	2762	12	2704	124	2.6
51	62352	107	0.08939	0.00077	2.92883	0.16887	0.23763	0.01355	1413	17	1374	70	3.0
52	24604	106	0.07202	0.00109	1.57598	0.09865	0.15871	0.00964	987	31	950	53	4.0
53	208490	104	0.18739	0.00136	13.15129	0.73260	0.50901	0.02811	2719	12	2652	119	3.0
54	11850	102	0.07442	0.00151	1.85363	0.11465	0.18064	0.01055	1053	41	1070	57	-1.8
55	100080	122	0.19187	0.00141	13.15731	0.75594	0.49734	0.02834	2758	12	2602	121	6.9
56	75519	122	0.08540	0.00076	2.60687	0.15972	0.22139	0.01342	1325	17	1289	70	3.0
57	19579	116	0.07429	0.00273	1.88173	0.12866	0.18371	0.01059	1049	74	1087	57	-3.9
58	24069	110	0.07152	0.00151	1.64210	0.10929	0.16652	0.01051	972	43	993	58	-2.3
59	27830	117	0.07610	0.00126	1.87503	0.11254	0.17871	0.01031	1098	33	1060	56	3.7
60	7971	121	0.07717	0.00239	2.05199	0.12830	0.19284	0.01048	1126	62	1137	56	-1.1
61	425912	81	0.19215	0.00130	13.50325	0.77048	0.50967	0.02888	2761	11	2655	122	4.7
62	99219	62	0.18899	0.00135	12.87859	0.73601	0.49424	0.02803	2733	12	2589	120	6.4
63	81270	61	0.09993	0.00085	3.81269	0.22170	0.27672	0.01592	1623	16	1575	80	3.3
64	61409	80	0.07474	0.00066	1.76426	0.10496	0.17121	0.01007	1061	18	1019	55	4.3
65	42734	74	0.09199	0.00093	3.04614	0.17411	0.24017	0.01351	1467	19	1388	70	6.0
66	906221	75	0.18107	0.00121	12.60496	0.72763	0.50489	0.02895	2663	11	2635	123	1.3
67	190036	74	0.07556	0.00056	1.84302	0.10124	0.17691	0.00963	1083	15	1050	53	3.3
68	25760	41	0.06833	0.00113	1.51176	0.08907	0.16046	0.00907	879	34	959	50	-9.9

Saint-Victor Formation E 818868
Sample 09SV01 N 5115491

Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	age (Ma) ^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
69	170425	213	0.07908	0.00074	2.07512	0.11867	0.19031	0.01074	1174	19	1123	58	4.7
70	92425	69	0.07611	0.00069	1.87269	0.10981	0.17845	0.01034	1098	18	1058	56	3.9
71	54790	43	0.08455	0.00088	2.59050	0.15125	0.22221	0.01277	1305	20	1294	67	1.0
72	368079	42	0.07272	0.00056	1.61697	0.08797	0.16127	0.00869	1006	16	964	48	4.5
73	81589	34	0.09000	0.00078	3.00738	0.17228	0.24236	0.01373	1425	16	1399	71	2.1
74	26279	26	0.07102	0.00116	1.63901	0.09611	0.16737	0.00943	958	33	998	52	-4.4
78	107939	47	0.08185	0.00072	2.25508	0.14364	0.19982	0.01260	1242	17	1174	67	6.0
76	35086	40	0.07187	0.00089	1.52286	0.09885	0.15368	0.00979	982	25	922	54	6.6
77	19314	44	0.07025	0.00128	1.60996	0.09670	0.16621	0.00951	936	37	991	52	-6.4
78	55243	43	0.08816	0.00088	2.66896	0.14954	0.21957	0.01210	1386	19	1280	64	8.5
79	51983	34	0.08465	0.00085	2.39114	0.12983	0.20486	0.01093	1308	19	1201	58	8.9
80	49602	23	0.07802	0.00076	1.95129	0.10872	0.18140	0.00995	1147	19	1075	54	6.9
81	349459	25	0.11373	0.00078	4.81231	0.27042	0.30690	0.01712	1860	12	1725	84	8.2
82	342365	26	0.11329	0.00079	4.73762	0.27126	0.30330	0.01724	1853	13	1708	85	8.9
83	77850	46	0.07129	0.00068	1.48662	0.08317	0.15124	0.00834	966	19	908	47	6.4
84	248422	40	0.07352	0.00054	1.62410	0.09826	0.16022	0.00962	1028	15	958	53	7.4
85	20685	27	0.12833	0.00135	6.53766	0.43117	0.36948	0.02406	2075	18	2027	112	2.7

Saint-Victor Formation E 737469
Sample 09SV02 N 5031591

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
1	24720	47	0.05221	0.00119	0.42930	0.01891	0.05964	0.00225	294	52	373	14	-27.6
2	41568	52	0.09922	0.00171	3.13680	0.12178	0.22929	0.00798	1610	32	1331	42	19.2
3	318474	58	0.07258	0.00100	1.51230	0.05523	0.15113	0.00511	1002	28	907	29	10.1
4	10248	56	0.06755	0.00214	1.39394	0.06583	0.14966	0.00524	855	66	899	29	-5.5
5	41297	74	0.05507	0.00120	0.51728	0.02173	0.06812	0.00244	415	49	425	15	-2.4
6	52960	78	0.07214	0.00188	1.30342	0.08667	0.13105	0.00802	990	53	794	46	21.0
7	69341	58	0.05451	0.00058	0.50959	0.02133	0.06780	0.00275	392	24	423	17	-8.1
9	107422	88	0.11610	0.00178	4.24839	0.21181	0.26540	0.01259	1897	28	1517	64	22.4
10	42310	69	0.06780	0.00123	1.00002	0.05511	0.10697	0.00557	862	38	655	32	25.3
11	221688	87	0.07383	0.00111	1.35366	0.24230	0.13298	0.02372	1037	30	805	134	23.8
12	12300	64	0.08232	0.00285	2.31363	0.39519	0.20385	0.03410	1253	68	1196	180	5.0
13	67017	46	0.05382	0.00073	0.51728	0.01998	0.06971	0.00252	363	31	434	15	-20.2
14	36608	84	0.10115	0.00207	3.37715	0.57800	0.24214	0.04115	1645	38	1398	210	16.7
15	67317	29	0.05402	0.00071	0.51683	0.01870	0.06939	0.00234	372	30	432	14	-16.9
16	59565	67	0.09892	0.00144	3.26468	0.55500	0.23937	0.04054	1604	27	1383	208	15.3
17	30028	56	0.07065	0.00145	1.43676	0.24133	0.14750	0.02459	947	42	887	137	6.8
18	38246	42	0.05423	0.00130	0.51564	0.02262	0.06897	0.00254	381	54	430	15	-13.4
19	58649	48	0.05400	0.00074	0.51593	0.01992	0.06929	0.00250	371	31	432	15	-17.0
20	81858	69	0.11342	0.00172	4.11349	0.69460	0.26303	0.04424	1855	27	1505	222	21.1
21	68623	68	0.05694	0.00148	0.54980	0.02503	0.07003	0.00262	489	57	436	16	11.2
22	37807	73	0.08229	0.00127	2.32095	0.17934	0.20456	0.01549	1252	30	1200	82	4.6
24	32511	79	0.05446	0.00096	0.48997	0.03863	0.06525	0.00501	390	40	407	30	-4.6
25	13643	63	0.06731	0.00211	1.64933	0.12981	0.17772	0.01283	847	65	1055	70	-26.5
27	26973	85	0.05629	0.00127	0.50874	0.03971	0.06555	0.00490	464	50	409	30	12.1
28	131509	74	0.07223	0.00106	1.53807	0.11564	0.15445	0.01139	992	30	926	63	7.2
29	58159	56	0.05624	0.00100	0.53109	0.02056	0.06849	0.00235	462	40	427	14	7.8
30	25735	77	0.05345	0.00158	0.50125	0.04128	0.06802	0.00523	348	67	424	31	-22.7
31	33548	95	0.07078	0.00145	1.66908	0.25471	0.17103	0.02586	951	42	1018	141	-7.6
32	94906	78	0.07226	0.00104	1.70301	0.25988	0.17094	0.02597	993	29	1017	141	-2.6
33	74276	38	0.05433	0.00075	0.50491	0.01887	0.06741	0.00234	385	31	421	14	-9.6
34	66980	88	0.07381	0.00126	1.71467	0.26082	0.16848	0.02546	1036	35	1004	139	3.4
35	28980	112	0.07410	0.00180	1.61373	0.24978	0.15796	0.02415	1044	49	945	133	10.2
36	54000	37	0.05351	0.00114	0.49636	0.01986	0.06727	0.00227	351	48	420	14	-20.4

Saint-Victor Formation E 737469
Sample 09SV02 N 5031591

Grain #	^{206}Pb (cps)	^{204}Pb (cps)	$^{207}Pb/^{206}Pb$	2 σ	$^{207}Pb/^{235}U$	2 σ	$^{206}Pb/^{238}U$	2 σ	age (Ma) $^{207}Pb/^{206}Pb$	2 σ	$^{206}Pb/^{238}U$	2 σ	disc.
37	37931	98	0.07224	0.00125	1.75355	0.26411	0.17605	0.02634	993	35	1045	143	-5.7
39	78701	80	0.07261	0.00113	1.69946	0.25692	0.16976	0.02553	1003	32	1011	139	-0.8
41	65246	48	0.08384	0.00142	2.45143	0.25701	0.21207	0.02194	1289	33	1240	116	4.2
42	45262	53	0.07201	0.00130	1.66831	0.17314	0.16802	0.01717	986	37	1001	94	-1.6
43	36835	47	0.06805	0.00114	1.42344	0.14684	0.15172	0.01544	870	35	911	86	-5.0
44	64539	51	0.07184	0.00116	1.58044	0.16382	0.15955	0.01633	981	33	954	90	3.0
45	138238	63	0.05554	0.00060	0.50643	0.02082	0.06613	0.00262	434	24	413	16	5.1
46	55111	51	0.09101	0.00137	2.94837	0.31024	0.23495	0.02447	1447	29	1360	126	6.6
47	127955	71	0.07261	0.00112	1.60381	0.16561	0.16019	0.01636	1003	31	958	90	4.9
48	55584	72	0.05604	0.00131	0.53741	0.02239	0.06955	0.00240	454	52	433	14	4.6
49	85751	75	0.05460	0.00070	0.51715	0.01901	0.06870	0.00237	396	29	428	14	-8.5
50	44062	50	0.05316	0.00131	0.48218	0.05177	0.06578	0.00687	336	56	411	41	-23.0
51	37581	70	0.08105	0.00450	1.92124	0.21440	0.17191	0.01664	1223	109	1023	91	17.7
52	45513	47	0.05335	0.00085	0.47422	0.04571	0.06447	0.00613	344	36	403	37	-17.8
53	40314	60	0.05469	0.00141	0.50114	0.02241	0.06646	0.00242	399	58	415	15	-4.0
54	15145	43	0.05390	0.00449	0.51559	0.06416	0.06938	0.00641	367	188	432	39	-18.5
55	22297	68	0.05441	0.00337	0.48625	0.03570	0.06482	0.00255	388	139	405	15	-4.5
56	66833	95	0.05569	0.00198	0.51697	0.02595	0.06733	0.00238	440	79	420	14	4.7
58	17542	64	0.07791	0.00515	1.95753	0.22157	0.18222	0.01675	1145	131	1079	91	6.2
59	41116	47	0.07324	0.00160	1.66857	0.15628	0.16523	0.01505	1021	44	986	83	3.7
60	51574	38	0.07410	0.00132	1.65307	0.15398	0.16179	0.01479	1044	36	967	82	8.0
61	14574	46	0.06771	0.00287	1.71992	0.08676	0.18424	0.00503	860	88	1090	27	-29.2
62	32389	38	0.05366	0.00123	0.49618	0.01676	0.06706	0.00166	357	52	418	10	-17.8
64	22469	28	0.06849	0.00135	1.54099	0.04819	0.16318	0.00395	883	41	974	22	-11.1
65	25123	40	0.07129	0.00178	1.93033	0.06365	0.19638	0.00424	966	51	1156	23	-21.5
67	34778	198	0.05757	0.00332	0.52521	0.03840	0.06617	0.00297	513	127	413	18	20.2
68	37582	46	0.07438	0.00202	1.72468	0.05927	0.16817	0.00354	1052	55	1002	20	5.1
69	88669	37	0.05572	0.00105	0.59390	0.01889	0.07731	0.00198	441	42	480	12	-9.2
70	31319	35	0.06778	0.00158	1.65567	0.05081	0.17716	0.00354	862	48	1051	19	-23.9
71	65931	47	0.08206	0.00136	2.31556	0.14984	0.20465	0.01280	1247	32	1200	68	4.1
72	66805	60	0.05512	0.00096	0.48865	0.02618	0.06430	0.00326	417	39	402	20	3.8
73	46113	59	0.05274	0.00106	0.45988	0.02202	0.06324	0.00275	318	46	395	17	-25.2

Saint-Victor Formation E 737469
Sample 09SV02 N 5031591

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
74	104009	97	0.05638	0.00120	0.52602	0.03015	0.06767	0.00360	467	47	422	22	10.0
75	32451	61	0.05454	0.00249	0.55484	0.03804	0.07378	0.00378	394	102	459	23	-17.2
76	353921	75	0.07894	0.00114	2.17683	0.12033	0.19999	0.01067	1171	29	1175	57	-0.4
79	96713	57	0.07165	0.00171	1.59846	0.09582	0.16181	0.00890	976	49	967	49	1.0
80	69167	76	0.05405	0.00055	0.51428	0.01937	0.06901	0.00250	373	23	430	15	-15.9
81	117030	37	0.06555	0.00108	0.94641	0.08323	0.10471	0.00905	792	34	642	53	19.9
82	64313	68	0.05393	0.00071	0.50595	0.01917	0.06804	0.00242	368	30	424	15	-15.7
83	35631	31	0.07160	0.00145	1.64387	0.09257	0.16652	0.00875	975	41	993	48	-2.0
84	75621	25	0.09983	0.00149	3.59115	0.20379	0.26089	0.01428	1621	28	1494	73	8.7
85	129947	40	0.07397	0.00113	1.84332	0.11975	0.18075	0.01141	1041	31	1071	62	-3.2

Compton Formation - Milan member E 800320
Sample 09MILAN01 N 5056403

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
4	48218	7	0.06797	0.00208	1.61491	0.08457	0.17231	0.00733	868	32	1025	40	-19.6
7	56022	10	0.06439	0.00190	1.36875	0.06499	0.15417	0.00574	754	31	924	32	-24.2
11	296610	51	0.10253	0.00257	4.00416	0.17315	0.28325	0.00998	1670	23	1608	50	4.2
12	44914	35	0.07954	0.00259	2.35102	0.17169	0.21436	0.01402	1186	32	1252	74	-6.2
18	106628	42	0.05385	0.00142	0.50098	0.02210	0.06747	0.00239	365	30	421	14	-15.8
19	44259	43	0.06801	0.00206	1.51794	0.06966	0.16188	0.00558	869	31	967	31	-12.2
20	95147	25	0.10008	0.00256	3.79217	0.18058	0.27481	0.01104	1626	24	1565	56	4.2
21	163971	42	0.07696	0.00195	1.95923	0.10535	0.18465	0.00875	1120	25	1092	47	2.7
26	57092	30	0.06884	0.00203	1.43466	0.07576	0.15115	0.00662	894	30	907	37	-1.6
27	113987	51	0.07177	0.00186	1.54332	0.07926	0.15595	0.00691	979	26	934	38	5.0
29	30714	48	0.07445	0.00218	2.11174	0.11069	0.20572	0.00895	1054	29	1206	48	-15.9
30	65406	52	0.08818	0.00242	2.77726	0.12364	0.22843	0.00800	1386	26	1326	42	4.8
32	45752	47	0.06791	0.00211	1.53355	0.07378	0.16377	0.00602	866	32	978	33	-13.9
33	58553	63	0.09858	0.00273	3.83108	0.21251	0.28186	0.01355	1597	26	1601	68	-0.2
34	75633	58	0.09989	0.00261	3.74227	0.22601	0.27171	0.01480	1622	24	1549	75	5.0
36	54753	56	0.07094	0.00201	1.75280	0.08569	0.17919	0.00715	956	29	1063	39	-12.1
38	51289	77	0.06919	0.00197	1.55639	0.06831	0.16315	0.00545	904	29	974	30	-8.3
43	163876	66	0.05471	0.00140	0.52415	0.02386	0.06948	0.00262	400	29	433	16	-8.4
45	102350	70	0.07224	0.00192	1.72275	0.08627	0.17296	0.00734	993	27	1028	40	-3.9
46	148705	73	0.16671	0.00418	11.06694	0.65409	0.48146	0.02577	2525	21	2534	111	-0.4
47	44762	59	0.08477	0.00245	2.74620	0.14691	0.23495	0.01058	1310	28	1360	55	-4.2
48	56174	49	0.11084	0.00286	4.81771	0.26800	0.31523	0.01554	1813	23	1766	76	3.0
49	544493	63	0.07606	0.00191	1.95241	0.09755	0.18618	0.00804	1097	25	1101	44	-0.4
54	44311	11	0.06625	0.00205	1.47208	0.08269	0.16114	0.00756	814	32	963	42	-19.7
56	83238	23	0.20171	0.00511	15.80533	1.03613	0.56831	0.03436	2840	21	2901	140	-2.7
57	143060	18	0.19755	0.00495	14.50804	0.73413	0.53264	0.02342	2806	20	2753	98	2.3
59	85099	14	0.08299	0.00227	2.35490	0.11071	0.20579	0.00787	1269	27	1206	42	5.4
60	84452	16	0.07902	0.00210	2.24020	0.11382	0.20562	0.00891	1173	26	1205	47	-3.1
61	29097	54	0.05714	0.00114	0.47520	0.05528	0.06032	0.00691	497	22	378	42	24.7
62	35559	59	0.05747	0.00081	0.48183	0.04511	0.06081	0.00563	510	16	381	34	26.1
63	68317	52	0.08186	0.00106	1.82435	0.17980	0.16164	0.01579	1242	13	966	87	23.9
65	321875	44	0.07716	0.00071	1.70378	0.14343	0.16015	0.01340	1125	9	958	74	16.0
66	57052	48	0.08080	0.00094	1.83361	0.17655	0.16458	0.01573	1217	11	982	86	20.8

Compton Formation - Milan member E 800320
Sample 09MILAN01 N 5056403

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
67	17032	46	0.05598	0.00109	0.47110	0.04168	0.06104	0.00527	451	22	382	32	15.8
68	40264	40	0.07372	0.00108	1.40940	0.12688	0.13866	0.01231	1034	15	837	69	20.3
69	14295	41	0.05673	0.00131	0.48116	0.04201	0.06151	0.00518	481	25	385	31	20.6
70	149899	51	0.07561	0.00073	1.55505	0.14087	0.14916	0.01343	1085	10	896	75	18.6
71	182656	62	0.08326	0.00075	2.43740	0.21944	0.21231	0.01902	1275	9	1241	100	3.0
72	51064	70	0.07344	0.00079	1.77209	0.18350	0.17500	0.01802	1026	11	1040	98	-1.4
73	7440	64	0.06068	0.00247	0.71752	0.06361	0.08577	0.00675	628	44	530	40	16.1
74	45352	80	0.05702	0.00064	0.64549	0.06847	0.08211	0.00866	492	12	509	51	-3.5
75	68189	82	0.07527	0.00080	1.93438	0.17232	0.18639	0.01649	1076	11	1102	89	-2.6
76	26667	77	0.07613	0.00111	1.99289	0.14399	0.18986	0.01343	1098	15	1121	72	-2.2
77	49302	96	0.05570	0.00058	0.60766	0.05375	0.07912	0.00695	441	12	491	41	-11.9
78	26022	91	0.05602	0.00091	0.58946	0.05283	0.07632	0.00673	453	18	474	40	-4.8
79	5087	99	0.06065	0.00301	0.69477	0.06138	0.08308	0.00607	627	54	514	36	18.6
80	88792	111	0.07372	0.00068	1.73764	0.16947	0.17096	0.01660	1034	9	1017	91	1.7
81	35043	120	0.05821	0.00110	0.53089	0.03814	0.06614	0.00459	538	21	413	28	24.0
82	47570	118	0.07324	0.00086	1.35318	0.10464	0.13400	0.01024	1021	12	811	58	21.9
83	153809	123	0.09296	0.00084	2.68156	0.22608	0.20921	0.01754	1487	9	1225	93	19.4
84	168886	130	0.05536	0.00052	0.42780	0.03424	0.05604	0.00445	427	11	352	27	18.1

Compton Formation - Milan member E 800320
Sample 09MILAN01a grain set N 5056403

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	$^{207}Pb/^{206}Pb$	2 σ	$^{207}Pb/^{235}U$	2 σ	$^{206}Pb/^{238}U$	2 σ	$^{207}Pb/^{206}Pb$	2 σ	$^{206}Pb/^{238}U$	2 σ	disc.
1	71754	73	0.05560	0.00068	0.51704	0.02334	0.06744	0.00293	436	27	421	18	3.7
2	126039	109	0.05536	0.00058	0.47090	0.01914	0.06170	0.00242	427	24	386	15	9.8
3	73036	73	0.05609	0.00051	0.52470	0.02315	0.06784	0.00293	456	20	423	18	7.4
4	116715	91	0.05451	0.00050	0.47020	0.02057	0.06256	0.00268	392	20	391	16	0.3
5	23653	95	0.05526	0.00142	0.47900	0.02452	0.06287	0.00278	423	57	393	17	7.2
6	211815	111	0.05467	0.00044	0.47159	0.01723	0.06256	0.00223	399	18	391	14	1.9
7	113250	103	0.05483	0.00049	0.47204	0.01795	0.06244	0.00231	405	20	390	14	3.7
8	174877	98	0.05447	0.00043	0.47322	0.01787	0.06301	0.00233	390	18	394	14	-0.9
9	86807	109	0.05473	0.00044	0.46472	0.01943	0.06158	0.00253	401	18	385	15	4.2
10	109151	97	0.05485	0.00050	0.48483	0.01829	0.06410	0.00235	406	20	401	14	1.5
11	57076	126	0.05677	0.00065	0.53980	0.01927	0.06897	0.00233	482	25	430	14	11.3
12	161856	116	0.05448	0.00039	0.47089	0.01774	0.06269	0.00232	391	16	392	14	-0.3
13	92964	107	0.05614	0.00047	0.52930	0.02275	0.06839	0.00288	458	18	426	17	7.1
14	30700	100	0.05585	0.00084	0.51351	0.02164	0.06668	0.00262	446	34	416	16	7.0
15	136124	119	0.07715	0.00069	1.86420	0.08562	0.17524	0.00789	1125	18	1041	43	8.1
16	153361	88	0.05437	0.00041	0.45990	0.01660	0.06135	0.00217	386	17	384	13	0.7
17	82008	130	0.05691	0.00054	0.56017	0.02890	0.07139	0.00362	488	21	445	22	9.2
19	227467	90	0.05469	0.00041	0.46823	0.02153	0.06209	0.00282	400	17	388	17	2.9
20	35386	102	0.05428	0.00064	0.45326	0.01697	0.06056	0.00215	383	26	379	13	1.0
21	90252	101	0.05594	0.00050	0.54016	0.02344	0.07003	0.00298	450	20	436	18	3.2
22	204799	128	0.05452	0.00040	0.45269	0.01900	0.06022	0.00249	392	16	377	15	4.1
23	206693	110	0.06008	0.00057	0.67209	0.02574	0.08113	0.00301	607	21	503	18	17.8
24	69684	130	0.07338	0.00059	1.63709	0.05792	0.16180	0.00557	1025	16	967	31	6.1
25	38166	123	0.05446	0.00045	0.45548	0.01658	0.06065	0.00215	390	18	380	13	2.8
26	47323	130	0.05655	0.00075	0.52895	0.02230	0.06784	0.00272	474	29	423	16	11.1
27	245140	119	0.07433	0.00050	1.70667	0.09897	0.16654	0.00959	1050	14	993	53	5.9
28	125420	139	0.05497	0.00047	0.47654	0.01825	0.06287	0.00235	411	19	393	14	4.5
29	262781	118	0.07410	0.00054	1.67742	0.05895	0.16418	0.00564	1044	15	980	31	6.6
30	358233	110	0.07174	0.00050	1.54307	0.06077	0.15601	0.00605	978	14	935	34	4.8

Compton Formation - Milan member E 836025
Sample 09MILAN02 N 5109089

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
2	27667	65	0.07776	0.00102	2.12977	0.12602	0.19864	0.01146	1141	13	1168	61	-2.6
3	56045	65	0.07394	0.00069	1.84293	0.13042	0.18078	0.01268	1040	9	1071	69	-3.3
4	285428	64	0.16868	0.00130	11.73750	0.67463	0.50467	0.02875	2545	6	2634	122	-4.3
5	41568	75	0.07849	0.00085	2.10336	0.13246	0.19435	0.01206	1159	11	1145	65	1.4
6	19651	82	0.07800	0.00143	2.16927	0.12861	0.20170	0.01137	1147	18	1184	61	-3.6
7	38715	62	0.05561	0.00070	0.51557	0.02886	0.06724	0.00367	437	14	420	22	4.1
8	29772	62	0.05565	0.00074	0.53113	0.02934	0.06922	0.00371	438	15	431	22	1.6
9	27705	56	0.07609	0.00098	1.83292	0.11263	0.17471	0.01050	1097	13	1038	57	5.9
10	41808	67	0.05478	0.00062	0.48350	0.03248	0.06401	0.00424	403	13	400	26	0.8
11	29300	30	0.07251	0.00094	1.67506	0.10970	0.16754	0.01076	1000	13	999	59	0.2
12	53155	34	0.08829	0.00081	2.78404	0.16674	0.22870	0.01354	1389	9	1328	71	4.9
13	51166	28	0.05496	0.00062	0.51729	0.03002	0.06826	0.00389	411	13	426	23	-3.7
14	46278	17	0.05539	0.00072	0.53247	0.02797	0.06972	0.00355	428	14	434	21	-1.6
16	74903	14	0.05967	0.00060	0.82267	0.04547	0.09999	0.00543	592	11	614	32	-4.0
17	70194	14	0.07295	0.00073	1.39906	0.07333	0.13910	0.00716	1012	10	840	40	18.2
18	602632	15	0.07583	0.00059	1.95871	0.12339	0.18734	0.01171	1091	8	1107	63	-1.6
19	15702	10	0.07172	0.00141	1.74816	0.09833	0.17677	0.00932	978	20	1049	51	-7.9
20	80164	14	0.05593	0.00051	0.50492	0.03017	0.06548	0.00387	450	10	409	23	9.4
21	29258	40	0.05532	0.00074	0.54553	0.03001	0.07152	0.00381	425	15	445	23	-4.9
22	8653	48	0.07144	0.00211	1.81241	0.12355	0.18400	0.01130	970	30	1089	61	-13.3
23	62479	36	0.07425	0.00070	1.75472	0.11611	0.17141	0.01123	1048	10	1020	61	2.9
24	66592	46	0.10356	0.00096	4.14042	0.23486	0.28997	0.01623	1689	9	1641	81	3.2
25	63962	51	0.05999	0.00064	0.80154	0.04417	0.09691	0.00524	603	11	596	31	1.2
26	25184	48	0.07434	0.00094	1.79198	0.10465	0.17483	0.00997	1051	13	1039	54	1.2
27	40156	50	0.07421	0.00083	1.69303	0.11188	0.16547	0.01078	1047	11	987	59	6.2
28	246542	67	0.18226	0.00142	12.73073	0.79226	0.50659	0.03128	2674	6	2642	132	1.4
29	28378	54	0.05589	0.00079	0.54593	0.03447	0.07085	0.00436	448	16	441	26	1.5
30	4003	48	0.05811	0.00374	0.67585	0.06095	0.08436	0.00533	534	70	522	32	2.3
31	63225	53	0.07427	0.00065	1.77475	0.09855	0.17331	0.00950	1049	9	1030	52	1.9
32	18920	49	0.07861	0.00131	2.42358	0.16720	0.22360	0.01497	1162	16	1301	78	-13.2
33	23674	54	0.08676	0.00109	2.59857	0.16884	0.21722	0.01385	1355	12	1267	73	7.2
34	17417	61	0.07419	0.00138	1.92703	0.13902	0.18838	0.01313	1047	19	1113	71	-6.9
35	9555	47	0.05801	0.00185	0.62263	0.03740	0.07784	0.00396	530	35	483	24	9.2

Compton Formation - Milan member E 836025
Sample 09MILAN02 N 5109089

									age (Ma)				
Grain #	**^{206}Pb (cps)**	**^{204}Pb (cps)**	**^{207}Pb/^{206}Pb**	**2 σ**	**^{207}Pb/^{235}U**	**2 σ**	**^{206}Pb/^{238}U**	**2 σ**	**^{207}Pb/^{206}Pb**	**2 σ**	**^{206}Pb/^{238}U**	**2 σ**	**disc.**
36	203354	54	0.07713	0.00064	1.93710	0.12100	0.18215	0.01128	1125	8	1079	61	4.4
37	169555	42	0.11170	0.00092	4.65193	0.26195	0.30204	0.01682	1827	7	1701	83	7.8
38	43779	44	0.07705	0.00086	1.93031	0.14312	0.18169	0.01332	1123	11	1076	72	4.5
39	14398	55	0.07244	0.00123	1.77575	0.10755	0.17779	0.01034	998	17	1055	56	-6.1
40	220862	67	0.11343	0.00092	4.99399	0.29453	0.31931	0.01865	1855	7	1786	90	4.2
41	96488	77	0.05546	0.00052	0.55121	0.03129	0.07209	0.00404	431	10	449	24	-4.3
42	330588	90	0.17209	0.00135	11.37042	0.83512	0.47921	0.03500	2578	7	2524	151	2.5
43	42266	74	0.07388	0.00086	1.71539	0.10949	0.16839	0.01057	1038	12	1003	58	3.6
44	81972	72	0.07891	0.00071	2.02097	0.11476	0.18575	0.01041	1170	9	1098	56	6.7
45	59516	88	0.07428	0.00074	1.72116	0.10604	0.16806	0.01022	1049	10	1001	56	4.9
46	30183	90	0.07166	0.00096	1.53737	0.09953	0.15559	0.00986	976	14	932	55	4.9
47	58800	83	0.07321	0.00086	1.61731	0.10366	0.16022	0.01009	1020	12	958	56	6.5
48	57235	80	0.18521	0.00159	11.82873	0.69321	0.46320	0.02685	2700	7	2454	117	11.0
49	74854	81	0.07433	0.00070	1.72201	0.10193	0.16803	0.00982	1050	9	1001	54	5.0
50	62055	75	0.07334	0.00077	1.60998	0.09326	0.15922	0.00907	1023	11	952	50	7.4
51	31741	64	0.06859	0.00136	1.41411	0.14716	0.14954	0.01527	886	21	898	85	-1.5
54	113694	93	0.05825	0.00068	0.69204	0.06753	0.08617	0.00835	539	13	533	49	1.2
55	112048	107	0.05543	0.00049	0.52567	0.04220	0.06878	0.00549	430	10	429	33	0.2
56	48093	114	0.05302	0.00071	0.48329	0.05099	0.06611	0.00692	330	15	413	42	-26.0
57	220975	121	0.07390	0.00062	1.61684	0.15782	0.15867	0.01543	1039	8	949	85	9.3
58	26797	133	0.06836	0.00122	1.33880	0.14747	0.14204	0.01544	879	18	856	87	2.8
59	158565	151	0.07335	0.00062	1.61952	0.19075	0.16013	0.01881	1024	9	957	104	7.0
60	24454	132	0.06340	0.00124	1.08876	0.16273	0.12456	0.01846	722	21	757	105	-5.2
61	91218	156	0.05471	0.00049	0.48359	0.03991	0.06411	0.00526	400	10	401	32	-0.1
63	73558	173	0.05405	0.00057	0.46318	0.04888	0.06215	0.00653	373	12	389	39	-4.3
64	60179	160	0.05446	0.00063	0.49787	0.04304	0.06631	0.00568	390	13	414	34	-6.3
66	246227	171	0.11404	0.00094	4.55340	0.45195	0.28958	0.02864	1865	7	1639	142	13.7
67	103188	154	0.05499	0.00056	0.49278	0.04903	0.06500	0.00643	412	11	406	39	1.4
68	110448	155	0.10083	0.00092	3.38450	0.29116	0.24345	0.02083	1639	8	1405	107	15.9
69	96818	140	0.05528	0.00056	0.49430	0.04357	0.06485	0.00568	423	11	405	34	4.5
70	24690	97	0.06985	0.00117	1.47255	0.14062	0.15291	0.01438	924	17	917	80	0.8
71	101143	38	0.09030	0.00086	2.94223	0.17492	0.23632	0.01387	1432	9	1368	72	5.0
74	111356	18	0.06978	0.00081	1.54330	0.09570	0.16042	0.00977	922	12	959	54	-4.4

Compton Formation - Milan member E 836025

Sample 09MILAN02 N 5109089

									age (Ma)				
Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
79	104596	19	0.19411	0.00183	12.81392	0.80612	0.47878	0.02978	2777	8	2522	129	11.1
83	1546378	41	0.09616	0.00073	3.25336	0.19574	0.24537	0.01464	1551	7	1415	75	9.8
84	173320	26	0.10926	0.00102	4.47719	0.27776	0.29720	0.01823	1787	9	1677	90	7.0
86	109564	20	0.08286	0.00102	2.46738	0.15533	0.21597	0.01333	1266	12	1261	70	0.5

Compton Formation - Milan member E 836025
Sample 09MILAN02a grain set N 5109089

Grain #	^{206}Pb (cps)	^{204}Pb (cps)	^{207}Pb/^{206}Pb	2 σ	^{207}Pb/^{235}U	2 σ	^{206}Pb/^{238}U	2 σ	age (Ma) ^{207}Pb/^{206}Pb	2 σ	^{206}Pb/^{238}U	2 σ	disc.
1	174650	65	0.05382	0.00058	0.45446	0.01360	0.06124	0.00171	364	24	383	10	-5.5
3	82728	58	0.05592	0.00078	0.55309	0.02980	0.07174	0.00373	449	31	447	22	0.6
4	33195	50	0.06897	0.00101	1.44147	0.05628	0.15159	0.00548	898	30	910	31	-1.4
5	185374	53	0.07251	0.00077	1.58698	0.05401	0.15873	0.00513	1000	22	950	28	5.4
6	67231	62	0.05567	0.00102	0.52728	0.02076	0.06870	0.00239	439	41	428	14	2.5
7	83622	62	0.05474	0.00063	0.51317	0.01823	0.06799	0.00229	402	26	424	14	-5.8
8	170770	48	0.05604	0.00061	0.56466	0.01886	0.07307	0.00231	454	24	455	14	-0.1
9	136843	40	0.05543	0.00061	0.52861	0.01926	0.06917	0.00240	429	24	431	14	-0.4
10	113854	60	0.05516	0.00064	0.55167	0.01922	0.07254	0.00238	419	26	451	14	-8.1
11	219035	38	0.05451	0.00056	0.48054	0.01475	0.06394	0.00185	392	23	400	11	-2.0
12	50823	164	0.06021	0.00173	0.61341	0.03075	0.07390	0.00304	611	62	460	18	25.7
13	159336	45	0.05901	0.00064	0.72690	0.02480	0.08934	0.00289	567	24	552	17	2.9
14	48710	37	0.05787	0.00075	0.64017	0.02050	0.08024	0.00235	525	29	498	14	5.4
15	48770	23	0.05446	0.00067	0.50974	0.01656	0.06789	0.00204	390	28	423	12	-8.8
16	110646	51	0.05562	0.00064	0.54231	0.01722	0.07071	0.00209	437	26	440	13	-0.7
17	28091	45	0.05420	0.00085	0.53464	0.01693	0.07155	0.00197	379	35	445	12	-18.1
18	135897	44	0.05400	0.00061	0.46154	0.01376	0.06199	0.00171	371	26	388	10	-4.6
19	416600	71	0.07252	0.00073	1.58240	0.05404	0.15825	0.00517	1001	20	947	29	5.8
20	79274	56	0.05640	0.00079	0.54315	0.01854	0.06985	0.00217	468	31	435	13	7.2
22	35528	76	0.06183	0.00194	0.77637	0.03468	0.09107	0.00290	668	67	562	17	16.6
23	195379	102	0.05546	0.00068	0.48225	0.01491	0.06306	0.00179	431	27	394	11	8.8
24	155106	56	0.07418	0.00076	1.70847	0.05378	0.16703	0.00497	1046	21	996	27	5.2
25	37730	65	0.07359	0.00096	1.67888	0.05596	0.16546	0.00507	1030	26	987	28	4.5
26	105935	69	0.05445	0.00059	0.47773	0.01629	0.06363	0.00206	390	24	398	12	-2.1
28	56068	68	0.05438	0.00068	0.48350	0.01840	0.06449	0.00232	387	28	403	14	-4.3
29	101814	79	0.07383	0.00083	1.69167	0.05660	0.16617	0.00524	1037	23	991	29	4.8
30	246354	84	0.11108	0.00113	4.64211	0.19107	0.30309	0.01209	1817	18	1707	60	6.9
31	83943	77	0.05500	0.00065	0.50616	0.01619	0.06675	0.00198	412	27	417	12	-1.1

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