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Magmatic Evolution and Crustal Accretion in the Early Proterozoic: The Geology and Geochemistry of the Narsajuaq Terrane, Ungava Orogen, Northern Quebec.

par

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Thèse presentée à la Faculté des études supérieures
en vue de l'obtention du grade de
Philosophiae Doctor (Ph.D.)
en géologie

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Faculté des études supérieures

Cette thèse intitulée:

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présentée par:
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a été évaluée par un jury composé des personnes suivantes:

[redacted]

[redacted]

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[redacted]

Thèse acceptée le: 23 octobre 1995

ABSTRACT

Evaluation of the geochemical evolution of the Earth involves the study and evaluation of various crustal compositions and mantle reservoirs throughout geological time. This present study focuses on rocks of Early Proterozoic age as the interval 1.8-2.0 Ga represented an important period of global continental growth and accretion. The Ungava Orogen of northern Quebec contains rocks that document Early Proterozoic rifting of a continental margin (the Superior Province), development of an oceanic basin and the eventual closure by subduction-related processes which generated the Narsajuaq terrane magmatic arc. More than 200 Ma of plate divergence and convergence is recorded within the tectonostratigraphic units of the Ungava Orogen.

The Narsajuaq terrane of the Ungava Orogen represents a mid-crustal section through an Early Proterozoic magmatic arc and contains four distinct plutonic suites: Older, Younger, Cape Smith and Late suites. The geochemical and isotopic signature of these plutons indicates an origin by subduction-related processes involving several different sources, namely juvenile depleted mantle, average 2.7 Ga Superior Province crust and old (3.1 Ga) Superior Province crust. In addition, the rocks of the arc itself may have served as a source for some of the younger granitic plutons.

Northward-directed subduction of a previously developed oceanic basin under a crustal fragment that marked the transition across the rifted Superior Province crust from ≈ 1.90-1.82 Ga generated the Older, Younger and Cape Smith suites of the Narsajuaq terrane. Accretion of the arc complex to the southern Superior Province occurred after 1.82 Ga, the age of the youngest arc-related pluton. Large sheets of granite associated with the Younger suite were intruded at ca. 1.80 Ga, near the end of arc accretion, which were derived by partial melting of the surrounding crust during tectonic thickening and metamorphic hydration of the crust. Continuing deformation and crustal thickening resulted in crustal anatexis and the generation of the Late suite granitic magmas at ≈ 1.76-1.74 Ga.

Evaluation of the mafic magmas of the Ungava Orogen documented the geochemical and isotopic composition of rocks associated with the rifting of the Superior Province and the development of an oceanic basin at ≈ 2.0-1.9 Ga. The spectrum of magma compositions erupted during this period have provided us with a window into the mantle and permitted us to define mantle reservoir compositions at 2.0 Ga for depleted mantle ($\varepsilon_{Nd} = +4.5$ to $+5.5$) and enriched mantle ($\varepsilon_{Nd} = +2.5$ to $+3.5$). Continental rifting may have been associated with, or contemporaneous with, the arrival of a mantle plume which generated the Povungnituk Group, as well as the Eskimo and Flaherty Formations. Melting within the plume could have produced the enriched mantle signature of these rocks with crustal contamination superimposed upon the Eskimo Formation and western Povungnituk Group samples. Complete rupture of the crust resulted in the generation of an oceanic basin into which the Watts Group, Chukotat Group and Ottawa Islands lavas were deposited. Most of these samples were derived from the depleted mantle reservoir although the Watts Group contains lavas having both depleted and enriched characteristics indicating the synchronous existence and tapping of these reservoirs.

The chemical and isotopic similarity of these Proterozoic magmas to modern-day magmas provides strong evidence that the interplay between depleted mantle, OIB (ocean island basalt) mantle and sub-continental mantle during the Proterozoic was comparable to that of the modern Earth, indicating a continuity in magmatic processes through time. The presence of these mantle reservoirs at 2.0 Ga also indicates that the geochemical and tectonic processes that generate such reservoirs were already in existence at this time, which suggests that crust-mantle recycling was active prior to 2.0 Ga.

RÉSUMÉ

L'évolution géochimique de la terre doit être abordée par l'étude de sections crustales et de réservoirs mantelliques de différentes compositions à différentes époques géologiques. Ce type de travail doit se concentrer sur l'évolution d'une suite de roches d'un âge particulier et la comparer ensuite avec des suites d'autres époques géologiques. Le travail présenté dans cette thèse de doctorat se focalise sur des séquences du Protérozoique inférieur (1.8 - 2.0 Ga). Cette époque représente une importante période d'accrétion et de croissance globale des boucliers continentaux. Les articles présentés dans cette thèse traitent de l'évolution magmatique, ainsi que de l'accrétion crustale à l'intérieur de l'Orogène de l'Ungava (nord du Québec) d'âge Protérozoïque inférieur.

Caractériser la composition de la croûte continentale est essentiel pour comprendre l'évolution géochimique de la terre. Cet aspect est l'objet du premier chapitre de la thèse, ce dernier décrivant la géologie et la géochimie du terrane de Narsajuaq. Ce terrane représente la racine d'un complexe magmatique d'arc du Protérozoïque inférieur soudé à la marge de la Province du Lac Supérieur vers 1.82 Ga. Le terrane de Narsajuaq est constitué de quatre suites plutoniques différentes couvrant à elles seules plus de 150 Ma d'activité magmatique.

Les plutons les plus âgés du terrane de Narsajuaq se trouvent dans la suite "Cape Smith". Celle-là contient des intrusions représentant plus de 60 Ma d'activité plutonique, de 1898 à 1839 Ma. Les plutons de la suite "Cape Smith" sont de composition dioritique, tonalitique et granitique, d'affinités essentiellement calco-alcalines, avec quelques échantillons tholéiitiques. Des enrichissements faibles à modérés de LILE (les éléments hygromagmatophile) et de terres rares, ainsi que des anomalies négatives de Nb et Ti caractérisent ces plutons. La signature isotopique de cette suite est bien définie avec ε_{Nd} initial = +3.2 à +1.5 et un rapport $^{87}Sr/^{86}Sr_i$ de 0.7020 - 0.7024. Les plutons de cette suite ont été formés par fractionnement d'une

source mantellique légèrement enrichie en LILE et terres rares légères, accompagnée d'une faible composante de contamination crustale attribuée à la subduction de sédiments. Ces plutons recoupent surtout la croûte océanique mafique des Groupes de Watts, Parent et Spartan.

La suite dominante du terrane de Narsajuaq, nommée "Older suite", est formée d'une séquence rubanée de diorite-tonalite-granite. Cette séquence présente des concentrations plus élevées et plus variables de LILE et de terres rares que la suite "Cape Smith". Les compositions isotopiques de la "Older suite" sont aussi plus variables que celles de la suite "Cape Smith", avec un ε_{Nd} initial de +4.0 à -10.7 et un rapport $^{87}Sr/^{86}Sr_i$ de 0.7025 - 0.7048. La "Older suite", mise en place de 1863 à 1844 Ma, est en partie contemporaine de la suite "Cape Smith". Il est postulé que la "Older suite" a été formée sur un fragment de croûte continentale archéene ayant interagi avec des magmas ascendants, produisant ainsi le large spectre de signatures géochimiques et isotopiques observé.

Une série de plutons homogènes de tailles kilométriques recoupe la "Older suite" et constitue la suite nommée "Younger suite". Ces plutons ont été mis en place principalement de 1836 à 1821 Ma, avec une série plus jeune se présentant sous forme d'épanchements granitiques très étendus datant de 1803 - 1800 Ma. La "Younger suite", composée de plutons de diorite, tonalite et granite, a une signature isotopique et géochimique similaire à la "Older suite". Cependant, les valeurs de ε_{Nd} initial et de $^{87}Sr/^{86}Sr_i$ varient plus largement dans la "Younger suite". Ces deux suites de roches semblent avoir une une pétrogénèse et une évolution globalement similaire. Une série distincte de plutons de monzodiorite située presque exclusivement dans la "Younger suite", contient des concentrations très élevées de LILE et de terres rares légères, et a des patrons de terres rares très fractionnés. La fusion partielle d'une source mantellique enrichie peut expliquer leur signature géochimique. Nous suggérons que la fusion partielle résulte de la délamination d'une plaque subductée, combinée à la remontée de manteau asthénosphérique. Ce dernier aurait été modifié par la subduction, après une inversion de la géométrie de la zone de subduction entre 1844 et 1836 Ma. Les plutons granitiques les plus jeunes de la "Younger suite" (IVUJ et CNF) contiennent de fortes concentrations de LILE et de terres rares légères. Ils ont des compositions isotopiques qui indiquent un apport significatif d'une composante crustale (ε_{Nd} = -4.9 à -6.2 (IVUJ) et -17.5 à -18.5 (CNF)). Ces granites furent mis en place durant les étapes tardives de l'accrétion de l'arc magmatique sur la marge de la Province du Lac Supérieur. Leur génération est attribuée à la fusion partielle de la croûte, causée par l'épaississement tectonique et par l'influx de fluides au métamorphisme rétrograde subi par la croûte.

La suite nommée "Late suite" se compose d'une série de dykes de syénogranite pegmatitique non-métamorphisés et non-déformés. Ceux-ci recoupent toutes les unités de l'orogène ainsi que de rares plutons granitiques situé essentiellement dans le terrane de Narsajuaq. Cependant, il est à noter que seul un pluton granitique recoupe toute les unités tectono-stratigraphiques. Les dykes de pegmatite ont été mis en place il y a 1758 Ma, alors que les plutons granitiques semblent légèrement plus jeunes, un pluton ayant été daté à 1742 Ma. Une étude géochronologique détaillée de ce pluton (pluton du Lac Duquet) est l'objet du chapitre 4. Cette étude, effectuée à l'aide de la technique de l'ablation par laser ICP-MS, a permis l'identification de zircons hérités, ayant des âges de 1.7 à 3.2 Ga. Plus de 80% de ces âges correspondent à ceux des roches environnantes. Cette étude géochronologique illustre le potentiel de la méthode d'analyse de l'ablation par laser ICP-MS pour déterminer l'histoire crustale d'un pluton granitique. Les donnés géochimiques, isotopiques et géochronologiques des plutons de la "Late suite" indiquent que l'anatexie crustale est le principal mécanisme ayant produit ces plutons.

Une section crustale composite de 15-20 km d'épaisseur du terrane de Narsajuaq est présenté. Les roches plutoniques et sédimentaires de Narsajuaq représentent une section de profondeur intermédiaire d'une croûte d'arc magmatique du Protérozoique inférieur. La partie supérieure de la section crustale est manquante. Cependant, les roches volcaniques du Groupe de Parent pourraient représenter cette dernière. Les

métasédiments du Groupe de Sugluk, ainsi que les grandes intrusions granitiques, sont plus abondants vers le sommet de la section crustale, alors que les plutons mafiques sont plus communs vers la base de la section. La détermination de la composition géochimique composite du terrane de Narsajuaq est en cours. Le but est de comparer la composition globale de cette croûte du Protérozoïque inférieur avec des sections de croûtes archéennes et phanérozoïques, afin d'évaluer les variations géochimiques séculaires, au cours du Précambrien principalement, et un particulier à la transition Archéen-Protérozoïque. Une décroissance du rapport $(La/Yb)_n$ et un appauvrissement global des terres rares lourdes documentées dans les roches post-Archéennes sont corroborées par les données de la présente étude. Ces variations résulteraient d'un changement fondamental du processus de génération magmatique, allant de la fusion de plaques subductées jusqu'à la fusion du manteau sus-jacent, conséquences du refroidissement séculaire de la terre. Nous désirons poursuivre l'investigation de ces variations géochimiques temporelles pour pouvoir, entre autres, évaluer la signification de la transition Protérozoïque-Archéen.

Une évaluation des réservoirs mantelliques présents durant la formation des roches de l'Orogène de l'Ungava est l'objet du chapite 3. Le rifting du socle de la Province du Lac Supérieur, vers 2.04 Ga, a conduit au développement d'une séquence de marge de rift volcanique et d'un bassin océanique caractérisés par de volumineux épanchements magmatiques. Quatre suites distinctes de roches volcaniques furent mises en places durant plus de 100 Ma, reliées à la présence de différents manteaux et réservoirs crustaux (manteau appauvri, manteau enrichi et croûte continentale). Les suites associées avec le rifting continental et le développement d'une marge volcanique incluent: (1) les basaltes continentaux contaminés par la croûte de la Formation Eskimo et du Groupe de Povungnituk occidental et central; (2) les laves mafiques de la Formation de Flaherty, du Groupe de Povungnituk oriental et certaines laves du Groupe de Watts qui ont une signature géochimique et isotopique légèrement enrichie; et (3) les laves alcalines de Povungnituk, hautement enrichies. Durant le développement d'un bassin océanique, un quatrième groupe a été mis en place, comprenant des basaltes

magnésiens et des basaltes komatiitiques du Groupe de Chukotat, Iles d'Ottawa et quelques échantillons du Groupe de Watts qui ont des compositions géochimiques et isotopiques similaires aux N-MORB.

La gamme de compositions magmatiques des roches mises en place durant l'événement de rifting et de génération du bassin océanique, procure une fenêtre du manteau à 2.0 Ga et nous permet de définir les compositions des réservoirs mantelliques: pour le manteau appauvri (ε_{Nd} = +4.5 à +5.5); et pour le manteau enrichi (ε_{Nd} = +2.5 à +3.5). Le rifting continental peut être associé, ou être contemporain, à l'arrivée d'un diapir mantellique ayant donné naissance au Groupe de Povungnituk ainsi qu'aux formations d'Eskimo et de Flaherty. La fusion à l'intérieur du diapir mantellique pourrait avoir produit la signature de manteau enrichi des roches, avec contamination crustale superposée à la formation d'Eskimo et au Groupe de Povungnituk occidental. La rupture complète de la croûte a abouti à l'ouverture d'un bassin océanique, dans lequel les Groupes de Watts, de Chukotat et les Iles Ottawa ont été déposés. La plupart de ces roches ont été dérivées du réservoir mantellique appauvri. Cependant le Groupe de Watts contient des laves ayant des caractéristiques de manteau appauvri et enrichi, indiquant l'existence synchrone de ces réservoirs.

Un réservoir additionnel qui a servi de source au magmatisme de type arc a aussi été documenté, sur la base des compositions géochimiques et isotopiques des magmas les plus primitifs du terrane de Narsajuaq. Ce réservoir a une composition isotopique de Nd similaire au manteau appauvri. L'étude plus approfondie des différents magmas du terrane de Narsajuaq, présentée au chapitre 2, suggère qu'un réservoir mantellique enrichi a aussi été impliqué dans la génération des magmas de type arc, c'est à dire dans la formation des monzodiorites de la "Younger suite". Cependant, il est postulé que cet enrichissement était de nature locale, directement relié aux processus de subduction actifs en ce temps.

Les similarités chimiques et isotopiques entre les magmas protérozoïques étudiés ici et les magmas contemporains suggèrent que l'interaction entre manteau appauvri, manteau OIB et manteau sub-continental pendant le Protérozoïque est semblable à la situation actuelle. Les processus magmatiques sont donc continus au cours du temps. L'existence de ces réservoirs mantelliques il y a 2.0 Ga, indique de plus que les processus tectoniques et géochimiques responsables de leur origine existaient déjà en ce temps. Le recyclage manteau-croûte aurait donc été actif il y a plus de 2.0 Ga.

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ACKNOWLEDGEMENTS

The work presented in this thesis was possible due to the help, cooperation and encouragement of a number of people. First of all, I would like to thank my thesis director, John Ludden, for all his help during the course of my work, particularly for all the discussions which helped me develop my ideas, not only about the subject of study, but also at a larger, more global scale, and to see the implications of my work. I would also like to warmly thank Marc St-Onge and Steve Lucas of the Geological Survey of Canada, who provided the opportunity to study the rocks of the northern Ungava Peninsula through participation in their GSC Ungava project. The field experience was invaluable in understanding the origin of these rocks, and the discussions with both of them shed much light on the geological meaning of the rocks, and the Ungava Orogen as a whole. Field mapping as part of their team was an excellent experience and although Mother Nature did not always provide the best working conditions, Marc and Steve did their best to make the field seasons most agreeable and fun! From an analytical perspective many people kindly provided me with their expertise ensuring that my data was the best quality possible and teaching me the "ins and outs" of isotopic analyses. I therefore thank the following people for their help: Brigitte Dionne and Gilles Gauthier (Université de Montréal), Richard Stern and Randy Parrish (Geological Survey of Canada), and Ross Stevenson and Steve Previc (GEOTOP, Université du Québec à Montréal). Randy Parrish and Ross Stevenson are thanked for their generosity in allowing me to perform isotopic analyses in their labs. Discussions with many people have helped me develop and refine my ideas and I wish to thank (in addition to John Ludden, Marc St-Onge and Steve Lucas, noted above) Richard Stern, Paul Bédard, Randy Parrish, Don Francis, John Stix, Louis Gauthier, Dave Scott and Lisa Campbell. I wish to also thank all my fellow grad students and staff at the Université de Montréal for their help and support during my studies, in particular, Shirley Peloquin (for the "TBS" sessions) and Anne Sherman (for the gab and food sessions). I also thank most immensely Louis Gauthier for all his support, help, understanding and love from the very beginning of my work, and for putting up with talk about rocks that don't have any mines in them! Finally, thanks to my parents and Jacques Guigne, who believed in me and supported me all through my academic studies and gave me the confidence to accomplish anything and everything.

for my parents

INTRODUCTION

0.1 Scope of thesis

0.1.1 *Rationale*

The geochemical evolution of the Earth has been a subject of study by many researchers who have approached this topic from various angles, involving different techniques (Sm-Nd, Rb-Sr, Lu-Hf and U-Pb isotopic studies, as well as trace element studies) on a variety of rock types such as mafic to ultramafic basalt and komatiite flows, mantle xenoliths, granitic differentiated plutons and sedimentary rocks of varying ages. For example, the chemical differentiation of the crust from the mantle was studied by Hofmann (1988) who proposed a model whereby early separation of the bulk of the continental crust was followed by rehomogenization of the residual mantle and that the present-day mantle heterogeneities were developed at a latter stage. Taylor and McLennan (1985) used the composition of sedimentary rocks derived from weathering and erosion of exposed crust to determine average crustal compositions through time, and documented systematic changes such as enrichments in the LILE's, development of negative Eu anomalies and depletion in Ni, Co, Zr and Nb in post-Archean rocks relative to Archean crust. Martin (1986) and Drummond and Defant (1990) have interpreted geochemical changes across the Archean-Proterozoic boundary as a reflection of the change from subducted slab to mantle wedge melting in subduction zones where trondhjemite-tonalite-dacite (TTD) suites are generated. Bennett et al. (1993) documented the existence of a transient highly depleted mantle reservoir early in Earth's evolution based on the presence of Early Archean (3.9-3.8 Ga) rocks having high initial ε_{Nd} values ($\approx > +4$), while the evolution of enriched mantle reservoirs through time was studied by Campbell and Griffiths (1992) who noted a change in the geochemical signature of komatiites from depleted to enriched, related to changes in the style of mantle convection at approximately 4.0 Ga. It is clear therefore that important geochemical changes have occurred in the geological past. However, many studies have based their conclusions on data from recent and Archean suites, while there has been a paucity of data for suites at the critical boundary of Early Proterozoic age.

Other studies have looked at the present-day distribution of crust and noted the presence of significant quantities of Late Archean (3.0-2.7 Ga) and Early Proterozoic (2.0-1.7 Ga) rocks on the Earth's surface (e.g., Patchett and Bridgwater 1984; Nelson and DePaolo 1985; Patchett and Arndt 1986; Jacobsen 1988), which appear to represent important periods of crustal growth during Earth's evolution. Such growth can occur through a variety of mechanisms including magma addition, crustal accretion and tectonic stacking. The existence and acceptance of such processes as real and viable mechanisms in modern environments is firmly entrenched, where spectacular evidence for subduction, arc volcanism, accretion and collision is occurring (e.g., Aleutians, Indonesia, Japan). The nature of plate tectonics in the past, however, is not as clearly evident. Archean tectonics is believed to have involved thicker oceanic crust, rapid subduction and turbulent convection due to the higher geothermal gradient at that time (Sleep and Windley 1982). The amalgamation and accretion of different domains into the distinct granite-greenstone cratonic units that typify many Archean blocks is believed to be related to a precursory form of plate tectonics (e.g., Card 1990; Kimura et al. 1993). One of the oldest examples of an orogenic belt where unequivocal evidence for modern-style Wilson-cycle plate tectonics has been documented is the Early Proterozoic Ungava Orogen (Fig. 1) of northern Quebec where continental rifting, opening of an oceanic basin, subduction and arc magmatism, continental collision and the development of a thrust-fold belt accompanied by ophiolite obduction has been documented (e.g., Hynes and Francis 1982; Hoffman 1985; Picard et al. 1990; St-Onge and Lucas 1990a; Scott et al. 1991; Lucas et al. 1992).

The Ungava Orogen (Fig. 1) contains a well exposed series of rocks of Early Proterozoic (2.0-1.8 Ga) and Archean age that offers a superb opportunity to study crust-mantle evolution and tectonics during this period in Earth's development. One segment of the orogen in particular, the Narsajuaq terrane (Fig. 1), represents a mid-crustal section through an accreted Early Proterozoic magmatic arc complex (Dunphy and Ludden 1992; St-Onge et al. 1992), and investigation of the composition and origin of these rocks will help constrain the overall problem of crustal evolution during this time.

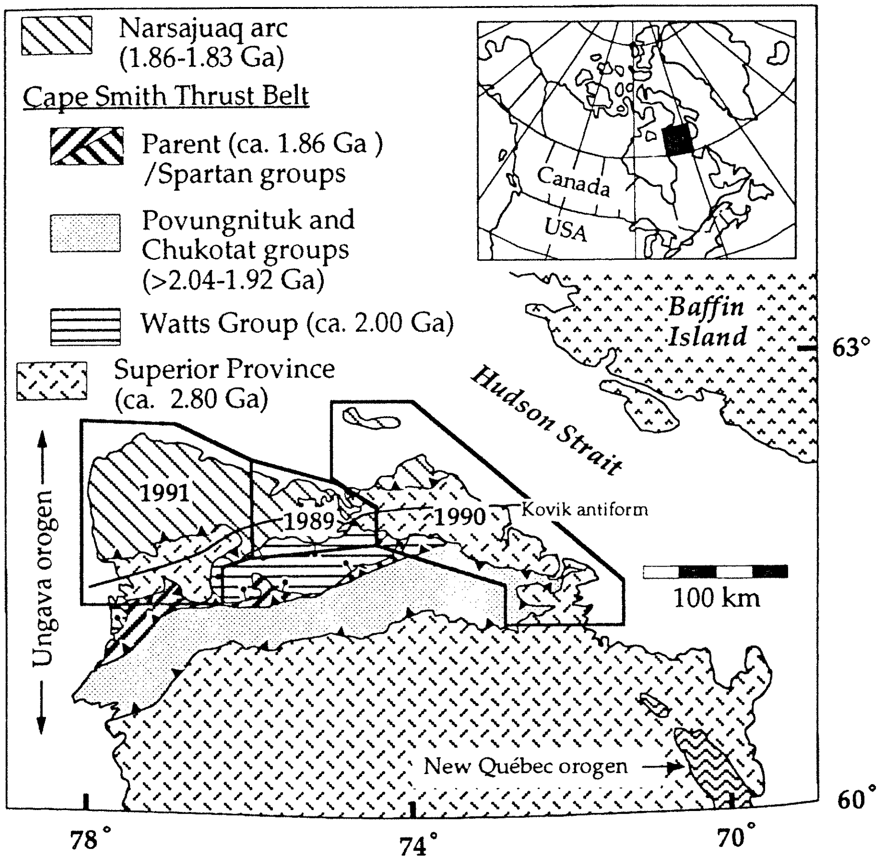

Figure 0.1. Location map of the Ungava Orogen and its tectonostratigraphic elements (adapted from St-Onge and Lucas 1992). Outlined areas mark the extent of mapping for the three summers of the project.

Characterization of the bulk composition of these rocks will permit a better documentation of the geochemical changes from the Archean to the Proterozoic. Another aspect of crustal evolution that can be addressed through study of this Early Proterozoic orogen is the evolution of mantle reservoirs through time. As was earlier stated, data for rocks of this age is scarce, and an evaluation and comparison of the composition of the mantle reservoirs that were involved in magma genesis at this time ($\approx$ 2.0 Ga) with younger reservoirs can shed light on this problem.

Although the northern Ungava Peninsula has been the focus of geological investigations for over a century (see historical background, below), most work has focused on the region with the greatest mineral potential, the Cape Smith Belt (Fig. 1). In 1989 the Geological Survey of Canada (GSC) decided to focus their resources on the region north of the Cape Smith Belt, and a multidisciplinary project was begun under the leadership of Dr. M.R. St-Onge and Dr. S.B. Lucas. The objectives of the GSC study were several (St-Onge and Lucas 1990b): (1) to provide regional-scale (1:100 000) geological coverage of this area, complimenting the maps published by the Geological Survey for the eastern Cape Smith Belt and the Ministère de l'Énergie et des Ressources du Québec (MERQ, now Ministère des Ressources Naturelles) for the western part of the belt (see historical background section for references); (2) to assess the potential for economic mineralization within the region; and (3) to strive towards an understanding of the geological evolution of the area through thematic studies on the structure, metamorphism, geochronology, igneous petrology and geochemistry of the various geological units in the area.

Discussions between myself, John Ludden, Marc St-Onge (GSC) and Steve Lucas (GSC) about the GSC project and the geological problems presented by the rocks of the Ungava region led to the decision to participate in the GSC project as a PhD student in order to investigate and address some of the above mentioned geological problems. Prior to the initiation of the GSC project in 1989, the only geological maps for this region of the Ungava Peninsula were based on reconnaissance scale mapping (1:250 000) undertaken in 1973 by Taylor (1974, 1982). This mapping indicated the presence of both Archean and Proterozoic (Aphebian) rocks north of the Cape Smith Belt, primarily granitic gneissic rocks with extensive swaths of paragneiss. Granulite-grade rocks were mapped as a separate rock type ("granulite") located in two geographically restricted areas. Several small felsic intrusives were also identified. Taylor's mapping identified most of the major units and geological contacts of the region, but the scale of mapping was such that geological observations were made only once every ≈5 km in order to cover the entire map area (≈50 000 km^2) during the course of two months in the summer of 1973. This scale of observation precluded detailed documentation of the geology of the region which restricted geological interpretations and models. No geochemical analyses were available for the rocks located north of the Cape Smith Belt, and the only geochronology consisted of Rb-Sr dates, all of which indicated Archean ages (Beall et al. 1963; Doig 1983, 1987).

0.1.2 *Goals*

During the course of the field studies and subsequent laboratory-based analyses the main subject of study for the thesis was more fully defined and the principal broad goal was delimited: to document the composition, age, origin and tectonic significance of the rocks located north of the Cape Smith Belt. Study was restricted to the Early Proterozoic rocks of the Narsajuaq terrane (Fig. 1), as another student was to undertake a MSc thesis on a suite of the Archean rocks (Harvey 1995). The main goal may be broken down into several detailed objectives: (1) to aid in documenting the geological character of the Narsajuaq terrane through field studies and regional scale mapping of the northern Ungava Peninsula in conjunction with the Geological Survey of Canada; (2) to provide a detailed geochemical characterization of this Early Proterozoic crust which has been interpreted as a magmatic arc; (3) to undertake isotopic studies to determine its source and evolution through time; and (4) to provide a detailed geochemical and petrological composite crustal section of the Narsajuaq arc for comparison with other crustal sections of varying ages in order to document the evolution of crustal compositions and crustal formation processes through time. These goals have been addressed in the first paper of the thesis, presented in Chapter 1.

As an introduction to the process of geochemical, isotopic and geochronological data acquisition that was meant as a preparation for the work ahead, a smaller-scale study was proposed to determine the age, origin and evolution of a small granitic pluton (the Lac Duquet pluton) that was part of a suite of undeformed and unmetamorphosed intrusions in the Narsajuaq terrane. The Lac Duquet monzogranite was chosen as the best candidate for this study as the geological context was well constrained due to detailed mapping by the author in 1990. It is also the only pluton that cross-cuts a D_4 fold containing all tectonostratigraphic units, thereby constraining the timing of this youngest deformation event. Conventional U-Pb geochronology (TIMS - thermal ionization mass spectrometery) constrained the age of emplacement of the pluton and documented the presence of a significant amount of inherited zircon (Dunphy et al. 1991), which given its geological setting and location was not surprising, while the geochemical and isotopic data indicated a crustal origin. It was subsequently decided that further documentation of the inheritance history of the pluton would better enable us to constrain its source and that laser ablation ICP-MS analysis (inductively coupled plasma mass spectrometry) was the most suitable technique for this job. It had previously been documented that the ICP-MS technique had the potential to produce meaningful ages ($^{207}Pb/^{206}Pb$ ages) for geological samples; however, it had not yet been fully evaluated using a case study. The Lac Duquet pluton was an ideal example as its age of emplacement and origin were constrained by the earlier work. The original objectives for the study of the Lac Duquet pluton were met, and these results along with the new geochronological ICP-MS data have been presented for publication (Dunphy et al. in press, Chapter 3).

Another aspect of study for the thesis was to document the composition of the various mantle reservoirs that were present in this area at the time of formation of the rocks, at approximately 2.0-1.8 Ga. The presence of distinct magmatic suites of rocks such as alkaline volcanics (Picard et al. 1990; Gaonac'h et al. 1992), tholeiitic basalts

(Francis et al. 1983; Picard et al. 1990; Scott et al. 1991) and arc volcanics (Picard et al. 1990) and plutonics (Dunphy and Ludden 1992; St-Onge et al. 1992) indicated the existence of different mantle reservoirs (e.g., OIB-like, MORB-like) in this region. It has only been recently, over the past 10 years, that a sufficient body of geochemical data has existed for the rocks of the orogen to permit an evaluation of these possible reservoirs and their evolution through time. Most previous geochemical and petrological studies on these rocks involved the presentation and discussion of a small amount of data from a relatively restricted area. Although the authors of these studies made attempts to integrate their results into the context of previous studies, no one had yet synthesized all the available data and made interpretations based on a large amount of data at the scale of the orogen. This was now possible due to a compilation of all existing geochemical data for the igneous rocks of the orogen by myself and resulted in the development of a model for the geochemical evolution of mafic magmas within the orogen (Dunphy et al. 1995, Chapter 2).

Finally, a fourth study was initiated as a result of the generosity of the Ministère des Ressources Naturelles du Québec (MRN) who invited me to evaluate the geological data that they had accumulated for a series of plutons intrusive into the Cape Smith Belt lithologies. I had collected samples and data for these rocks during the course of mapping with the GSC and planned to integrate my data into the bigger Narsajuaq picture, but welcomed the chance to examine more data in order to better determine the origin of these rocks and their relationship to the other elements in the Ungava Orogen. The results of this study was subsequently published as a MRN report (Dunphy 1994, Chapter 4).

0.2 Structure of the thesis

The thesis has been divided into 4 main chapters which present papers discussing various aspects of the geological evolution of the Ungava Orogen with chapters of introduction and conclusion at the beginning and end of the thesis. This present **introductory chapter** outlines the scope and goals of the thesis and the rationale behind its initiation. A detailed description of the scientific methods used to obtain the data and the means used for presentation and diffusion of the results is also presented. An overview of the previous geological studies that have been undertaken in the area is included with a detailed documentation of the most recent studies. Finally a brief discussion of the general geological setting of the Ungava Orogen is presented in order to give the reader a framework of reference for the rest of the thesis.

Chapter 1 presents a detailed description of the geology and geochemistry of the Narsajuaq terrane as well as a petrogenetic interpretation and a geotectonic model for its origin and evolution. Comparison of the Narsajuaq terrane to other large plutonic terranes such as the Peninsular Ranges batholith and various other granitoid complexes along the western margin of North and South America is made, as well as comparisons to other plutonic and volcanic arc complexes situated within the same orogenic belt as the Narsajuaq terrane (the Trans-Hudson Orogen). The data and results of Chapter 1 are to be published as two separate papers that are currently in preparation. The main body of information pertaining to the geology, geochemistry, petrogenesis and geotectonic model for the Narsajuaq terrane will be submitted to the Journal of Petrology in the fall of 1995. A second paper presenting the Narsajuaq terrane as an excellent example of a crustal section through an Early Proterozoic magmatic arc, as well as comparing and contrasting its make-up and composition to other crustal sections of different ages, is also in preparation for eventual submission to Precambrian Research for spring 1996. Both of these papers are to be authored by myself and J. Ludden. All of the data for these articles was collected by myself and the papers have also been written by myself. Field visits and discussions with J. Ludden greatly contributed to the development of the ideas presented, and it was the incentive of Ludden to present the Narsajuaq terrane as a crustal section. Evaluation and review of the articles by Ludden resulted in significant improvements.

The emphasis of **Chapter 2** is a regional-scale synthesis of the geochemistry of the mafic magmas of the Ungava Orogen and the implications of these results for mantle reservoir compositions at 2.0 Ga. This comprehensive review of the available data was able to document the existence of, and place constraints on, the composition of various mantle reservoirs that were involved in the petrogenesis of the magmas of the Ungava Orogen. Comparison of the composition of these Early Proterozoic reservoirs to modern mantle components illustrates their similarity and points to the continuity of magmatic processes through time. The paper presented in Chapter 2 has been published in Chemical Geology (Dunphy et al. 1995). For this paper most data was compiled from existing publications while some data was provided by the present author with other unpublished data coming from D. Francis and J. Ludden. Considerable discussion with both J. Ludden and D. Francis developed the ideas presented in the paper, which was written by myself and subsequently reviewed and revised by the coauthors.

The application of different geochronological techniques to the evaluation of the age of emplacement and inheritance history of a late granitic intrusion is the focus of **Chapter 3**. Two techniques are presented: conventional TIMS and the newer ICP-MS laser ablation methods. Both techniques have their strengths and weaknesses and together they provide a powerful means of deducing the geological history of the pluton. The age of emplacement of the pluton was constrained and the extensive zircon inheritance was documented. This paper has been submitted to the Canadian Journal of Earth Sciences and is expected to be published later this year (Dunphy et al. in press). The U-Pb geochronology for this paper was partly performed by myself under the supervision of R. Parrish, who completed the analyses. Collection of the ICP-MS data and the geochemical data was entirely done by the myself. Considerable discussion with Parrish about the TIMS geochronology technique as well as the limits and merits of other techniques was indispensable. J. Ludden contributed greatly to detailed discussions about the significance and origin of the pluton. I wrote the paper and it was subsequently critically reviewed by both Ludden and Parrish.

The last paper presented in the thesis (**Chapter 4**) discusses the geology and geochemistry of a series of plutons that are intrusive into the Cape Smith Belt lithologies. This paper was written as a contract for the Ministère des Ressources Naturalles du Québec (MRN) using the samples and data provided by the MRN (Dunphy 1994). The origin of these plutons having calc-alkaline characteristics in the predominantly mafic oceanic crust of the Watts Group as well as in the Parent and Spartan Groups was unknown and puzzling. Study of the data indicated an arc origin for these samples and that they were probably related to the arc magmas that generated the plutons of the Narsajuaq terrane and the volcanics/volcaniclastics of the Parent Group. Some of the data presented in the MRN paper (Dunphy 1994) is integrated with my own data for this suite of plutons and is presented in the paper of Chapter 1 as the Cape Smith suite. The data presented in the MRN paper was entirely provided by the MRN but was interpreted and written up solely by myself.

A final chapter (**General discussion and conclusions**) summarizes the major results of these above studies and discusses the possible future work that could be done to further improve our knowledge and understanding of the geological evolution of the Narsjauq terrane and the Ungava Orogen. A brief discussion of the regional-scale tectonics of the supercontinent Laurentia and the role of the Narsajuaq terrane in its development is also presented.

0.3 Methodology

In order to attain the aforementioned goals a detailed plan of action was drawn up with modifications added as conditions warranted. The fundamental foundation or base for this thesis is the regional-scale (1:100,000 scale) geological mapping that was undertaken by the Geological Survey of Canada during the course of three summers (1989-1991, Fig. 1). A total of approximately 20 000 km^2 was mapped during this period. Each summer one base camp was set up and from this camp daily traverses were undertaken. The base camps were located along the Rivière Foucault in the Vallée Narsajuaq (≈ 75°31'E, 62°01'N) near the village of Salluit (1989), at the mouth of the Rivière Wakeham (≈ 72°16'E, 61°34'N) near the village of Kangiqsujuaq (1990) and along the northwestern coast of the peninsula (≈ 77°43'E, 62°27'N) close to Ivujivik

(1991). Field work generally began by late June and terminated by the end of August due to the deterioration of weather conditions.

The strategy of the mapping program was to perform systematic and detailed solo foot traverses (perpendicular to the regional geological strike) 12-18 km in length and separated from each other by an average of 3-4 km. Helicopter support during the full three summers was vital to this type of work, which consisted of early morning drop-off's and late afternoon/early evening pick-ups. The northerly latitude facilitated the work as extended daylight hours permitted long and fruitful days. Extrapolation between traverses was generally possible due to the excellent exposure (≈ 70-80 %) of the rocks which assisted the interpretation. In areas where such correlations were not possible extra traverses or helicopter sampling was added. Daily compilation and group discussion of the geological data was undertaken which greatly aided in the interpretation of the data.

I was a principal member of the mapping team during the three summers of the project, and the geological knowledge obtained in the field was indispensable. Although this study would have still been possible had I not participated directly in the mapping program, the benefits of participation were numerous. A much better knowledge of the rocks was achieved, as well as a greater understanding of the geological relationships between the various lithological units within the orogen. Better control over the sampling procedure and the types of samples was possible. Many stimulating discussions and conversations about the geological data occurred during the course of the mapping and resulted in continuous development and refinement of thoughts and ideas. Practical experience related not only to the geological mapping but also to setting up and maintaining a field camp was also invaluable. And of course, the comradery and friendships, both professional and personal, that developed during this period have been maintained and will continue long in the future.

All geological samples that were collected are the property of the Geological

Survey of Canada and upon returning to the university samples for geochemical analysis were collected from the GSC and brought to Montreal. Geochemical samples were chosen with several goals in mind: representivity of the rock type; lack of alteration; geographic distribution; and availability. A relatively uniform geographic distribution of the samples was not always possible due to the primary desire to obtain samples that were truly representative of the rock type in question. In addition, the lithological distribution was such that certain compositions were only found in particular areas, hence a large geographical distribution was not possible. Lack of alteration is always important, if not vital, for proper and precise geochemical analyses, and in most cases it was not a problem to find appropriate samples as the outcrops were generally fresh. Recent glaciations helped expose such outcrops and the annual frost-and-thaw cycle provided angular blocky fragments that were perfect for sampling. The tundra environment with sparse vegetation also facilitated sampling, as did the fact that snow covers the area for 8-10 months a year, inhibiting growth of most plants and some lichens.

Sample preparation for geochemical analysis follows standard procedures of crushing and powdering (the specific preparation techniques for the particular analysis are detailed in Appendix 2). Most samples were prepared by the author although some batches were contracted out to undergraduate students. Geochemical analyses were preformed by X-ray Fluorescence (XRF) at the Université de Montréal and at McGill University and by instrumental neutron activation analysis (INAA) at the Université de Montréal, with some samples being run by myself, although most were run by technicians. XRF data was processed and presented to me in final form whereas the INAA data was processed by myself. Preparation of samples for isotope geochemistry and the subsequent analysis on thermal ionization mass spectrometers (TIMS) were entirely done by myself. All samples for isotope analysis were prepared (techniques in Appendix 1) at the Université de Montréal. Nd isotopic analyses were performed at the Geological Survey of Canada (Ottawa) and Sr isotopic analyses were done at the Université de Montréal and at GEOTOP at the Université de Québec à Montréal.

Mineral separates for geochronological analyses were prepared at the Geological Survey of Canada. TIMS analyses were also undertaken at the GSC while laser ablation ICP-MS analyses were performed by myself at the Université de Montréal.

Following the physical gathering of the samples and data, the results and interpretations were written up and presented. Some of the results have already been published in as papers in journals or reports (Dunphy 1994; Dunphy et al. 1995; Dunphy et al. in press) and at least two more papers are in preparation (Chapter 1). Results were also presented at talks and poster sessions at various conferences around the world which was a valuable forum for discussion and development of ideas: GAC-MAC 1991 in Toronto (Dunphy et al. 1991), AGU Spring meeting 1992 in Montreal (Dunphy and Ludden 1992), IVACEI 1993 in Canberra, Australia (Dunphy et al. 1993), ICOG-8 1994 in San Francisco (Dunphy and Ludden 1994). In addition, several other presentations have been made integrating the results of this present study with that of another project which I am involved in (Dunphy et al. 1994; Dunphy and Skulski 1995a, 1995b).

0.4 Historical overview of geological studies in the Ungava Orogen

The Ungava peninsula of northern Quebec has long been the subject of geoscientific investigations, beginning with observations by Bell (1885) and Low (1899, 1902) for the Geological Survey of Canada in the areas of Hudson Strait and Hudson Bay in the late 19th and early 20th century. These early studies describe many different facets of the Earth Sciences including geology, geography, hydrology, climatology, as well as biology and the natural history of the region. Early mineral exploration in the Cape Smith Belt began in the 1930's (Gunning 1934) and was followed by the discovery of nickel and other base metals as well as an asbestos deposit in the 1950's. The asbestos deposit (Asbestos Hill) was mined from 1972 to 1984 by the Asbestos Corporation Limited, and over the years since first being staked in 1961, a nickel deposit (Raglan horizon) was evaluated and is finally going into production under the ownership of Falconbridge Limited. Significant concentrations of the platinum-group elements (PGE's) are associated with the Ni-Cu deposits (Giovenazzo 1986; Giovenazzo et al.

1989; Barnes and Barnes 1990; Barnes and Giovenazzo 1990).

The early mineral exploration created significant interest in the region. Subsequent mapping of the Cape Smith Belt by a variety of workers (Beall 1959, 1960; Bergeron 1957, 1959; DeMontigny 1959; Gélinas 1962; Gold 1962; Stam 1961) delimited major structures and was followed by regional reconnaissance mapping of the entire Ungava Peninsula by Taylor in 1973 (Taylor 1974, 1982). Other studies of the Cape Smith Belt of a stratigraphic and geochemical nature were undertaken by Dimroth et al. (1970), Baragar (1974, 1984), Schwarz and Fujiwara (1977), Francis and Hynes (1979), Baragar and Scoates (1981), Hynes and Francis (1982), and Francis et al. (1983). Rb-Sr geochronology on rocks from within and surrounding the belt was undertaken by Beall et al. (1963) and Doig (1983, 1987). Hoffman (1985) reviewed all available geological data for the region and presented a tectonic model proposing that the Cape Smith Belt was a klippe.

Subsequent detailed mapping by the Ministère des Ressources Naturalles du Québec (MRN) in the western and central portions of the Cape Smith Belt (Lamothe et al. 1983, 1986; Hervet 1985; Moorhead 1985, 1987, 1988, 1989; Tremblay 1987, 1989, 1991; Barrette 1988, 1989, 1990; Picard 1989a, 1989b; Roy 1989) and the Geological Survey of Canada (GSC) in the eastern region (St-Onge and Lucas 1986, 1989-1990, 1990a, 1993; St-Onge et al. 1986, 1987, 1988, 1989) produced a wealth of information on the geological field relationships within the Belt. Complimentary thematic studies on the structure and metamorphism (Lucas 1989, 1990; Lucas and St-Onge 1989a, 1989b; Lucas and Byrne 1992; Bégin 1989, 1992a, 1992b; St-Onge and Lucas 1991), petrology and geochemistry (Gaonac'h et al. 1989, 1992; Hegner and Bevier 1989, 1991; Scott et al. 1989, 1991, 1992; Thibert et al. 1989; Picard et al. 1990; Scott and Bickle 1991), geochronology (Parrish 1989; Machado et al. 1993) and mineralization (Giovenazzo 1986; Giovenazzo and Lefebvre 1986; Giovenazzo et al. 1989; Barnes and Giovenazzo 1990; Barrette 1991) has led to a better understanding of the geological evolution of the fold-thrust belt.

Most recently a three year multidisciplinary project was initiated by the Geological Survey of Canada under the leadership of M.R. St-Onge and S.B. Lucas to investigate the hinterland of the Cape Smith Belt in the Ungava Orogen. Many different publications have resulted from this project since its initiation during the summer of 1989. Presentation and documentation of the field geological observations have been published as a series of Geological Survey of Canada Current Research papers (St-Onge and Lucas 1990b, 1992; Lucas and St-Onge 1991) and were followed by publications presenting more detailed information and interpretations, as well as an updated geotectonic model for the origin of the orogen incorporating the new geological data (Lucas and St-Onge 1992; Lucas et al. 1992; St-Onge et al. 1992). Subsequent structural studies have detailed the nature of basement thrusting during the development of the orogen (Lucas and Byrne 1992), the origin of the compositional layering in the Archean and Proterozoic plutonic units (Lucas and St-Onge 1995) and detailed studies of the metamorphic history of the Archean basement are currently under review (St-Onge and Lucas 1995; St-Onge and Ijewliw, in press; Scott and St-Onge, in press). A MSc thesis (Monday 1994) evaluated the metamorphic conditions of the Narsajuaq terrane while this present PhD thesis documents the magmatic evolution of various components of the orogen. The geochemical evolution of mafic magmas of the orogen has been presented (Dunphy et al. 1995) as well as the results of smaller-scale studies on the arc-related plutons of the Cape Smith Belt (Dunphy 1994) and on geochronological constraints of late magmatism within the orogen (Dunphy et al. in press). The geological and geochemical characteristics of the Narsajuaq terrane and its petrogenesis comprise an article in preparation (Chapter 1, Dunphy and Ludden, in preparation).

0.5 Summary of the geological setting of the Ungava Orogen

The geological setting of the Ungava Orogen (Fig. 1) has been investigated by various workers looking at different parts of the orogen, principally the Cape Smith Belt (see historical overview). The origin of the mafic volcanic and sedimentary rocks of Cape Smith Belt has been debated for many years and several different models have been presented. Early models suggested that the Cape Smith Belt was an ensialic fold belt (Dimroth et al. 1970; Baer 1977; Baragar and Scoates 1981). Another interpretation was that the belt marks the site of a major collisional geosuture (Gibb and Walcott 1971; Burke et al. 1977; Thomas and Gibb 1977). Hynes and Francis (1982), Francis et al. (1981, 1983) and Picard et al. (1990) suggested that the belt marks the site of a small short lived ocean basin. Hoffman (1985) proposed that the belt is essentially a klippe with its root zone (geosuture) located further to the north and separated from the belt by a post-thrusting antiform (the Kovik antiform, Fig.1).

The more recent work by the MRN in the western portion of the belt and the GSC in the eastern region and in the northern hinterland has demonstrated (1) the presence of continuous basement (Superior Province) exposure around the eastern end of the Cape Smith Belt (Taylor 1982; St-Onge et al. 1992); (2) the existence of two autochthonous basement half-windows coring the large-scale antiform north of the belt (Lucas and St-Onge 1991; St-Onge and Lucas 1992); (3) a ca. 2.0 Ga ophiolite (the Purtuniq ophiolite) preserved in the belt (St-Onge et al. 1988; Scott et al. 1989, 1991); (4) volcanic and volcanoclastic rocks of the Parent Group, interpreted as a subduction-related volcanic arc (Picard et al. 1990; Barrette 1994); and (5) that the hinterland of the Cape Smith Belt is characterized by a 1.9-1.8 Ga magmatic arc terrane (the Narsajuaq terrane; Dunphy and Ludden 1992; St-Onge et al. 1992).

Incorporation of all the geological data, including these more recent observations, resulted in the development of several somewhat similar geotectonic models for the origin of the orogen by a variety of authors (Picard et al. 1990; St-Onge and Lucas 1990a; Lucas et al. 1992; St-Onge et al. 1992; Dunphy et al. 1995). These models all involve continental rifting, development of an oceanic basin, subsequent subduction associated with a change in plate geometry and eventual closure of the oceanic basin.

The lithological and geochemical characteristics of the Povungnituk Group (Fig. 1) and associated Eskimo and Flaherty Formations of the Belcher Islands have been interpreted as deposits related to rifting of the Superior Province continental crust (Hynes and Francis 1982; Francis et al. 1983; Arndt et al. 1987; Picard et al. 1990; St-Onge and Lucas 1990b; Legault et al. 1994; Dunphy et al. 1995). The mafic volcanics of the Chukotat Group (and correlative volcanics from the Ottawa Islands) appear to record the transition from accumulation of continental rift-related rocks to the formation of transitional to oceanic crust (Francis and Hynes 1979; Francis et al. 1983; Arndt et al. 1987; Picard et al. 1990). Rifting appears to have been underway by 2038 Ma, the age of a gabbro sill emplaced in the lower Povungnituk Group sedimentary rocks (Machado et al. 1993), and continued through until at least 1959 Ma, the age of a rhyolite at the top of the Povungnituk Group (Parrish 1989). The Chukotat Group has been dated at 1918 Ma (Parrish 1989) which produces a gap in the geological record (between 1959 and 1918 Ma) that is not well understood.

A series of layered mafic and ultramafic rocks, massive and pillowed basalt flows, sheeted dykes and rare plagiogranite intrusions make up the Watts Group (Fig. 1) which has been interpreted as an ophiolite (St-Onge et al. 1988; Scott et al. 1991) having an age of 1998 Ma (Parrish 1989). The Watts Group structurally overlies the younger Chukotat Group due to later thrusting and deformation associated with the development of the Cape Smith thrust-fold Belt (St-Onge and Lucas 1990b). The geochemical and isotopic character of the Watts Group rocks indicate that both depleted MORB-like and enriched OIB-like magmas were involved in its petrogenesis (Hegner and Bevier 1991; Scott et al. 1991).

The rifting event that produced these various lithologies of the Ungava Orogen may have been related to hot-spot activity and the arrivial of a mantle plume which could have produced the volume of mafic magmas and their depleted and enriched geochemical signatures (Hegner and Bevier 1991; Scott et al. 1991; Dunphy et al. 1995). A change in the regional tectonics from extensional to compressional, possibly related to the assembly of the supercontinent of Laurentia (Hoffman 1988; Van Kranendonk et al. 1993), may have been responsible for the development of a subduction system which gave rise to the arc-related deposits of the Parent Group, Spartan Group and the

Narsajuaq terrane (Fig. 1; Picard et al. 1990; Lucas et al. 1992; St-Onge et al. 1992; Dunphy et al. 1995). The initiation of collisional is constrained as younger than 1918 Ma, the age of the youngest rift-related rock (Parrish 1989), and older than 1898 Ma, the age of the oldest arc-related pluton (Machado et al. 1993).

Arc-related magmatism spanned nearly 80 Ma (1898-1821 Ma; Parrish 1989; St-Onge et al. 1992; Machado et al. 1993; R. Parrish, personal communication 1994) and involved several pulses of activity (i.e., older, younger and Cape Smith suites). Construction of the magmatic arc was synchronous with the early development of the Cape Smith Thrust and Fold Belt which began ca. 1870 Ma, the age of a gabbroic sill in the Chukotat Group that contains Archean xenocrystic zircons that are interpreted as inherited from the Superior Province (St-Onge et al. 1992). Basin closure was nearly complete by 1821 Ma (the youngest truely arc-related pluton), although magmatism continued for another 70 Ma, in several distinct pulses. Large granitic sheets with isotopic compositions indicating a significant crustal signature were intruded ca. 1800 Ma (R. Parrish, personal communication 1994), possibly related to the last-stages of arc magmatism whereby mafic magmas derived from subduction processes pooled at the base of the crust and caused partial melting of that crust, which rose and intruded at higher crustal levels. Final accretion of the Narsajuaq terrane to the Superior Province margin had occurred by 1758 Ma, the age of late undeformed syenogranite pegmatite dykes which cross-cut all units of the orogen (Parrish 1989; St-Onge et al. 1992). A final stage of deformation about N-S axes gave rise to the distinct dome-and-basin interference pattern which is visible in the map traces of the faults (Fig. 1). Following this event a final small pluse of granitic magmatism occurred at 1742 Ma, related to partial melting of the underlying crust due to structural thickening during the waning stages of the orogeny (Dunphy et al. in press).

References

Arndt, N.T., Brugmann, G.E., Lehnert, K., Chauvel, C., and Chappell, B.W. 1987. Geochemistry, petrogenesis and tectonic environment of Circum-Superior Belt basalts, Canada. *In* Geochemistry and Mineralization of Proterozoic Volcanic Suites. *Edited by* T.C. Pharaoh, R.D. Beckinsale and D. Rickard. Geological Society, Special Publication **33**:133-145.

Baer, A.J. 1977. Gravity anomalies and deep structure of the Cape Smith Fold Belt, northern Ungava, Quebec: comment and reply. Geology, **5**:651-653.

Baragar, W.R.A. 1974. Volcanic studies in the Cape Smith - Wakeham Bay Belt, New Quebec. *In* Report of Activities, Part A. Geological Survey of Canada, Paper 74-1A, pp. 155-157.

Baragar, W.R.A. 1984. Pillow formation and layered flows in the Circum-Superior Belt of eastern Hudson Bay. Canadian Journal of Earth Sciences, **21**:781-792.

Baragar, W.R.A. and Scoates, R.F.J. 1981. The Circum-Superior Belt: a Proterozoic plate margin? *In* Precambrian Plate Tectonics. *Edited by* A. Kroner. Elsevier, New York, pp. 297-330.

Barnes, S.-J., and Barnes, S.J. 1990. A new interpretation of the Katiniq Nickel Deposit, Ungava, northern Quebec. Economic Geology, **85**:1269-1272.

Barnes, S.-J., and Giovenazzo, D. 1990. Platinum-group elements in the Bravo intrusion, Cape Smith Fold Belt, northern Quebec. Canadian Mineralogist, **28**:431-449.

Barrette, P.D. 1988. Géologie de la région du lac Bilson, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 87-23.

Barrette, P.D. 1989. Géologie de la région du lac Bolduc, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 88-17.

Barrette, P.D. 1990. Géologie de la région du lac Bolduc (Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, ET 89-03.

Barrette, J.P. 1991. Étude gîtologique de la partie occidentale de la fosse de l'Ungava, région des lac Bilson, Bolduc, Vigneau et de la baie Korak. Ministère de l'Energie et des Ressources du Québec, MB 91-22.

Barrette, P.D. 1994. Lithostratigraphy and map-scale structure in the western Cape Smith Belt, northern Quebec: a tentative correlation between two tectonic domains. Canadian Journal of Earth Sciences, **31**:986-994.

Beall, G.H. 1959. Cross Lake area, New Quebec. Preliminary Report 396, Preliminary Map 1267.

Beall, G.H. 1960. Laflamme Lake area, New Quebec. Preliminary Report 435, Preliminary Map 1351.

Beall, G.H., Hurley, P.M., Fairbairn, H.W., and Pinson, W.H. 1963. Comparison of K-Ar and whole rock Rb-Sr dating in New Quebec and Labrador. American Journal of Science, **261**:571-580.

Begin, N.J. 1989. P-T conditions of metamorphism inferred from the metabasites of the Cape Smith Belt, northern Quebec. Geoscience Canada, **16**:151-154.

Begin, N.T. 1992a. Contrasting mineral isograd sequences in metabasites of the Cape Smith Belt, northern Quebec, Canada: three new bathograds for mafic rocks. Journal of Metamorphic Geology, **10**:685-704.

Begin, N.T. 1992b. Textural and compositional relationships of Ca-amphiboles in metabasites of the Cape Smith Belt, northern Quebec: implications for a miscibility gap at medium pressure. Journal of Petrology, **33**:1317-1343.

Bell, R. 1885. Observations on the geology, mineralogy, zoology and botany of the Labrador Coast, Hudson Strait and Bay. Geological Survey of Canada, Report of Progress, 1882-84.

Bennett, V.C., Nutman, A.P., and McCulloch, M.T. 1993. Nd isotopic evidence for transient, highly depleted mantle reservoirs in the early history of the Earth. Earth and Planetary Sciences Letters, **119**:299-317.

Bergeron, R. 1957. Cape Smith - Wakeham Bay belt, New Quebec, Quebec. Preliminary Report 355, Preliminary Maps 1090, 1196.

Bergeron, R. 1959. Povungnituk range area, New Quebec. Preliminary Report 392, Preliminary Map 1279.

Burke, K., Dewey, J.F., and Kidd, W.S.F. 1977. World distribution of sutures - the sites of former oceans. Tectonophysics, **40**:69-99.

Campbell, I.H., and Griffiths, R.W. 1992. The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. Journal of Geology, **92**:497-523.

Card, K.D. 1990. A review of the Superior province of the Canadian Shield, a product of Archaean accretion. Precambrian Research, **48**:99-156.

DeMontigny, P.A. 1959. Upper Deception River area, New Quebec. Preliminary Report 398, Preliminary Map 1285.

Dimroth, E., Baragar, W.R.A., Bergeron, R., and Jackson, G.D. 1970. The filling of the Circum-Ungava geosyncline. *In* Symposium on basins and geosynclines of the Canadian Shield. *Edited by* A.J. Baer. Geological Survey of Canada, Paper 70-40, pp. 45-142.

Doig, R. 1983. Rb-Sr isotopic study of Archean gneisses north of the Cape Smith fold belt, Ungava, Quebec. Canadian Journal of Earth Sciences, **20**:821-829.

Doig, R. 1987. Rb-Sr geochronology and metamorphic history of Proterozoic to Early Archean rocks north of the Cape Smith Fold Belt, Quebec. Canadian Journal of Earth Sciences, **24**:813-825.

Drummond, M.S., and Defant, M.J. 1990. A model for trondhjemite-tonalite-dacite genesis and crustal growth via slab melting: Archean to modern comparisons. Journal of Geophysical Research, **95**:21503-21521.

Dunphy, J. 1994. Evolution des roches plutoniques du domaine nord de la Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, MB 94-58.

Dunphy, J.M., and Ludden, J.N. 1992. The Narsajuaq terrane, northern Quebec: a mid-crustal section of an Early Proterozoic magmatic arc. Eos Transactions (American Geophysical Union), **73**:318.

Dunphy, J.M., and Ludden, J.N. 1994. Geochemical and isotopic characteristics of an Early Proterozoic accreted magmatic arc (Narsajuaq terrane), northern Quebec, Canada. Eighth international conference on geochronologgy (ICOG-8), U.S.G.S. Circular **1107**:88.

Dunphy, J.M., and Skulski, T. 1995a. The Early Proterozoic De Pas batholith, New Quebec Orogen, Canada: an Andean arc or a continental collision massif?

Precambrian 95 conference, program with abstracts, p.245.

Dunphy, J.M., and Skulski, T. 1995b. Early Proterozoic granitic magmatism in the Ungava and New Quebec Orogens: the Narsajuaq terrane and the De Pas batholith. *In*: Eastern Canadian Shield Onshore-Offshore Transect (ECSOOT) Meeting Report (Nov. 28-29, 1994). Wardle, R.J. and Hall, J. (*compilers*), LITHOPROBE Report 45, p.37-50.

Dunphy, J.M., Ludden, J.N., and Francis, D.M. 1993. Geochemistry of magmas from the Ungava orogen, Quebec, Canada: Implications for mantle evolution proir to 2.0 Ga. International Association of Volcanology and Chemistry of the Earth's Interior (IAVCEI) General Assembly 1993, Book of Abstracts, p. 28.

Dunphy, J.M., Ludden, J.N., and Francis, D. 1995. Geochemistry of mafic magmas from the Ungava orogen, Quebec, Canada and implications for mantle reservoir compositions at 2.0 Ga. Chemical Geology, **120**:361-380.

Dunphy, J.M., Ludden, J.N., and Parrish, R.R. in press. Stitching together the Ungava orogen, northern Quebec: geochemical, isotopic and geochronological constraints of late magmatic events. Canadian Journal of Earth Sciences, v. 32.

Dunphy, J.M., Parrish, R.R., and Ludden, J.N. 1991. Stitching together the Ungava orogeny: the geochemistry and age relationships of the Lac Duquet pluton. Geological Association of Canada - Mineralogical Association of Canada, Program with Abstracts, **16**:A32.

Dunphy, J.M., Skulski, T., and Brunet, S.B. 1994. Le batholite de De Pas a la latitude 56°30' (l'orogène du Nouveau-Québec): un arc Andéen ou un massif de collision? Séminaire d'information sur la recherche géologique, Ministère de l'Energie et des Ressources du Québec, DV 94-09, p. 49.

Francis, D.M., and Hynes, A.J. 1979. Komatiite-derived tholeiites in the Proterozoic of New Quebec. Earth and Planetary Science Letters, **44**:473-481.

Francis, D.M., Ludden, J.N., and Hynes, A.J. 1983. Magma evolution in a Proterozoic rifting environment. Journal of Petrology, **24**:556-582.

Francis, D.M., Hynes, A.J., Ludden, J.N., and Bédard, J. 1981. Crystal fractionation and partial melting in the petrogenesis of a Proterozoic high-MgO volcanic suite, Ungava Quebec. Contributions to Mineralogy and Petrology, **78**:27-36.

Gaonac'h, H., Picard, C., Ludden, J.N., and Francis, D.M. 1989. Alkaline rocks from a Proterozoic volcanic island in the Cape Smith Thrust Belt, New Quebec. Geoscience Canada, **16**:137-139.

Gaonac'h, H., Ludden, J.N., Picard, C., and Francis, D.M. 1992. Highly alkaline lavas in a Proterozoic rift zone: Implications for Precambrian mantle metasomatic processes. Geology, **20**:247-250.

Gélinas, L. 1962. Watts Lake area, New Quebec. Preliminary Report 471, Preliminary Map 1414.

Gibb, R.A., and Walcott, R.I. 1971. A Precambrian suture in the Canadian Shield. Earth and Planetary Science Letters, **10**:417-422.

Giovenazzo, D. 1986. La Fosse de l'Ungava: une province métallogénique enrichie en éléments du groupe de platine; dans Exploration en Ungava: données récentes sur la géologie et la gîtologie. Ministère de l'Energie et des Ressources du Québec., DV-86-16, pp. 75-81.

Giovenazzo, D., and Lefebvre, C. 1986. Classification des indices minéralisés de la Fosse de l'Ungava; dans Exploration en Ungava; donées recentes sur la géologie et la gîtologie. Ministère de l'Energie et des Ressources du Québec, DV 86-16, pp. 73-76.

Giovenazzo, D., Picard, C., and Guha, J. 1989. Tectonic setting of Ni-Cu-PGE deposits in the central part of the Cape Smith Belt. Geoscience Canada, **16**:134-136.

Gold, D.P. 1962. Brisebois Lake area, New Quebec. Preliminary Report 470, Preliminary Map 1413.

Gunning, H.C. 1934. Sulphide deposits at Cape Smith, east coast of Hudson Bay. In Gelogical Survey of Canada, Summary Report 1933, Part D pp. 139-154.

Harvey, D. 1995. Géochime et traceur isotopique du Nd dans les intrusifs Archéens de la parite Nord-est de la Péninsule d'Ungava. M.Sc. thesis, Université de Montréal, Montreal, Québec.

Hegner, E., and Bevier, M.L. 1989. Geochemical constraints on the origin of mafic rocks from the Cape Smith Belt. Geoscience Canada, **16**:148-151.

Hegner, E., and Bevier, M.L. 1991. Nd and Pb isotopic constraints on the origin of the Purtuniq ophiolite and Early Proterozoic Cape Smith Belt, northern Quebec, Canada. Chemical Geology, **91**:357-371.

Hervet, M. 1985. Région de la vallée de Narsajuaq, Ungava. Ministère de l'Energie et des Ressources du Québec, DP 85-05.

Hoffman, P.F. 1985. Is the Cape Smith Belt (northern Quebec) a klippe? Canadian Journal of Earth Sciences, **22**:1361-1369.

Hoffman, P.F. 1988. United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annual Reviews of Earth and Planetary Science, **16**:543-603.

Hynes, A.J., and Francis, D.M. 1982. A transect of the Early Proterozoic Cape Smith foldbelt, New Quebec. Tectonophysics, **88**:23-59.

Jacobsen, S.B. 1988. Isotopic and chemical constraints on mantle-crust evolution. Geochimica et Cosmochimica Acta, **52**:1341-1350.

Kimura, G., Ludden, J.N., Desrochers, J.P., and Hori, R. 1993. A model of ocean-crust accretion for the Superior province, Canada. Lithos, **30**:337-355.

Lamothe, D., Picard, C., and Moorhead, J. 1983. Région du lac Beauparlant, Bande de Cap Smith - Maricourt. Ministère de l'Energie et des Ressources du Québec, DP 84-39.

Lamothe, D., Gagnon, R., and Clarke, T., *(Editor)*. 1986. Exploration en Ungava: données récents sur la géologie et la gîtologie. Ministère de l'Energie et des Ressources du Québec Séminaire d'information DV 86-16 pp. 81.

Legault, F., Francis, D., Hynes, A., and Budkewitsch, P. 1994. Proterozoic continental volcanism in the Belcher Islands: implications for the evolution of the Circum Ungava Fold Belt. Canadian Journal of Earth Sciences, **31**:1536-1549.

Low, A.P. 1899. Report on an exploration of part of the south shore of Hudson Strait and Ungava Bay. Geological Survey of Canada, Annual Report, vol. XI, part L.

Low, A.P. 1902. Report on an exploration of the east coast of Hudson Bay from Cape Wolstenholme to the south end of James Bay. Geological Survey of Canada, Annual Report, volume XIII, part D.

Lucas, S.B. 1989. Structural evolution of the Cape Smith Thrust Belt and the role of out-of-sequence faulting in the thickening of mountain belts. Tectonics, **8**:655-676.

Lucas, S.B. 1990. Relations between thrust belt evolution, grain-scale deformation, and metamorphic processes: Cape Smith Belt, northern Quebec, Canada. Tectonophysics, **178**:151-182.

Lucas, S.B., and Byrne, T. 1992. Footwall involvement during arc-continent collision, Ungava orogen, northern Canada. Journal of the Geological Society, **149**:237-248.

Lucas, S.B., and St-Onge, M.R. 1989a. Shear zone softening at the base of the Cape Smith Belt: implication for the rheological evolution of thrust belts. Geoscience Canada, **16**:158-163.

Lucas, S.B., and St-Onge, M.R. 1989b. Structural evolution of the Cape Smith Belt from initial thrusting to basement-involved folding. Geoscience Canada, **16**:122-126.

Lucas, S.B., and St-Onge, M.R. 1991. Evolution of Archean and Early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec. *In* Current Research, Part C. Geological Survey of Canada, Paper 91-1C, pp. 109-119.

Lucas, S.B., and St-Onge, M.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 2: Structural and metamorphic history. Canadian Journal of Earth Sciences, **29**:765-782.

Lucas, S.B., St-Onge, M.R., Parrish, R.R., and Dunphy, J.M. 1992. Long-lived continent-ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. Geology, **20**:113-116.

Lucas, S.B., and St-Onge, M.R. 1995. Syn-tectonic magmatism and the development of compositional layering, Ungava orogen (northern Quebec, Canada). Journal of Structural Geology, **17**:475-491.

Machado, N., David, J., Scott, D.J., Lamothe, D., Philippe, S., and Gariépy, C. 1993. U-Pb geochronology of the western Cape Smith Belt, Canada: New insights on the age of initial rifting and arc magmatism. Precambrian Research, **63**:211-223.

Martin, H. 1986. Effect of steeper Archean geothermal gradient on geochemistry of subduction-zone magmas. Geology, **14**:753-756.

Monday, P. 1994. Narsajuaq terrane, Ungava Orogen: Thermobarometry of granulite-facies metasediments and metaigneous rocks and tectonic implications. M.Sc. thesis, Queens University, Kingston, Ontario.

Moorhead, J. 1985. Géologie de la région du lac Chukotat. Ministère de l'Energie et des Ressources du Québec, DP 86-01.

Moorhead, J. 1987. Géologie de la région du lac Hubert, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 86-31.

Moorhead, J. 1988. Géologie de la région du lac Vigneau, Nouveau Québec. Ministère de l'Energie et des Ressources du Québec, DP 88-05.

Moorhead, J. 1989. Géologie de la région du lac Chukotat (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 87-10.

Nelson, B.K., and DePaolo, D.J. 1985. Rapid production of continental crust 1.7-1.9 b.y. ago: Nd isotopic evidence from the basement of the North American mid-continent. Geological Society of America Bulletin, **96**:746-754.

Parrish, R.R. 1989. U-Pb geochronology of the Cape Smith Belt and Sugluk block, northern Quebec. Geoscience Canada, **16**:126-130.

Patchett, J.P., and Bridgwater, D. 1984. Origin of continental crust of 1.9-1.7 Ga age defined by Nd isotopes in the Ketilidian terrain of South Greenland. Contributions to Mineralogy and Petrology, **87**:311-318.

Patchett, J.P., and Arndt, N.T. 1986. Nd Isotopes and tectonics of 1.9-1.7 Ga crustal genesis. Earth and Planetary Science Letters, **78**:329-338.

Picard, C. 1989a. Lithochimie des roches volcaniques protérozoïques de la partie occidentale de la Fosse de l'Ungava (région au sud du lac Lanyan). Ministère de l'Energie et des Ressources du Québec, ET 87-14.

Picard, C. 1989b. Pétrologie et volcanologie des roches volcaniques protérozoïques de la partie centrale de la Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, ET 87-07.

Picard, C., Lamothe, D., Piboule, M., and Oliver, R. 1990. Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt

(New-Quebec). Precambrian Research, **47**:223-249.

Roy, C. 1989. Géologie de la région du lac Bélanger, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, MB 89-13.

Schwarz, E.J., and Fujiwara, Y. 1977. Komatiitic basalts from the Proterozoic Cape Smith Range in northern Quebec, Canada. *In* Volcanic regimes in Canada. *Edited* by W.R.A. Baragar, L.C. Coleman and J.M. Hall. Geological Association of Canada, Special Paper, **16**:193-201.

Scott, D.J., and Bickle, M.J. 1991. Field relationships in the Early Proterozoic Purtuniq ophiolite, Lac Watts and Purtuniq map areas, Quebec. In Current Research, Part C. Geological Survey of Canada, Paper 91-1C, pp. 179-188.

Scott, D.J., and St-Onge, M.R. in press. Constraints on Pb closure temperature in titanite based on rocks from the Ungava Orogen, Canada: implications for the U-Pb geochronology and P-T-t path determinations. Geology.

Scott, D.J., Helmstaedt, H., and Bickle, M.J. 1992. Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. Geology, **20**:173-176.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1989. The 1998 Ma Purtuniq ophiolite: imbricated and metamorphosed oceanic crust in the Cape Smith Thrust Belt, northern Quebec. Geoscience Canada, **16**:144-147.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1991. Geology and chemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. *In* Ophiolite Genesis and Evolution of the Oceanic Lithosphere. *Edited by* T. Peters, A. Nicolas and R.G. Coleman. Kluwer Academic Publishers, Amsterdam, pp. 817-849.

Sleep, N.H., and Windley, B.F. 1982. Archean plate tectonics: constraints and inferences. Journal of Geology, **90**:363-379.

St-Onge, M.R., and Lucas, S.B. 1986. Structural and metamorphic evolution of an Early Proterozoic thrust-fold belt, eastern Cape Smith belt (Ungava Trough), Quebec. *In*: Exploration en Ungava: donnees recentes sur la geologie et la gitologie.

Ministère de l'Energie et des Ressources du Québec, DV 86-16, pp. 31-39.

St-Onge, M.R., and Lucas, S.B. 1989-1990. Geology, eastern portion of the Cape Smith Thrust-Fold Belt, parts of the Wakeham Bay, Cratère du Nouveau-Québec and Nuvilik Lakes map areas, northern Quebec. Geological Survey of Canada, Map Maps 1721A-1735A, scale 1:50,000.

St-Onge, M.R., and Lucas, S.B. 1990a. Evolution of the Cape Smith Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. *In* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 313-351.

St-Onge, M.R., and Lucas, S.B. 1990b. Early Proterozoic collisional tectonics in the internal zone of the Ungava (Trans-Hudson) orogen, Lacs Nuvilik and Sugluk map areas, Quebec. *In* Current Research, Part C. Geological Survey of Canada, Paper 90-1C, pp. 119-132.

St-Onge, M.R., and Lucas, S.B. 1991. Evolution of regional metamorphism in the Cape Smith Thrust Belt (northern Quebec, Canada): interaction of tectonic and thermal processes. Journal of Metamorphic Geology, **9**:515-534.

St-Onge, M.R., and Lucas, S.B. 1992. New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. *In* Current Research, Part C. Geological Survey of Canada, Paper 92-1C, pp. 31-41.

St-Onge, M.R., and Lucas, S.B. 1993. Geology of the eastern Cape Smith Belt: Parts of the Kangiqssujuaq, Cratère du Nouveau-Québec, and Lacs Nuvilik map areas, Quebec. *In* Geological Survey of Canada, Memoir 438, pp. 110.

St-Onge, M.R., and Lucas, S.B. 1995. Large-scale fluid infiltration, metasomatism and re-equilibration of Archean basement granulites during Paleoproterozoic thrust belt construction, Ungava orogen, Canada. Journal of Metamorphic Geology, **13**:509-535.

St-Onge, M.R., and Ijewliw, O.J. in press. Mineral corona formation during high-P retrogression of granulitic rocks, Ungava Orogen, Canada. Journal of Petrology.

St-Onge, M.R., Lucas, S.B., and Parrish, R.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Canadian Journal of Earth Science, **29**:746-764.

St-Onge, M.R., Lucas, S.B., Scott, D.J., and Bégin, N.J. 1986. Eastern Cape Smith Belt: an Early Proterozoic thrust-fold belt and basal shear zone exposed in oblique section, Wakeham Bay and Cratère du Nouveau-Québec map areas, northern Quebec. *In* Current Research, Part A. Geological Survey of Canada, Paper 86-1A, pp. 1-14.

St-Onge, M.R., Lucas, S.B., Scott, D.J., and Bégin, N.J. 1987. Tectonostratigraphy and structure of the Lac Watts - Lac Cross - Rivière Déception area, central Cape Smith Belt, northern Quebec. *In* Current Research, Part A. Geological Survey of Canada, Paper 87-1A, pp. 619-632.

St-Onge, M.R., Lucas, S.B., Scott, D.J., and Bégin, N.J. 1989. Evidence for the development of oceanic crust and for continental rifting in the tectonostratigraphy of the Early Proterozoic Cape Smith Belt. Geoscience Canada, **16**:119-122.

St-Onge, M.R., Lucas, S.B., Scott, D.J., Bégin, N.J., Helmstaedt, H., and Carmicheal, D.M. 1988. Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec. *In* Current Research, Part C. Geological Survey of Canada, Paper 88-1C, pp. 1-18.

Stam, J.C. 1961. On the geology and petrology of the Cape Smith - Wakeham Bay Belt, Ungava, New Quebec. Geologie en Mijnbouw, **40**:412-421.

Taylor, F.C. 1974. Reconnaissance geology of a part of the Precambrian Shield, northern Quebec and Northwest Territories. *In* Geological Survey of Canada, Paper 74-21, pp. 10.

Taylor, F.C. 1982. Reconnaissance geology of a part of the Canadian Shield, northern Quebec and Northwest Territories. *In* Geological Survey of Canada, Memoir 399, pp. 32.

Taylor, S.R., and McLennan, S.M. 1985. The continental crust: its composition and evolution. Blackwell Scientific Publications, Boston, pp. 312.

Thibert, F., Picard, C., and Trzcienski, W. 1989. Pétrologie des filons-couches différenciés Roméo 1 et 2 dans la partie centrale de la bande du Cap Smith. Geoscience Canada, **16**:140-144.

Thomas, M.D., and Gibb, R.A. 1977. Gravity anomalies and deep structure of the Cape Smith Fold Belt, northern Ungava, Quebec. Geology, **5**:169-172.

Tremblay, G. 1987. Géologie de la région du lac Lessard, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 86-30.

Tremblay, G. 1989. Géologie de la région du lac Vanasse (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 87-08.

Tremblay, G. 1991. Géologie de la région du lac Lessard (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 88-09.

Van Kranendonk, M.J., St-Onge, M.R., and Henderson, J.R. 1993. Paleoproterozoic tectonic assembly of Northeast Laurentia through multiple indentations. Precambrian Research, **63**:325-347.

CHAPTER 1

Geochemical and isotopic characteristics of an Early Proterozoic accreted magmatic arc (Narsajuaq terrane), northern Quebec, Canada

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ABSTRACT

The Narsajuaq terrane is a major component of the Early Proterozoic Ungava (Trans-Hudson) Orogen of northern Quebec and is a well exposed mid-crustal block interpreted to be the plutonic roots of a magmatic arc preserved in oblique cross-section within the Ungava Orogen. The Ungava Orogen represents an arc - collision belt recording more than 200 Ma of divergent and convergent tectonic activity.

The Narsajuaq terrane is comprised of a series of metamorphosed (granulite- to amphibolite-grade) plutons and supracrustal rocks which were accreted to the Archean Superior Province basement during the Early Proterozoic. On the basis of geological field relationships and geochronological data the plutonic rocks of the Narsajuaq terrane are divided into 4 suites: (1) the *Older suite* ranges in age from 1863-1844 Ma and consists of a well layered sequence of diorite-tonalite, intruded and interlayered with variably deformed and metamorphosed granitic veins; (2) the *Younger suite*, consisting of discrete, km-size bodies of metamorphosed plutons (diorite - monzodiorite - tonalite - granite) intrusive into the Older suite, ranging in age from 1836-1800 Ma; (3) the *Late suite* of undeformed and unmetamorphosed granitic plutons and syenogranitic pegmatites that cross-cut all elements of the Ungava Orogen, dated at 1758-1742 Ma; and (4) the *Cape Smith suite*, consisting of various deformed and undeformed plutons that are found intrusive into the Cape Smith Belt lithologies and range in age from 1898-1839 Ma. A series of metamorphosed sedimentary rocks (predominantly semi-pelites) of the Sugluk Group are also found in the Narsajuaq terrane, and contain zircons of both Archean and Early Proterozoic age.

The various plutonic suites of the Narsajuaq terrane are predominately calc-alkaline and metaluminous to mildly peraluminous and contain variable but generally moderate to high concentrations of the LILE's and light REE's, pronounced negative Nb and Ti anomalies, and fractionated REE's. A range in isotopic compositions is noted for the suites, with the Older, Younger and Late suites having initial ε_{Nd} and $^{87}Sr/^{86}Sr_{(i)}$

values ranging from juvenile (+4.0 and 0.7020, respectively) to very enriched (-18.5 and 0.7064). The isotopic signature of the plutons indicate that at least three sources were involved in their petrogenesis: depleted mantle, average 2.7 Ga Superior Province crust and old (3.1 Ga) Superior Province crust. The Cape Smith suite has a much more restricted isotopic composition which is predominantly juvenile, and suggests a minor role for crustal contamination, primarily via sediment subduction. The geochemical and isotopic signature of most of the Narsajuaq terrane plutons can be derived by fractionation and variable amounts of contamination of a slightly LILE- and light REE-enriched mantle source, although a distinct series of monzodiorite plutons were derived by partial melting of a separate and different highly enriched mantle source.

Northward-directed subduction of a previously developed oceanic basin under a crustal fragment which marked the transition across the rifted Superior Province crust from ≈ 1.90-1.82 Ga generated the Older, Younger and Cape Smith suites of the Narsajuaq terrane. Accretion of the arc complex to the southern Superior Province occurred after 1.82 Ga, the age of the youngest arc-related pluton. Large sheets of granite associated with the Younger suite were intruded at ca. 1.80 Ga, near the end of arc accretion, which were derived by partial melting of the surrounding crust during tectonic thickening and metamorphic hydration of the crust. Continuing deformation and crustal thickening resulted in crustal anatexis and the generation of the Late suite granitic magmas at ≈ 1.76-1.74 Ga.

The Narsajuaq terrane represents a superb crustal section (15-25 km thick) through an Early Proterozoic magmatic arc which is presently exposed as a consequence of its accretion to the Superior Province margin and its subsequent deformation and exhumation along the northern flank of a regional-scale antiform in the Ungava Orogen. The uppermost part of the crustal section is missing although the volcanic rocks of the Parent Group may represent such crust. Metasediments of the Sugluk Group are more prevalent near the top of the crustal section, as are large granitic sheets, where as mafic (diorite) plutons are more common deeper in the section. Changes in the style of melting (related to a change in the geothermal regime of the Earth over time) and in mantle source compositions may account for the different geochemical signature of the Proterozoic Narsajuaq terrane crust and crustal compositions of Archean and Phanerozoic age.

RÉSUMÉ

Le terrane de Narsajuaq constitue un composant majeur de l'Orogène de l'Ungava (Trans-Hudsonien) du nord du Québec au Protérozoïque inférieur. Ce terrane représente une section crustale, des profondeurs intermédiaires, interprétée comme étant la racine plutonique d'arc magmatique préservée en section oblique à l'intérieur de l'Orogène de l'Ungava. Cet orogène représente une ceinture de collision arc magmatique-continent qui documente plus de 200 Ma d'activité tectonique divergente et convergente.

Le terrane de Narsajuaq se compose d'une série de plutons et de roches supracrustales métamorphisées (granulite à amphibolite) qui ont été soudé au socle Archéen de la Province du Lac Supérieur durant le Protérozoïque inférieur. Basé sur des relations géologiques de terrain et des données géochronologiques, les roches plutoniques du terrane de Narsajuaq se divisent en 4 suites: (1) la suite dominante ("Older suite"), comprenant des âges de 1863 à 1844 Ma, consiste en une séquence rubanée de diorite-tonalite, recoupée par des veines granitiques ayant subie divers degrés de déformation et métamorphisme; (2) une suite plus jeune ("Younger suite"), compose de plutons métamorphisés (diorite - monzodiorite - tonalite - granite) bien définis, de dimension kilométrique, recoupant la "Older suite". Les plutons de la "Younger suite" sont mise en place de 1836 à 1800 Ma; (3) la suite tardive ("Late suite") se compose de plutons granitiques, non-déformés et non-métamorphisés, et de pegmatites syénogranitiques, recoupant tout les éléments de l'orogène de l'Ungava. La "Late suite" est datée entre 1758 et 1742 Ma; et (4) la suite de Cap Smith ("Cape Smith suite"), consiste en une variété de plutons déformés et non-déformés, recoupant les lithologies de la ceinture de Cap Smith, et variant en âge de 1898 à 1839 Ma. Une série de roches sédimentaires

métamorphisées (surtout des méta-pélites) du Groupe de Sugluk se trouve aussi dans le terrane de Narsajuaq, et contient des zircons d'âge Archéen et Protérozoïque.

Les différentes suites plutoniques du terrane de Narsajuaq sont à prédominance calco-alcalines et métalumineuses, contiennent des concentrations modérées à fortes de "LILE" (les éléments hygromagmatophile) et terres rares légères, des patrons de terres rares fractionnés et des anomalies négatives prononcées de Nb et Ti. Une large gamme de compositions isotopiques caractérise les différentes suites, avec "Older", "Younger" et "Late" ayant des valeurs ε_{Nd} initial et $^{87}Sr/^{87}Sr_i$ de signature juvénile (+4.0 et 0.7020) à très enrichie (-18.5 et 0.7064). La signature isotopique des plutons indique qu'au moins trois sources ayant impliquées dans leur pétrogénèse: manteau appauvri, la croûte de la Province du Lac Supérieur d'âge 2.7 Ga et la croûte de la Province du Lac Supérieur ancienne (3.1 Ga). La suite "Cape Smith" a une composition isotopique à prédominance juvénile qui suggère un rôle mineur de la contamination crustale, essentiellement par subduction de sédiments. La signature géochimique et isotopique de la plupart de plutons du terrane de Narsajuaq peut être dérivée par fractionnation et contamination d'une source mantellique légèrement enrichie en "LILE" et terres rares. Cependant, une série distincte de plutons de monzodiorite a été dérivée par fusion partielle d'une source mantellique hautement enrichie.

La subduction dirigée vers le nord d'un bassin océanique sous un fragment de la croûte continentale de la Province du Lac Supérieur de 1.90 à 1.82 Ga a générée les suites "Older", "Younger" et "Cape Smith". L'accrétion du complexe en arc sur la Province du Lac Supérieur est survenue après 1.82 Ga, tel qu'indiqué par l'âge du pluton relié au magmatisme d'arc le plus jeune. De grandes intrusions de granites associées à la suite "Younger" ont été mises en place à 1.80 Ga, vers la fin de l'accrétion de l'arc. Ces intrusions sont dérivées par fusion partielle de la croûte environnante durant l'épaississement tectonique et l'hydratation métamorphique de la croûte. La déformation progressive, ainsi que l'épaississement crustal, ont causées l'anatexie crustale et la génération des suites granitiques "Late" vers 1.76 - 1.74 Ga.

Le terrane de Narsajuaq représente une superbe section crustale (15 - 25 km d'épaisseur) à travers d'arc magmatique du Protérozoïque inférieur. Cette section est présentement exposée conséquemment à son accrétion à la marge de la Province du Lac Supérieur et sa subséquente déformation et exhumation le long du flanc nord d'un antiforme d'échelle régionale dans l'Orogène de l'Ungava. La partie supérieure de la section crustale est manquante. Cependant, les roches volcaniques du Groupe de Parent, pourraient représentées la partie supérieure de la section. Les métasédiments du Groupe de Sugluk, ainsi que les grandes intrusions granitiques, sont plus abondant vers le sommet de la section crustale, alors que les plutons mafiques sont plus commun vers la base de la section. Des variations du mode de fusion (relié au changement de régime géothermale de la terre au cours du temps) et de compositions de sources mantelliques peuvent expliquer les différences entre les signatures géochimiques de la croûte du terrane de Narsajuaq du Protérozoïque et les compositions crustales Archéennes et Phanérozoïques.

1.1. INTRODUCTION

The interval 1.8-2.0 Ga represented an important period of global continental growth and accretion, resulting in the development of large continental blocks or supercontinents often consisting of many Archean cratons welded together by younger Proterozoic orogenic belts (e.g., Laurentia: Hoffman 1988; Van Kranendonk et al. 1993; Baltic Shield: Park 1991; Western Australia: Myers 1990). Detailed study of the components of these amalgamated terranes (including extensive geochemical, geochronological, structural and metamorphic analyses of the preserved rocks), is critical for the understanding of the magmatic and tectonic processes active at the time of their formation. For example, considerable study of the Cape Smith Thrust-Fold Belt in the Early Proterozoic Ungava Orogen has documented the existence of one of the oldest sections of oceanic crust (the Purtiniq ophiolite, St-Onge et al. 1988; Scott et al. 1989, 1991; St-Onge and Lucas 1990a) and an alkaline ocean island suite (Povungnituk Group alkaline lavas, Francis et al. 1983; Gaonac'h et al. 1989, 1992; Picard et al. 1990), which have been interpreted in a plate tectonic framework involving continental rifting, oceanic basin formation and eventual ophiolite obduction.

The Ungava Orogen (St-Onge and Lucas 1990b) of northern Quebec is located within the much larger orogenic belt of the Trans-Hudson Orogen (Fig. 1), which is one of the largest and best exposed Early Proterozoic orogenic belts in the world (Lewry and Collerson 1990; Hoffman 1990). The Trans-Hudson Orogen welds together the Archean Superior Province craton and cratonic blocks such as the Hearne and Rae found to the north and east thereby forming part of the Laurentian Shield (Hoffman 1988). The Narsajuaq terrane (St-Onge and Lucas 1990b) is one component of the Ungava Orogen (Fig. 2), consisting primarily of metaplutonic rocks, which is well exposed in oblique cross-section within the orogen, thereby providing an excellent opportunity to study magmatic processes and magmatic products during the Early Proterozoic. As no previous geochemical or isotopic data existed for this region prior to our study, the samples were chosen with the goal of characterizing the composition and evolution of the rocks and assessing any regional-scale variations over the entire terrane. Resolving the origin and petrogenesis of the rocks of the Narsajuaq terrane will place constraints on the tectonic models presented for the region.

The excellent exposure of the Narsajuaq terrane has provided us with a superb Early Proterozoic mid-crustal section. Characterization of the composition and evolution of these rocks will also lead to us a better understanding of the composition of the bulk continental crust during this period in Earth's evolution. Although a few key crustal sections have been documented at several locations on the Earth's surface (e.g., the Archean Kapuskasing Zone (Superior Province, Canada; Percival and Card 1983); the Paleozoic Ivrea Zone (Italy; Berckhemer 1969; Pin 1990); the Mesozoic Kohistan complex (Pakistan; Bard 1983)), data from Early Proterozoic sections has been lacking until now. The bulk composition of the Narsajuaq terrane crust is compared to that of the other sections in order to assess changes in crustal composition through time. Preliminary study of the Narsajuaq terrane rocks have shown that they are part of a calc-alkaline suite similar in composition to many volcanic arc plutons, and they have been

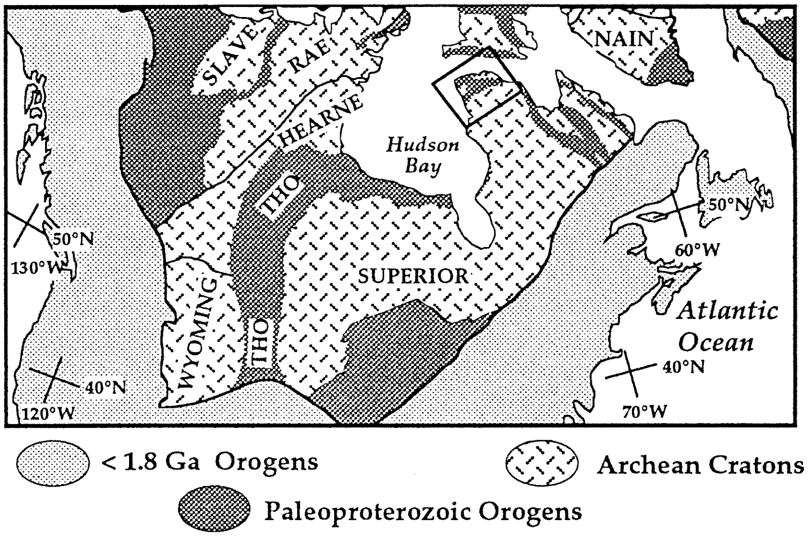


Fig. 1.1. Regional-scale map showing location of the Trans-Hudson Orogen (THO) and Archean cratons (from Lucas et al. 1994). The Ungava Orogen segment is outlined by the box.

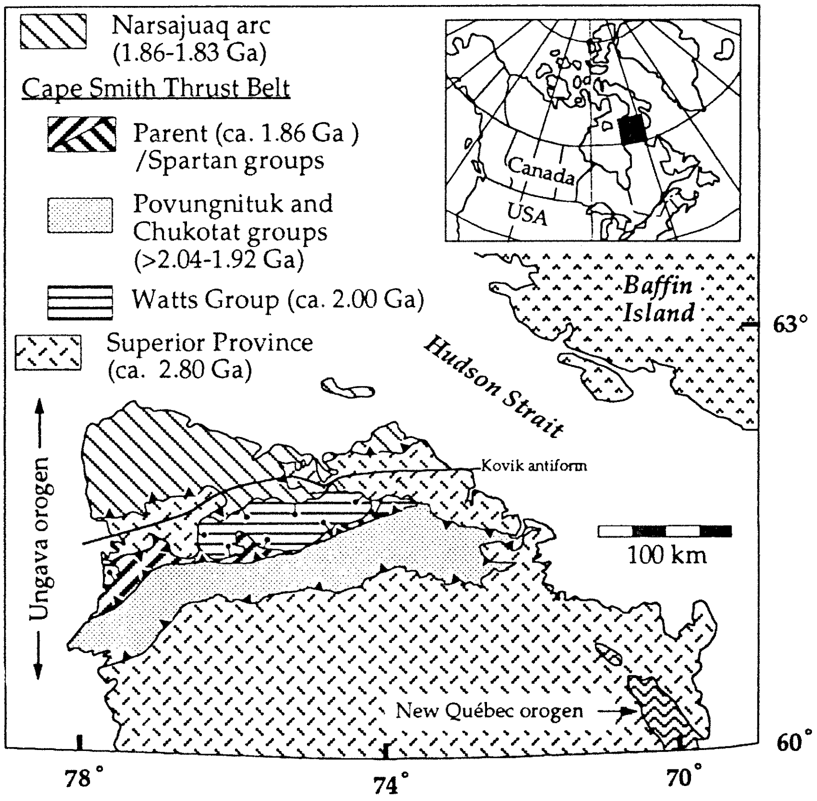


Fig. 1.2. Map illustrating the tectonic elements of the Ungava Orogen (St-Onge and Lucas 1992).

interpreted as representing the plutonic root of a ca. 1.90-1.83 Ga magmatic arc (Lucas et al., 1992; St-Onge et al., 1992). Lithologically the terrane consists of an older magmatic suite of layered diorite, tonalite and monzogranite having a gneissic appearance, which has been intruded by a younger suite of discrete plutons with a similar range in compositions. This paper presents the first complete set of geochemical and isotopic data for the Narsajuaq terrane, details the evidence for the magmatic arc origin for the plutons of the terrane, and integrates the results into a coherent geotectonic model.

Another important aspect of crustal studies focuses on the mechanism of growth of the continental crust. Although the actual concept of crustal *growth* is a hotly debated subject (e.g., Armstrong 1968; Veizer 1983; Armstrong vs. DePaolo 1981; Condie 1986; also see reviews by McLennan 1988 and Armstrong 1991), it has been acknowledged that limited crustal growth has occurred at particular places on the Earth during certain times in its history, although it may have been compensated by crustal recycling at other sites. Such growth occurs primarily through accretion of volcano-plutonic terranes, as has been documented within the Early Proterozoic supercontinent of Laurentia (Condie 1990), in the North American Cordillera (Coney et al. 1980), and can be predicted to occur in the western Pacific, in areas such as Japan and Indonesia, which will eventually collide with the Asian continent. It is not the purpose of this paper to jump into this debate, but to merely point out that continental growth did occur along the northern margin of the Superior Province when the Narsajuaq terrane (as well the other associated components in the Ungava Orogen) was generated and subsequently accreted in the Early Proterozoic, representing a significant contribution to the volume of continental crust at that time.

1.2. REGIONAL GEOLOGIC SETTING

Extensive work in the Ungava Orogen particularly during the last decade has provided a superb geological framework for the region (e.g., Francis et al. 1983; Picard et al. 1990; St-Onge and Lucas 1990a; Scott et al. 1991; St-Onge et al. 1992) and has led to a better understanding of the regional geology of the area, furnishing an excellent context for subsequent studies such as these presented here. The Ungava Orogen (Fig. 2) consists of three main tectonostratigraphic elements that were accumulated on or accreted to the northern margin of the Superior province during the period 1.8-2.0 Ga (St-Onge et al. 1992): (1) (par)-autochthonous plutonic and supracrustal rocks of the Archean Superior Province basement; (2) autochthonous and allochthonous sedimentary and volcanic units associated with a ca. 2.04-1.92 Ga rift margin and the adjacent oceanic basin (Povungnituk and Chukotat Groups); and (3) allochthonous crustal components of a ca. 2.00 Ga ophiolite (Watts Group) and a ca. 1.90-1.82 Ga magmatic arc and associated units (Narsajuaq terrane, Parent and Spartan Groups). More than 200 Ma of plate divergence and convergence are recorded in these geological units (St-Onge et al. 1992).

The Superior Province basement in the Ungava Orogen is exposed south of the Cape Smith Belt and outcrops continuously around the eastern margin of the belt northward to Foul Bay (Fig. 2; Taylor 1982; St-Onge et al. 1992). In the western part of the orogen (Fig. 2) Archean rocks (R. Parrish, personal communication 1994) are exposed in a tectonic window along the Kovik antiform (Hoffman 1985) and are interpreted as belonging to the Superior Province (St-Onge and Lucas 1992). The Superior Province basement consists primarily of metamorphosed felsic plutonic units (tonalite and granite) with lesser mafic-ultramafic units and metasedimentary rocks (St-Onge et al. 1992). U-Pb geochronology on the Superior Province plutons has produced a wide range in ages from approximately 2.8 Ga to 3.6 Ga (Parrish 1989; R. Parrish, personal communication 1994).

Detailed descriptions of the stratigraphy of the Cape Smith Belt (Fig. 2; Povungnituk, Chukotat, Spartan, Parent and Watts Groups) have been presented elsewhere (e.g., Hynes and Francis 1982; St-Onge and Lucas 1990a, 1993; Picard et al. 1990) and are only summarized here. The Povungnituk Group (Fig. 2) is the lowermost unit of the Cape Smith Belt and consists of autochthonous and allochthonous sequences of clastic metasedimentary units and overlying mafic pillowed and massive flows with minor interfingered clastic sediment. A series of alkaline lavas (basanites, nephelintes, phonolites) and pyroclastic rocks are also found within the Povungnituk Group (Gaonac'h et al. 1992). Structurally overlying and in thrust contact with the Povungnituk Group is the predominantly igneous Chukotat Group (Fig. 2) consisting of >5 km of submarine pillowed and massive basalt flows and mafic/ultramafic sills with minor associated sediments. Tectonically juxtaposed to the Chukotat Group is the Watts Group (Fig. 2) which comprises a distinct assemblage of layered mafic and ultramafic rocks, massive and pillowed basalt flows, mafic sills and sheeted dykes and rare plagiogranite intrusions that have been interpreted as an ophiolite with a preserved thickness of 7.5-8 km (St-Onge et al. 1988; Scott et al. 1991, 1992). Sandwiched between the Watts Group and the Chukotat Group in the western portion of the Cape Smith Belt is the Parent Group (Fig. 2) consisting of shales, siltstones and greywackes intercalated with basaltic to rhyolitic lavas (andesitic basalt being the most common) and volcaniclastic tuffs (Lamothe et al. 1983; Barrette 1994). Further east, at a similar stratigraphic position as the Parent Group, is the Spartan Group (Fig. 2) which is a clastic sedimentary sequence consisting primarily of pelites interbedded with semipelite and minor sandstone (St-Onge et al. 1988).

The deformational history of the Ungava Orogen is complex, as each of the three tectonostratigraphic domains underwent different episodes of deformation at different times. However, in general four deformation events can be documented (Lucas and St-Onge 1992): (1) a pre-accretion deformation (D_1) which is of Archean age for the Superior Province basement and Proterozoic age for the other two domains; (2) a Proterozoic accretion-related deformation (D_2) which is recorded in all elements of the orogen; (3) regional scale folding (D_3) of all tectonic elements about east-west axes; and (4) subsequent cross-folding of all units about north-south axes (D_4). More detailed description of the deformational structures in the Narsajuaq terrane are presented in a later section.

1.3. THE NARSAJUAQ TERRANE

1.3.1 Geology and geochronology of the Narsajuaq terrane

The Narsajuaq terrane is comprised of a complex series of high grade (amphibolite- to granulite-facies) plutons and minor supracrustal rocks which have been accreted to the Superior Province basement (Fig. 2; St-Onge and Lucas 1990b, 1992; Lucas and St-Onge 1991; St-Onge et al. 1992). These authors have described the general geology and preliminary geochronology of the various lithologies of the Narsajuaq terrane. Here we present a summary of their findings as well as new information as determined from subsequent petrological studies. Although the lithologies of the Narsajuaq terrane are metamorphosed, the prefix "meta" applies but is dropped for simplicity. Additionally, the term terrane is used here to denote a fault-bounded geological package with stratigraphic, structural and geochronological characteristics distinct from its neighbours, as defined by Coney et al. (1980).

Geological field relationships (including overall strain state, lithological association and the nature of intrusive contacts) and geochronological data have been used to divide the plutonic rocks of the Narsajuaq terrane into three main groups: (1) Older suite; (2) Younger suite; and (3) Late suite. In addition, a series of plutons intrusive into the lithologies of the Cape Smith Belt are included as a fourth group (the Cape Smith suite) as they have similar characteristics to the main Narsajuaq terrane plutons and may be related to them.

1.3.1.1 *Older suite*

The primary lithology (Fig. 3) of the Narsajuaq terrane is a well-layered (cm- to m-scale) sequence of diorite and tonalite (Fig. 4a), intruded by and interlayered with variably deformed and metamorphosed granitic (*senso lato*) veins (St-Onge et al. 1992). Although the actual composition of the diorites range from true diorite (quartz-absent) to quartz-rich diorite (IUGS classification scheme), the term diorite is used for simplicity. Similarly, the tonalites also include granodioritic compositions and granite includes monzogranite and syenogranite. Monzodiorite compositions are also noted

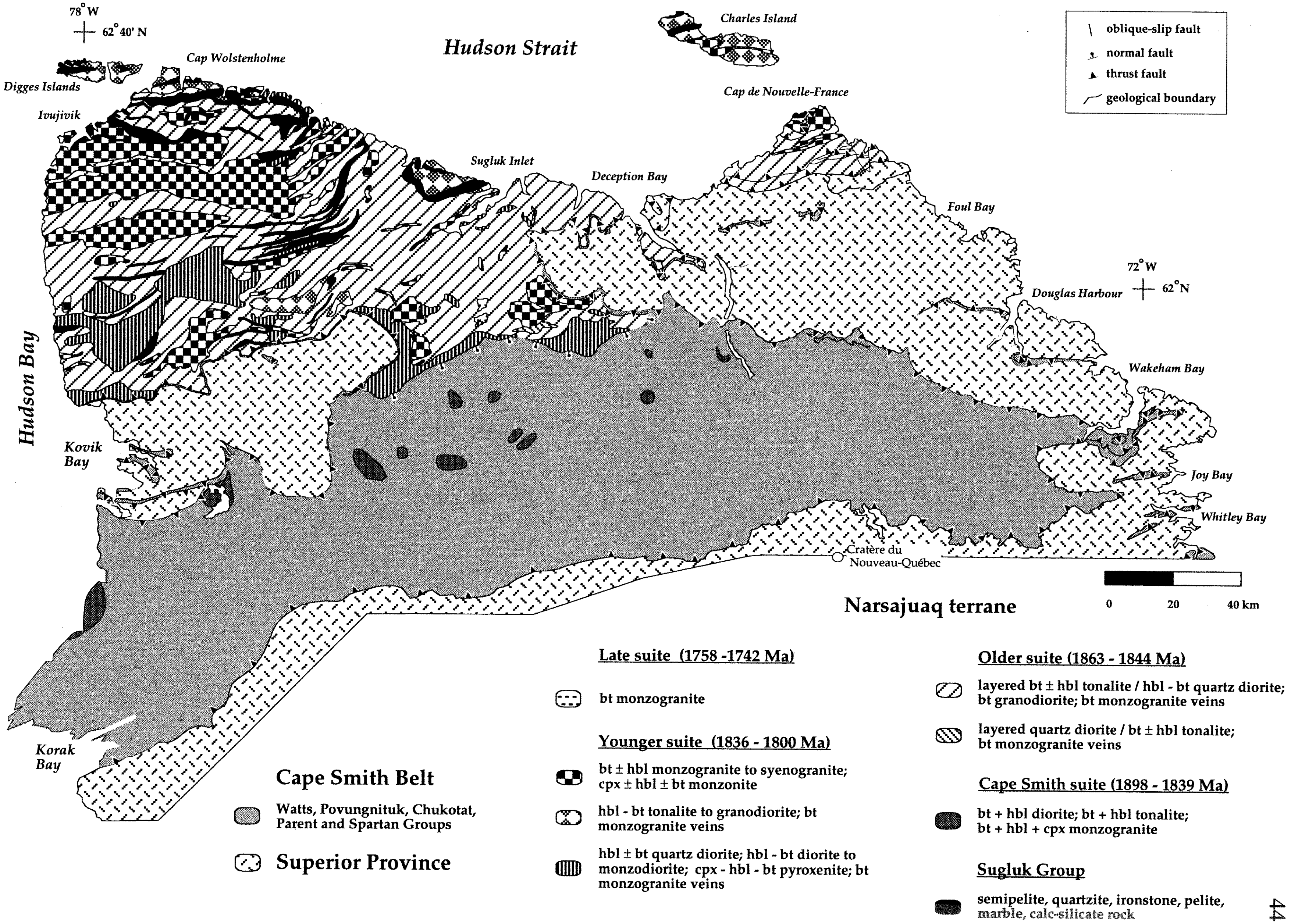


Fig. 1.3. Geological compilation map for the Narsajuaq terrane and other associated elements (Superior Province basement, Cape Smith Fold Belt) of the Ungava Orogen (modified from St-Onge and Lucas 1992). Geochronological age data from Parrish (1989), St-Onge et al. (1992), Machado et al. (1993), R. Parrish (personal communication, 1994) and Dunphy et al. (in press).

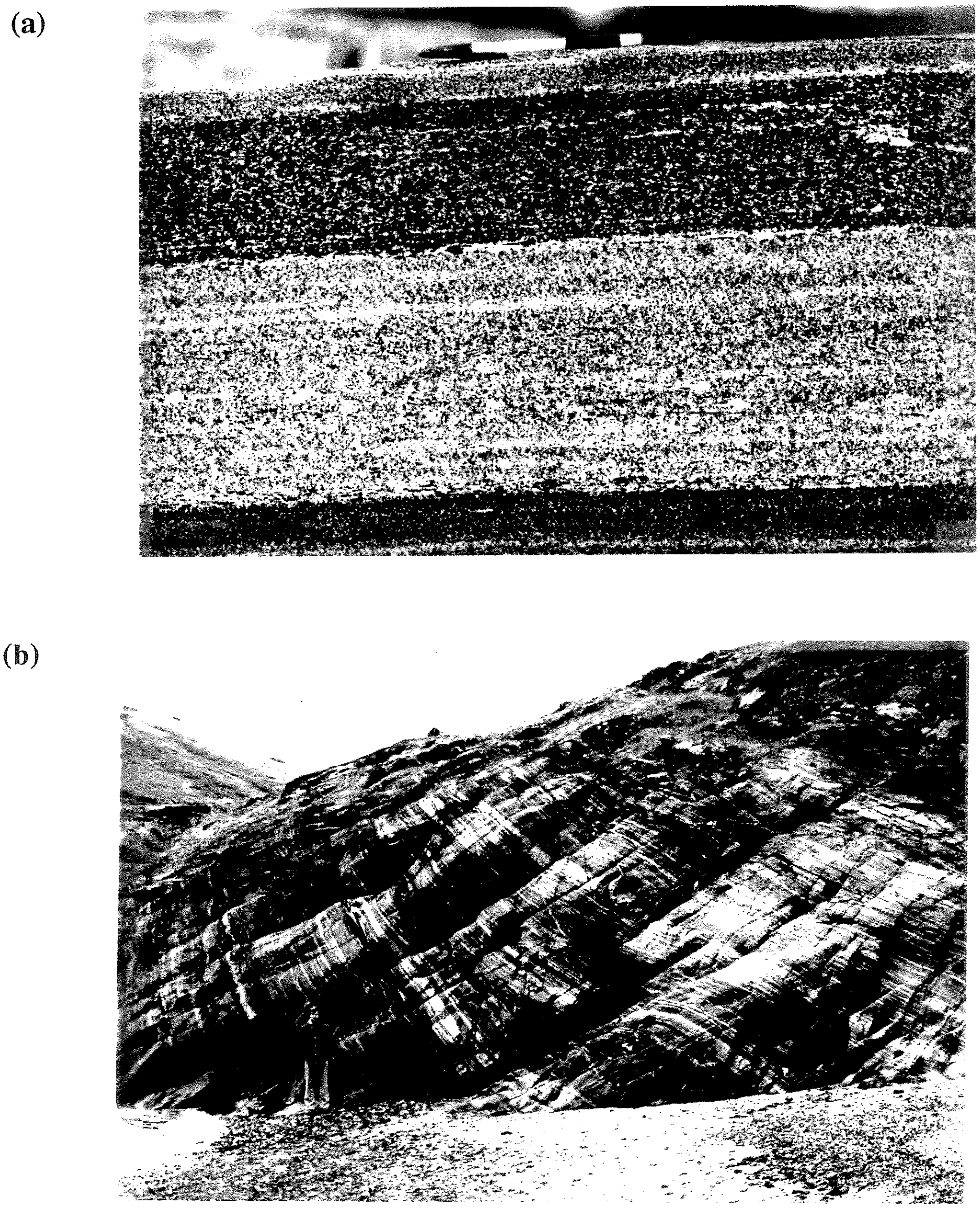


Fig. 1.4. Layered diorite-tonalite unit from the Older suite of the Narsajuaq terrane: (a) close up, with pen for scale; (b) outcrop-scale photo, with person for scale.

within the Older suite, although they are rare and do not make up a significant component of the suite. It is the compositional layering (diorite-tonalite-granite) which defines the principal planar fabric of the Narsajuaq terrane (Fig. 4b; Lucas and St-Onge 1992). Tonalites and diorites are generally biotite ± hornblende bearing with relict orthopyroxene, clinopyroxene and occasionally garnet heterogeneously distributed throughout the units. The predominant ENE-WSW foliation of these rocks is defined by the ferromagnesian minerals along with aggregates of quartz and feldspars. Volumetrically the tonalite comprises 70-80% of the Older suite although the proportions vary locally. Rarely preserved cross-cutting relationships indicate that the tonalite commonly intrudes the diorite, although diorite has also been observed cross-cutting the tonalite (St-Onge et al. 1992). A sample of the tonalite has been dated (all dates, unless otherwise indicated are U-Pb zircon ages) at 1863 Ma whereas two quartz diorite samples give ages of 1845 and 1844 Ma (Table 1; St-Onge et al. 1992; R. Parrish, personal communication 1994). A notable feature of this unit is its compositional and textural consistency across the entire orogen (>250 km strike length) and the ubiquitous occurrence of granitic (senso lato) veins.

The monzogranite to syenogranite veins of the Older suite are biotite ± hornblende bearing and range in thickness from several mm to greater than 10 m. Several generations of veins are present, with the oldest and most deformed veins concordant to the layering of the host tonalite-quartz diorite unit (Fig. 5; St-Onge and Lucas 1992). Younger, variably deformed veins also cross-cut the tonalite-quartz diorite unit (Fig. 5). Some of these veins may be related to younger granitic plutons (see below) while others have no clear spatial or temporal relationship to larger granite bodies. It is postulated that the parental bodies are not exposed at the present erosional surface. Syntectonic intrusion of granitic magmas sub-parallel to the existing compositional layering has been proposed as the mechanism of emplacement for the Older suite veins (Lucas and St-Onge 1995). This primary magmatic feature has been accentuated by later high-T deformation, enhancing the overall compositional layering observed in the Narsajuaq terrane rocks. A foliated granodiorite vein has been dated at

Table 1. Summary of U-Pb ages for the Narsajuaq terrane suites.

Suite	Sample number	Lithology	Age (± error) (Ma)	Reference *
Older suite	L-90a-89	diorite	1863 ± 2	2
	D-41b-89	granodiorite (vein)	1861 ± 2	2
	D-41a-89	granite (vein)	1848 ± 5	4
	L-90b-89	diorite	1845 ± 2	4
	D-41c-89	diorite	1844 +13/-10	2
Younger suite	D-20-89	monzodiorite	1836 ± 0.5	2
	PCA-89-56	monzogranite	1835 ± 1	2
	Sugluk-2-87	monzoganite	1835 ± 1	1
	D-43-89	monzodiorite	1834 ± 0.6	2
	Sugluk-5-87	tonalite	1830 ± 2	1
	PCA-89-32	tonalite	1826 ± 1	2
	Sugluk-3-87	tonalite	1825 ± 3	1
	S-215-91	diorite	1821 ± 1	4
	S-154a-90	granite (CNF)	1803 ± 3	4
	S-217-91	granite (IVUJ)	1800 ± 2	4
Late suite	Sugluk-1-87	syenogranite	1758 ± 1	1
	L-68-89	granite (Lac Duquet)	1742 ± 1	5
Cape Smith suite	Z3-84	granite	1898 +12/-9	1
	P-23-87	tonalite	1888 +6/-4	1
	S-162a-86	granite	1876 ± 1.5	1
	DL-5	diorite	1874 +4/-3	3
	Z5-84	tonalite	1870 ± 15	1
	CP-2	diorite	1860 ± 2	3
	CP-5	diorite	1859 ± 2	3
	Z2-84	granite	1848 +6/-5	1
	DL-1	granodiorite	1845 ± 6	3
	D-284-91	granite	1840 ± 2	4
	Z4-84	diorite	1839 +6/-4	1

* References: 1 = Parrish (1989); 2 = St-Onge et al. (1992); 3 = Machado et al. (1993); 4 = Parrish (personal communication, 1994); 5 = Dunphy et al. (in press).

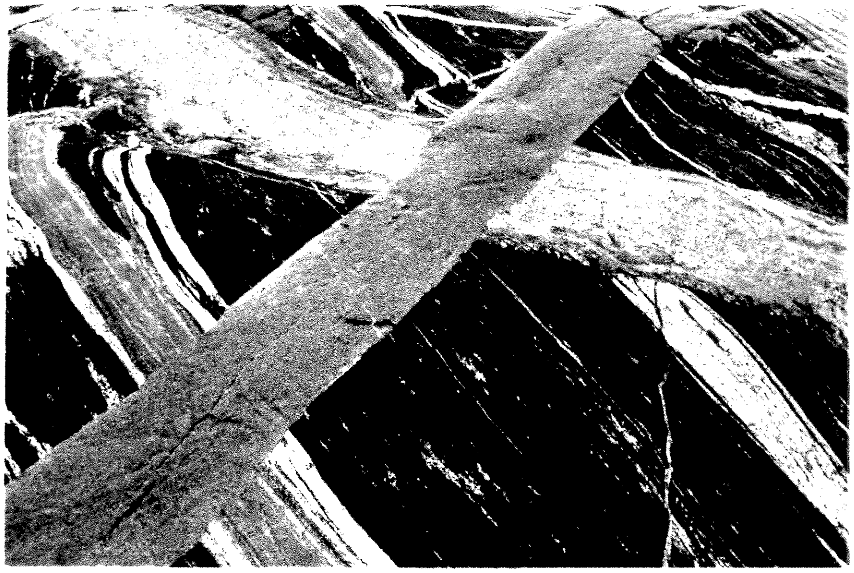


Fig. 1.5. Granitic veins of the Narsajuaq terrane. Several generations of vein are present from deformed layer-parallel and slightly oblique veins of the Older suite to undeformed veins that cross-cut all units and structures (probably related to Younger suite magmatism). Pen for scale.

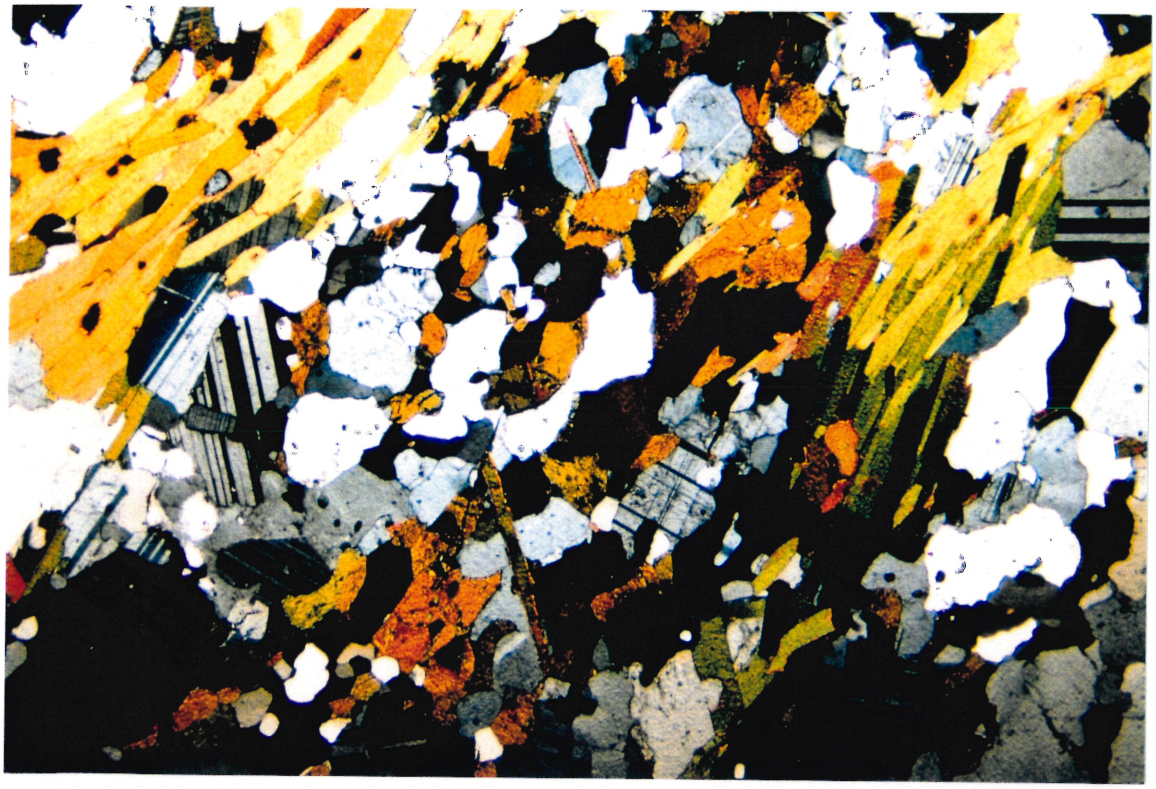
1861 Ma while a less deformed monzogranite vein is 1848 Ma old (Table 1; St-Onge et al. 1992; R. Parrish, personal communication 1994).

Rafts and enclaves of older to comagmatic mafic to ultramafic rocks and older metasediments (Sugluk Group; see below) are occasionally present in the Older suite and concordant sequences of metamorphosed quartz diorite, amphibolite and pyroxenite are found locally (St-Onge et al. 1992). Rare peridotite sills are also found intrusive into the Older suite lithologies, ranging in size from several metres to tens of metres and are variably sheared and deformed (St-Onge and Lucas 1990b). The absolute age of these peridotite bodies is unknown, although they postdate the granite veins but predate emplacement of Late suite syenogranite pegmatite dykes (see below).

Petrographic study of Older suite samples confirms the field observations and has permitted a more detailed study of the mineralogical composition of the rocks sampled for this present study. A complete summary of the mineralogy of the thin section samples can be found in Appendix 1. The diorites contain 25-90% mafic minerals, primarily hornblende and biotite, although clinopyroxene is observed in some samples (primarily as relict cores that are being replaced by hornblende). Minor epidote, tremolite-actinolite and chlorite are present as retrograde products of hornblende and biotite (Fig. 6a,b). No orthopyroxene has been noted in the samples that were studied in thin section as most of the samples collected for the present study were not at granulite grade, although field observations noted the presence of orthopyroxene in variable amounts throughout the unit. Tonalite samples (Fig. 6c) only contained hornblende and/or biotite (15-60%) as the mafic minerals, with trace amounts of epidote. The granitic vein samples are biotite bearing (1-15%) with traces of muscovite and chlorite (Fig. 6d). As was noted during the course of the field work, not all the veins are true granites as the amount of potassic feldspar varies greatly from 0-40%. Apatite, titanite and opaque oxides are ubiquitous accessory phases in all samples.

Fig. 1.6. Representative photomicrographs ((a) plane polarized light and (b) cross polars) of an Older suite diorite (D-41c) containing predominantly hornblende as the mafic mineral, although one retrograde biotite (going to chlorite) is present along the bottom edge of the photo, as well as plagioclase (generally twinned) and quartz. Width of field of view in each photo ≈ 7 mm.

(c)



(d)

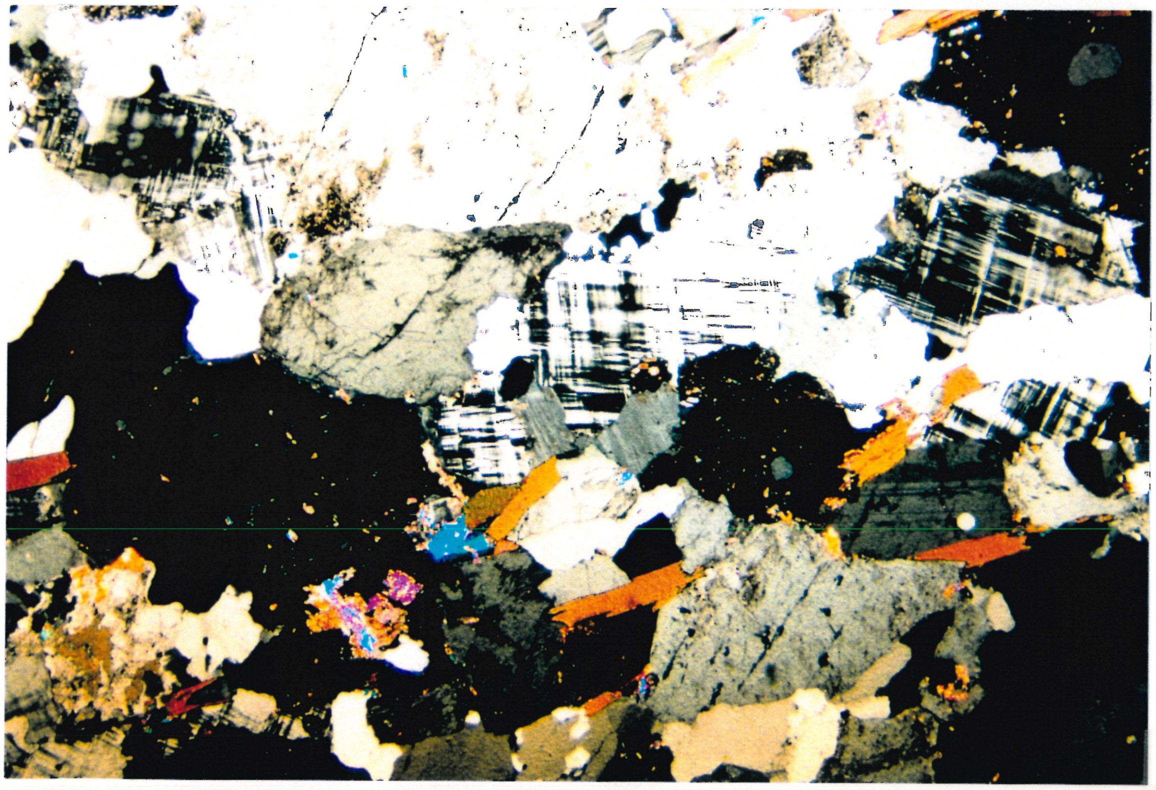


Fig. 1.6. (cont'd) (c) Older suite tonalite (D-211) in which the mafic minerals are segregated into discrete bands of biotite (along the outer edges of the photo) and hornblende (in the middle). Twinned plagioclase and clear quartz are the felsic minerals present; (d) Older suite granite vein (D-23) sample showing typical mineralogy of biotite - plagioclase - microcline - quartz with minor muscovite. Width of field of view in each photo ≈ 7 mm.

1.3.1.2 *Younger suite*

A series of younger plutons, termed the Younger suite (Fig. 3), are intrusive into this base unit (Older suite) and consist of discrete km-size bodies of metamorphosed diorite, monzodiorite, tonalite and granite ranging in age from 1836 to 1800 Ma (Table 1; Parrish 1989; St-Onge et al. 1992; R. Parrish, personal communication 1994). While intermediate compositions are present in the Younger suite, mafic and felsic plutons are more abundant (approximately 15% tonalite, 35% mafic compositions and 50% granite). The Younger suite plutons are variably foliated with the foliations defined by aligned mafic minerals, primarily biotite and hornblende, although locally granulite-facies mineralogies are present (Lucas and St-Onge 1992). The homogeneity of the Younger suite plutons contrasts markedly with the layered aspect of the Older suite and suggests that they were emplaced at higher crustal levels than the Older suite (St-Onge and Lucas 1992).

Plutons of dioritic composition are found predominantly in the southern portion of the map area, near the contact with the Cape Smith Belt (Fig. 3), and range from highly deformed and sheared bodies (particularly those immediately adjacent to the contact with the Cape Smith Belt) to largely undeformed plutons (Fig. 7). The monzodiorite compositions are restricted to two plutons found south of Sugluk Inlet (Fig. 3) and are hornblende - biotite ± clinopyroxene bearing and range from equigranular to k-feldspar megacrystic. A rough compositional polarity is noted in the distribution of the Younger suite plutons as the diorite-monzodiorite plutons are found mainly in the southern limits of the Narsajuaq terrane whereas the more felsic plutons are predominantly located further north (Fig. 3).

The youngest and least deformed plutons are huge (10's km in size) granitic sheets located near Ivujivik in the west and Cap de Nouvelle-France in the east (Fig. 3). These pink weathering granites (referred to as IVUJ and CNF), separated by more than 100 km, are remarkably similar in their overall bulk composition, mineralogy and general field appearance (grain size, deformation) although compositional variations

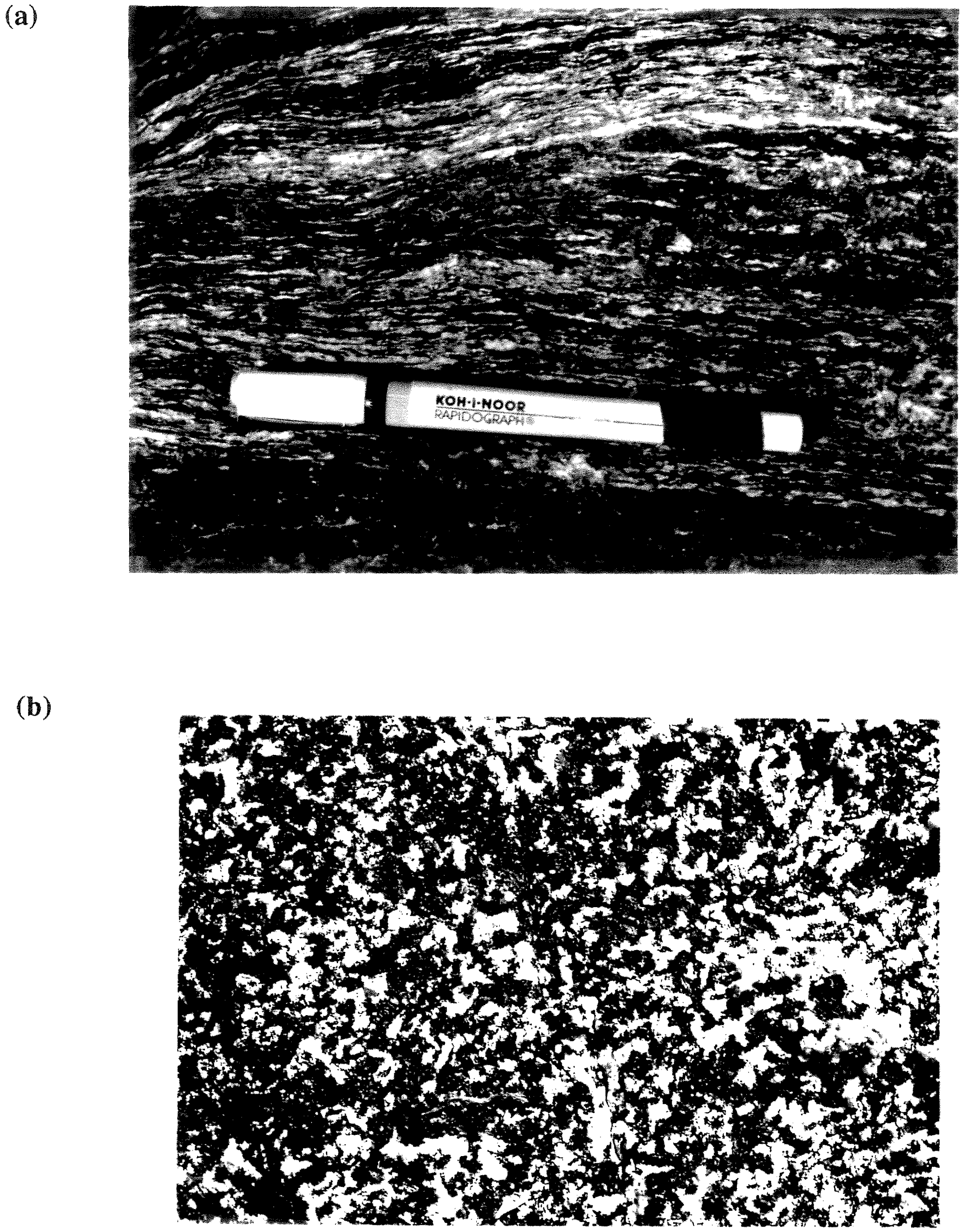


Fig. 1.7. Younger suite diorite plutons preserved in varying deformation states from (a) highly deformed and sheared (located near the terrane bounding fault next to the Cape Smith Belt) to (b) undeformed. Pen for scale in (a) and grain size in (b) is approximately 0.5-1 cm.

within each pluton from granite to potassic feldspar granite and syenite are noted, as well as variations in the grain size (equigranular to k-feldspar megacrystic) (Fig. 8; Lucas and St-Onge 1991; St-Onge and Lucas 1992). A sample from the IVUJ granite has been dated at 1800 Ma while the CNF granite has an age of 1803 Ma (Table 1; R. Parrish, personal communication 1994).

Enclaves of Older-suite lithologies as well as comagmatic pyroxenite, diorite and tonalite and metasedimentary inclusions are found locally in the Younger suite units, particularly in the tonalite and granite plutons located in the northern part of the terrane (Fig. 9; St-Onge et al. 1992). The previously mentioned peridotite sills that are rare in the Older suite are even rarer in the Younger suite, as there is only one diorite pluton that contains such sills.

Thin section study of the Younger suite diorites documents similar compositions as is observed in the Older suite diorites. Hornblende and biotite are the prevalent mafic minerals (50-80%), although several thin sections had minor quantities of clinopyroxene with significant overgrowth of hornblende (Fig. 10a). Epidote is generally a minor phase, although in two samples hornblende is completely replaced by epidote. The monzodiorite plutons are distinct in that they contain 5-50% potassic feldspar and in most samples have distinctive clinopyroxene crystals that have exsolved Fe-Ti oxides on their surface (Fig. 10b). In addition, one sample contains 3 relatively large allanite crystals with spectacular epidote overgrowths (Fig. 10c). Hornblende and biotite are also present in the monzodiorites. The Younger suite granites (IVUJ and CNF) contain very little mafic material (< 10% biotite ± minor hornblende and muscovite) and are generally very quartz-rich (up to 70%) (Fig. 10d). As was observed in the Older suite, all of the Younger suite plutons contain trace amounts of apatite, titanite and opaque oxides.

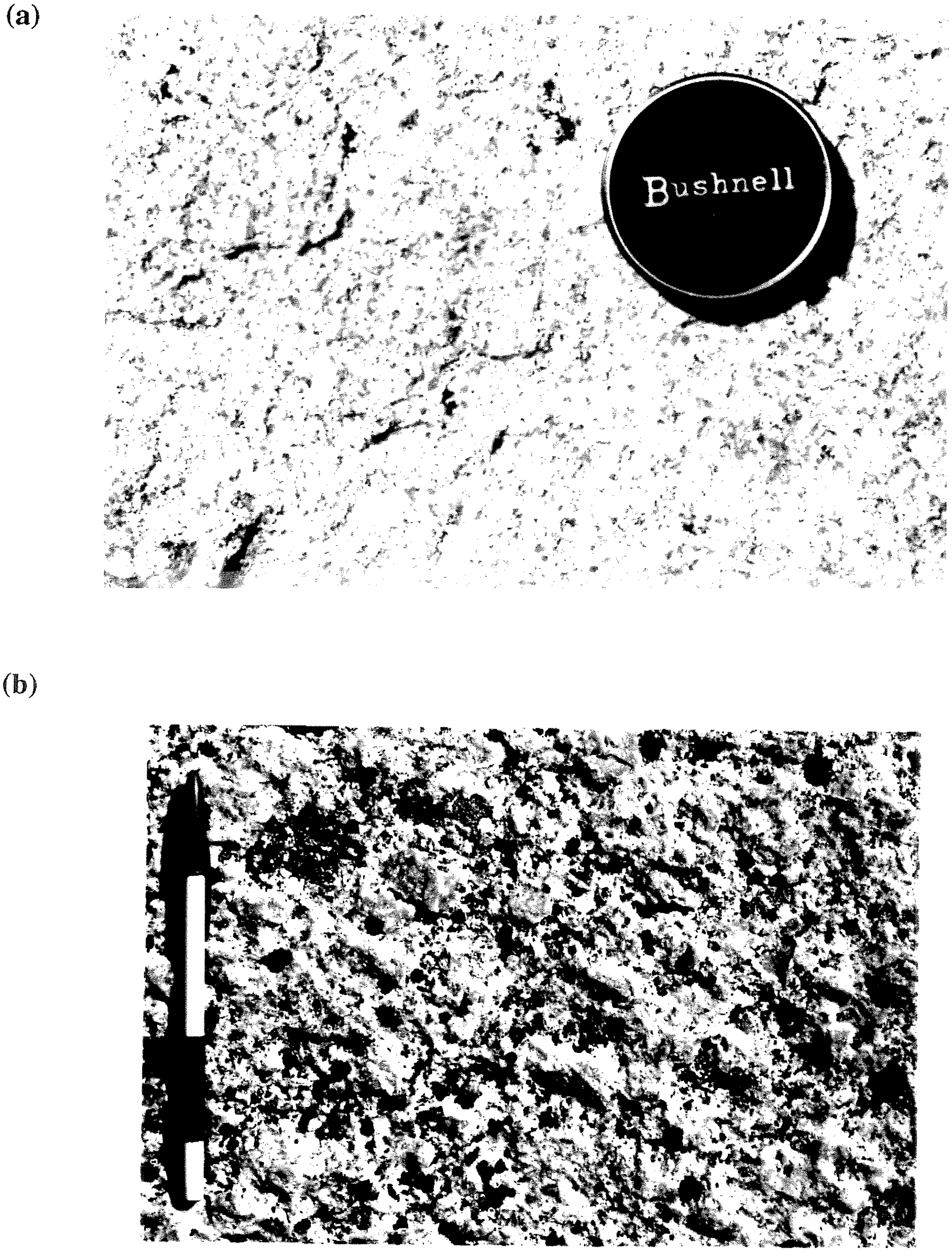


Fig. 1.8. Variation in appearance of the Younger suite granitic pluton (IVUJ) from (a) equigranular fine-medium grained to (b) K-feldspar megacrystic. Lens cap for scale in (a) and pen in (b).

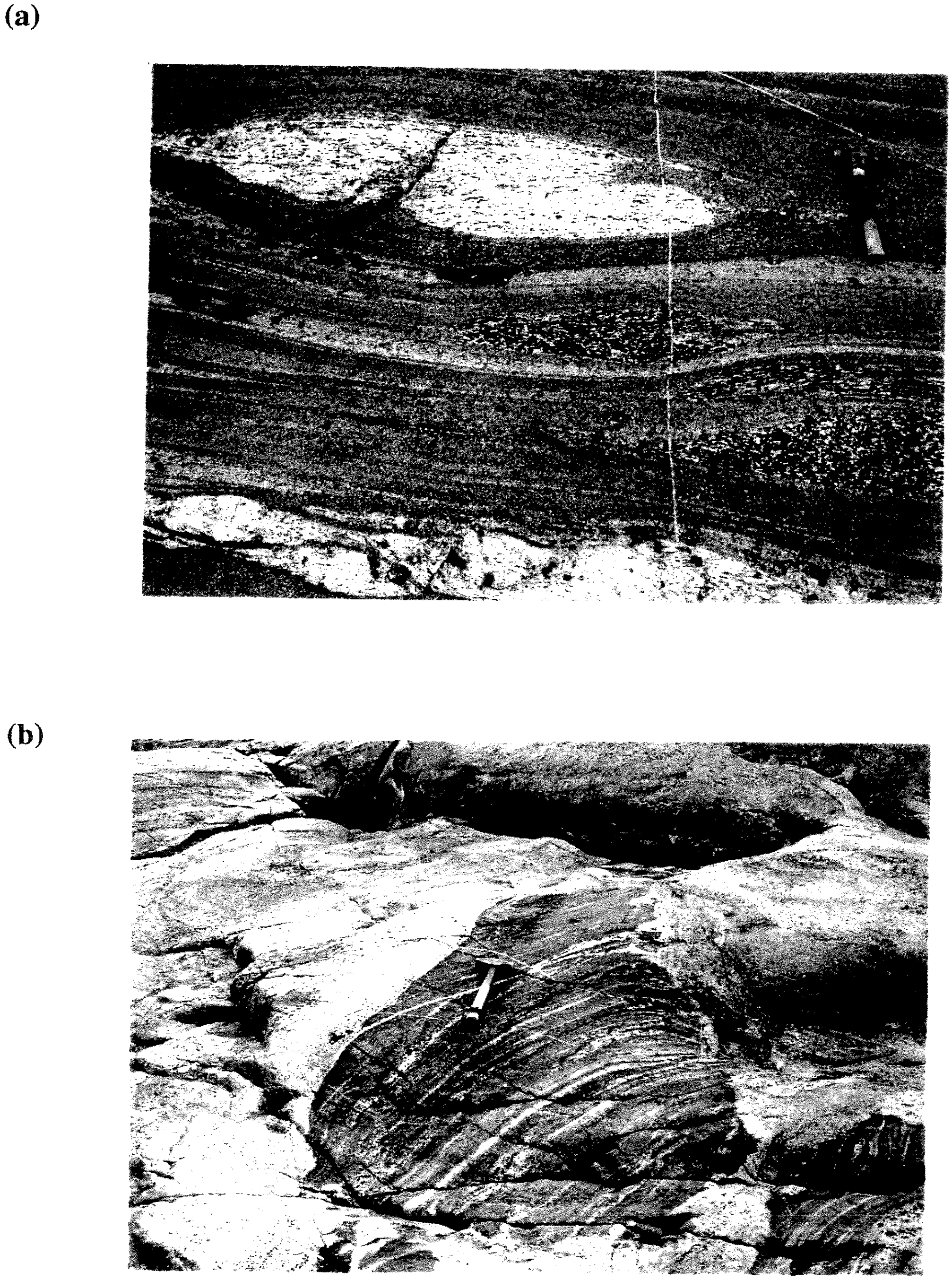


Fig. 1.9. Inclusions in Younger suite plutons: (a) tonalite (white) and diorite (dark) inclusions in a syntectonic Younger suite tonalite (hammer for scale) and (b) inclusion of the Older suite layered diorite-tonalite unit in the younger suite IVUJ granite (hammer for scale).

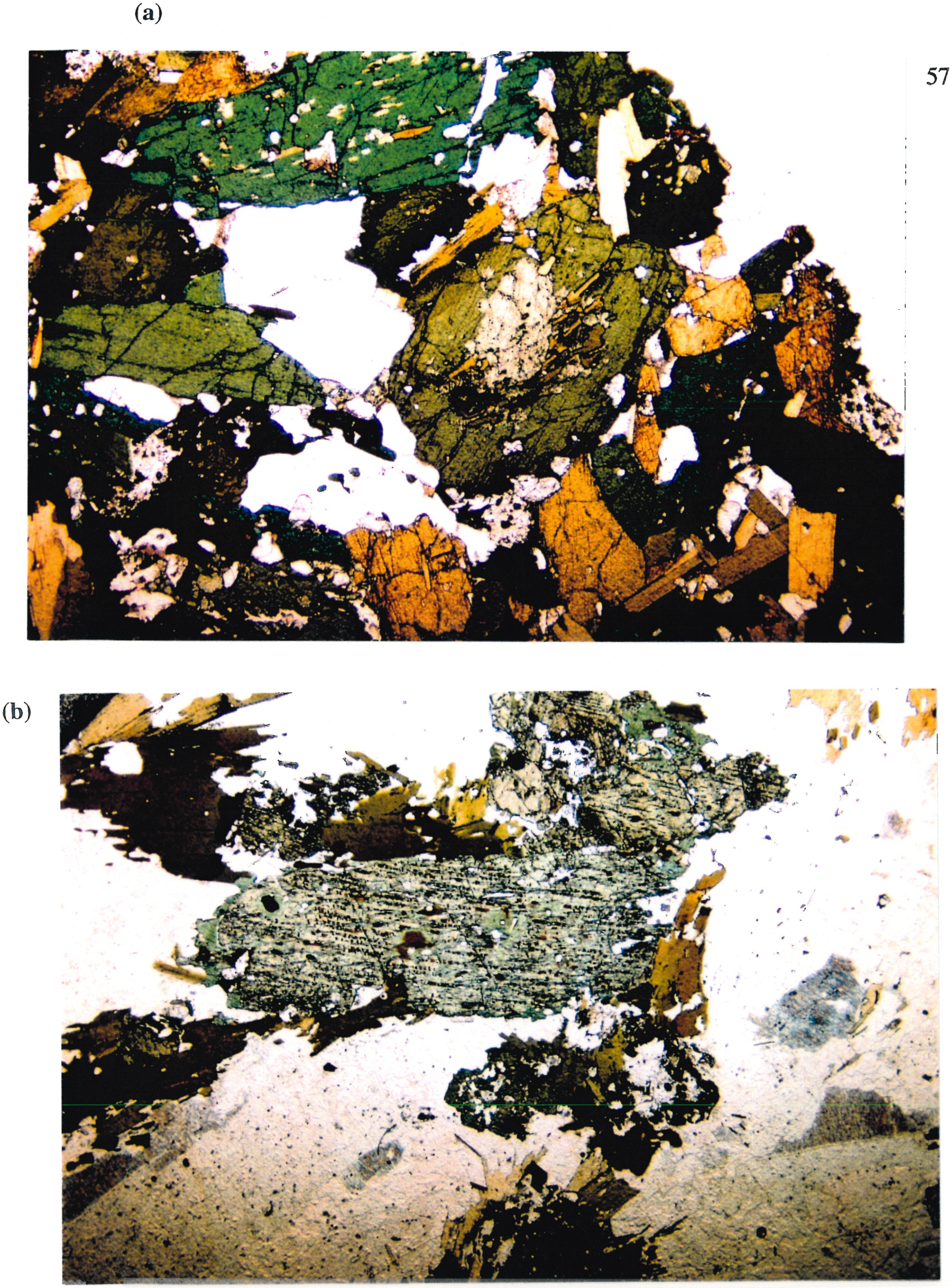
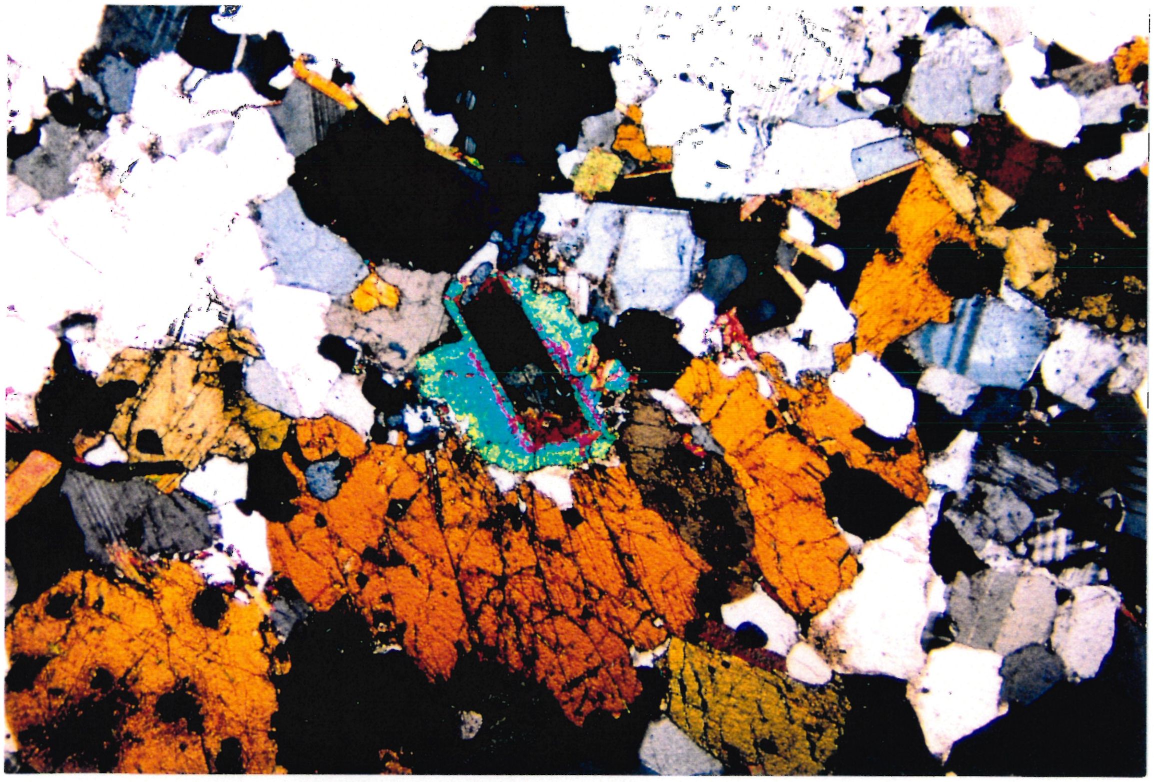


Fig. 1.10. Photomicrographs of Younger suite units: retrograde Younger suite diorite (L-18) in plane polarized light (a) containing clinopyroxene (centre of large grain) which is partially replaced by hornblende (blobs in centre and around rim of grain) and biotite. White areas are quartz grains; (b) Younger suite monzodiorite (L-38a) containing clinopyroxene partially replaced by hornblende, as well as exsolved Fe-Ti oxides (brownish streaks). The white regions are large plagioclase and microcline crystals. Width of field of view in each photo ≈ 7 mm.

(c)



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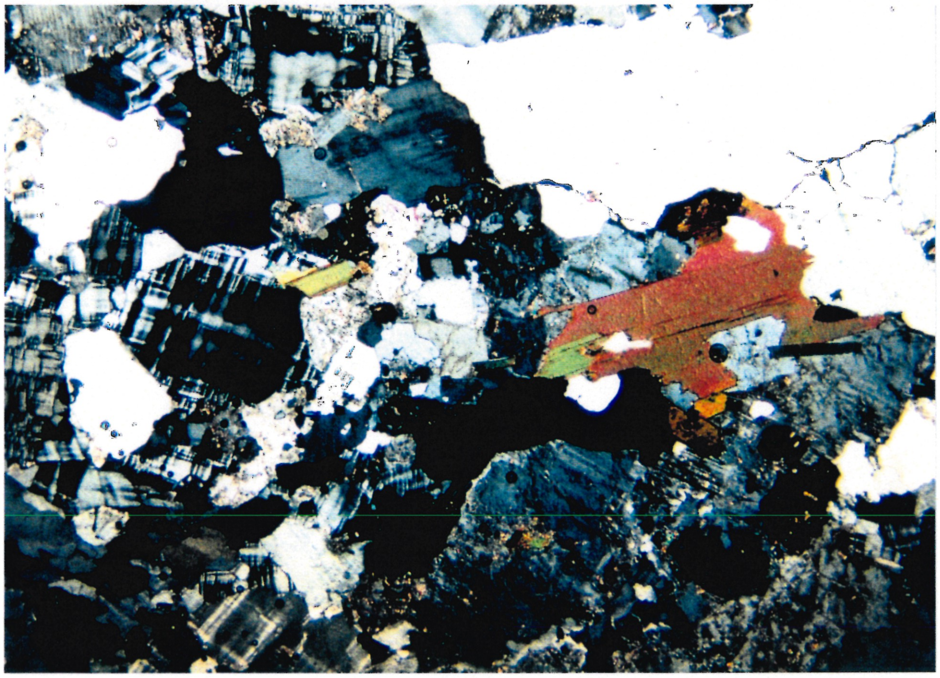


Fig. 1.10. (cont'd) (c) Younger suite monzodiorite (D-20) containing large allanite crystal with spectacular epidote reaction rim. The orange-coloured mineral is clinopyroxene with the remaining grains composed of plagioclase, microcline and minor quartz; (d) typical biotite granite (L-187) from the Younger suite with microcline, plagioclase and quartz dominating the mineralogy. Width of field of view in each photo ≈ 7 mm.

1.3.1.3 *Cape Smith suite*

A series of plutons ranging in composition from diorite to granite are found intrusive into the lithologies of the Cape Smith Belt, notably within the Watts, Spartan and Parent Groups (Fig. 3; Taylor 1982; St-Onge and Lucas 1990a, 1993; Dunphy 1994) and are termed the Cape Smith suite. The plutons occur as distinct bodies having sharp intrusive contacts with their host rocks although in some cases the margins of the plutons are marked by thrust faults. The Cape Smith suite plutons have undergone variable degrees of deformation and are preserved as massive, homogeneous bodies to well foliated and highly strained intrusions. They range in size from small plugs (1 km diameter) to larger compositionally zoned bodies (10 km diameter).

The primary mineralogy of hornblende + biotite ± clinopyroxene has largely been replaced by variable amounts of epidote, tremolite-actinolite and chlorite, and seritization of plagioclase is pervasive (Fig. 11). This mineralogy is typical of greenschist-facies metamorphic conditions and is consistent with the regional metamorphic grade of the host volcanic and sedimentary rocks of the Cape Smith Belt (Bégin 1992; St-Onge and Lucas 1990a, 1993). Thin section study of the plutons from the Cape Smith suite documented the presence of primary orthopyroxene in two samples of dioritic composition. Two samples also contained a noticeable amount of muscovite (2-6 %) which is generally only present in trace amounts in the plutons of the various suites of the Narsajuaq terrane, if at all.

The Cape Smith suite plutons range in age from 1898 - 1839 Ma (Table 1; Parrish 1989; Machado et al. 1993), representing almost 60 Ma of magmatic activity, and partially overlap with the ages of the plutons for the Older suite. This similarity in the timing of magmatism as well as the compositional similarities and close spatial relationship of the Cape Smith suite plutons and the plutons of the Narsajuaq terrane proper suggests that the plutonic events may be related (St-Onge et al. 1992; Dunphy 1994).

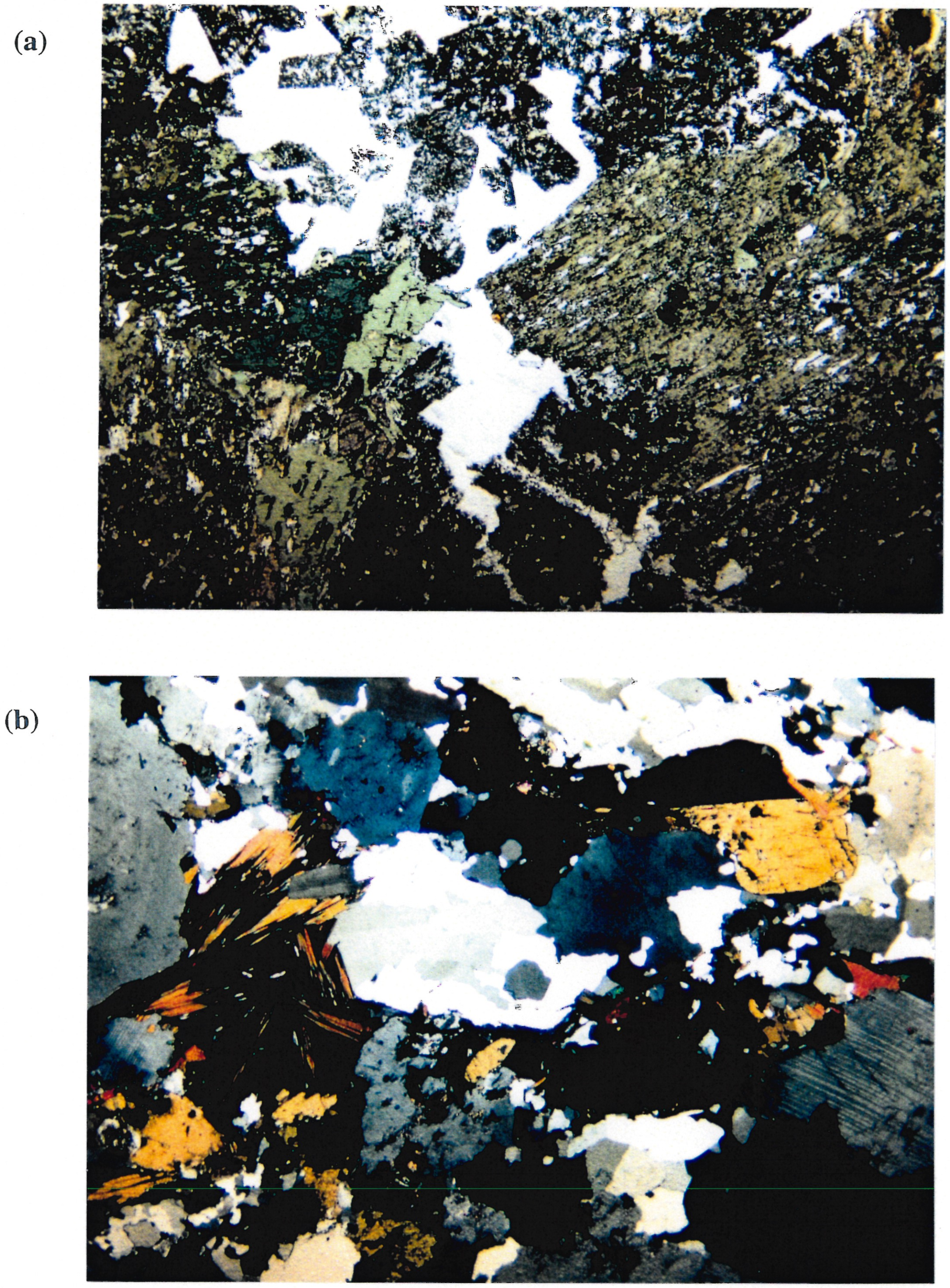


Fig. 1.11. Photomicrographs of Cape Smith suite lithologies: (a) retrograde diorite (D-274) in which the hornblende and biotite are partially replaced by epidote and chlorite (pale green and blue colours) and pervasive seritization of the plagioclase has occurred (darkened rectangluar grains). The large clear white grains are quartz; (b) biotite tonalite (D-284) in which the biotite (yellow-brown colour) is partially replaced by chlorite (blue). Twinned plagioclase and strained quartz make up the rest of the mineralogy. Width of field of view in each photo ≈ 7 mm.

1.3.1.4 *Late suite*

Plutonic rocks of the Late suite consist of small granitic plutons and dykes that were emplaced late in the development of the Narsajuaq terrane, syn- to post-tectonic, corresponding to the final stages of accretion of the Narsajuaq terrane onto the Superior Province margin and the subsequent regional-scale folding (St-Onge et al. 1992; Lucas and St-Onge 1992). The small (<3 km in diameter), rare granitic plutons and sills are concentrated in the Narsajuaq terrane (Fig. 3 - most are not shown due to their small size) whereas the pegmatite dykes cross-cut the entire Ungava Orogen (Fig. 12a; St-Onge et al. 1992). In general the Late suite granites are undeformed and unmetamorphosed and cut all structures in the host rocks, however some dykes record a weak deformation. The dykes vary from less than 0.5 m to greater than 2 m wide (Fig. 12b), with the sills reaching up to 2 km in size. Where these pegmatites intrude granulite-grade lithologies, distinct borders (cm- to m-scale) of retrogression to amphibolite-grade assemblages are observed (St-Onge et al. 1992). The Late suite granites are biotite bearing with trace amounts of muscovite, apatite, titanite, zircon, monazite and opaque phases. A syenogranite dyke has been dated at 1758 Ma (Parrish 1989) while one granitic pluton has an age of 1742 Ma (Table 1) and contains marked zircon inheritance up to 3.1 Ga (Dunphy et al. in press).

1.3.1.5 *Sugluk Group*

In addition to the plutonic rocks, a sequence of sedimentary rocks (termed the Sugluk Group) are present within the Narsajuaq terrane and are exposed primarily in the northern part of the terrane (Fig. 3; St-Onge and Lucas 1990b, 1992). These supracrustal rocks typically occur as elongate lenses and pods of restricted width (<1 km) but substantial strike lengths (up to 65 km).

The metasediments are predominantly graphitic semi-pelites with lesser pelite, quartzite, calc-silicate and mafic volcanic(?) bands. The semi-pelites typically contain garnet - biotite - quartz - plagioclase ± granitic pods with sillimanite occurring in the pelites (Lucas and St-Onge 1992). Locally higher pressure assemblages may be present

(a)

(b)

Fig. 1.12. Late suite syenogranite pegmatite dykes: (a) large-scale view of pervasive pegmatites cross-cutting the other Narsajuaq terrane units (cliff height is approximately 300 m); (b) large dyke cross-cutting a well foliated younger suite tonalite (person for scale).

(see metamorphic section, below). Intense deformation (no primary structures are preserved) and high grade metamorphism of the sediments have made the depositional environment difficult to constrain although St-Onge and Lucas (1992) have suggested that the relative scarcity of shallow water shelf-type deposits suggests deposition occurred in a deep-water environment.

U-Pb geochronology on a Sugluk Group quartzite by Parrish (1989) has documented two populations of detrital grains: a younger group of igneous grains at 1830-1863 Ma and older grains at >2525 Ma with Proterozoic metamorphic overgrowths. A second sample of semipelite has zircon cores with ages >2230 Ma which are overgrown by metamorphic rims at 1825-1829 Ma. These geochronological studies document the mixed provenance of the supracrustal rocks (Archean and Proterozoic detritus) and suggests that rapid deposition, burial and metamorphism (to granulite grade conditions) occurred in some of the samples. One possible scenario to explain such events is deposition of the sediments in a forearc or backarc basin and subsequent subduction and accretion to the bottom of an arc complex. In this scenario intrusion of mafic magmas at 1830 Ma could have been responsible for the granulite-facies metamorphism (Lucas and St-Onge 1992; St-Onge et al. 1992).

The map pattern (Fig. 3) clearly shows that the Younger suite plutons cross cut the supracrustal rocks (which frequently occur as rafts and inclusions within these plutons); however, due to strong deformation the contact of the metasediments and the Older suite plutons is more equivocal. North of Sugluk Inlet both intrusive and tectonic contacts between the metasediments and the Older diorite-tonalite suite have been documented (Fig. 13; St-Onge and Lucas 1990b). Therefore, while the existing geochronology on the Sugluk sediments suggests deposition of a Sugluk Group quartzite post-1830 Ma, the field relationships indicate that older sediments may exist, as they occur interlayered with, and as inclusions within, the Older suite plutons (1863-1844 Ma). The Sugluk Group (or parts of it) may represent country rock roof pendants within the plutonic domain (St-Onge and Lucas 1992).

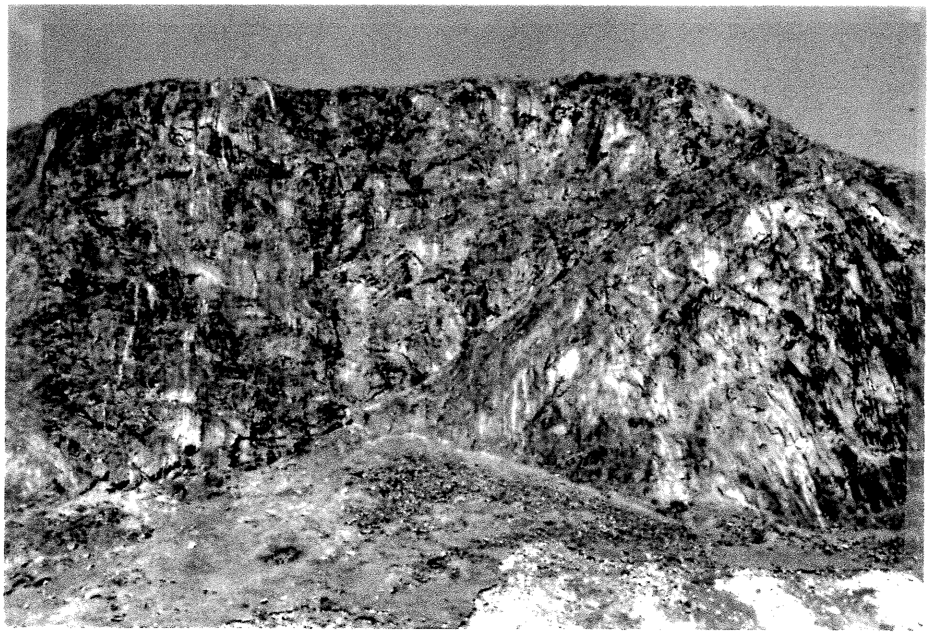


Fig. 1.13. Large fault-bound inclusion of rusty semipelite from the Sugluk Group in a Younger suite granite. The fault is the planar surface dipping approximately 45° from top right to bottom left of photo. Height of cliff approximately 150 m.

1.3.2 Metamorphic history

The plutonic and sedimentary rocks of the Narsajuaq terrane contain mineral assemblages ranging from amphibolite-facies to granulite-facies. All units however, locally contain assemblages consistent with granulite-facies conditions (Lucas and St-Onge 1992). Subsets of the maximum phase assemblage orthopyroxene - hornblende - biotite - plagioclase ± clinopyroxene ± garnet ± quartz (± k-feldspar in the granites) are found in the plutonic rocks whereas the Sugluk Group metasediments contain a variety of metamorphic mineral assemblages depending on the composition of the rock. For example, pelites have a maximum assemblage of sillimanite - hypersthene - cordierite - k-feldspar - garnet - biotite - plagioclase - quartz - granitic pods whereas the quartzites are garnet - orthopyroxene ± clinopyroxene bearing.

Geothermometric and geobarometric studies of the granulite grade lithologies in the Narsajuaq terrane has documented the presence of a P-T discontinuity which correlates with a zone of high strain and repetition of a sequence of the Sugluk Group supracrustal rocks (Monday 1994). The P-T measurements for the northern block of the terrane increase from N to S from 6.4-6.9 kbar and 800-825°C in the north to 9.2-9.9 kbar and 860-900°C in the south, then decrease across the high strain zone to 7.0-8.9 kbar and 700-775°C in the southern block (Fig. 14; Monday 1994). Monday (1994) interpreted these data as indicating the presence of north-side-up (thrust) fault that had not been identified during mapping that is situated along the linear band of metasediments in the central region of the terrane trending NW from Sugluk Inlet (Fig. 14). U-Pb dating of metamorphic zircon from Narsajuaq terrane samples has yielded ages of 1830-1825 Ma (Parrish 1989), corresponding to the waning stages of plutonic activity in the Narsajuaq terrane.

Variable overprinting and retrogression of the high-grade granulite-facies mineral assemblages to amphibolite-facies is observed which occurred during the tectonic events related to the development and amalgamation of the Cape Smith Thrust and Fold Belt (Lucas and St-Onge 1992; Scott and St-Onge, in press). Both the rocks of the Narsajuaq

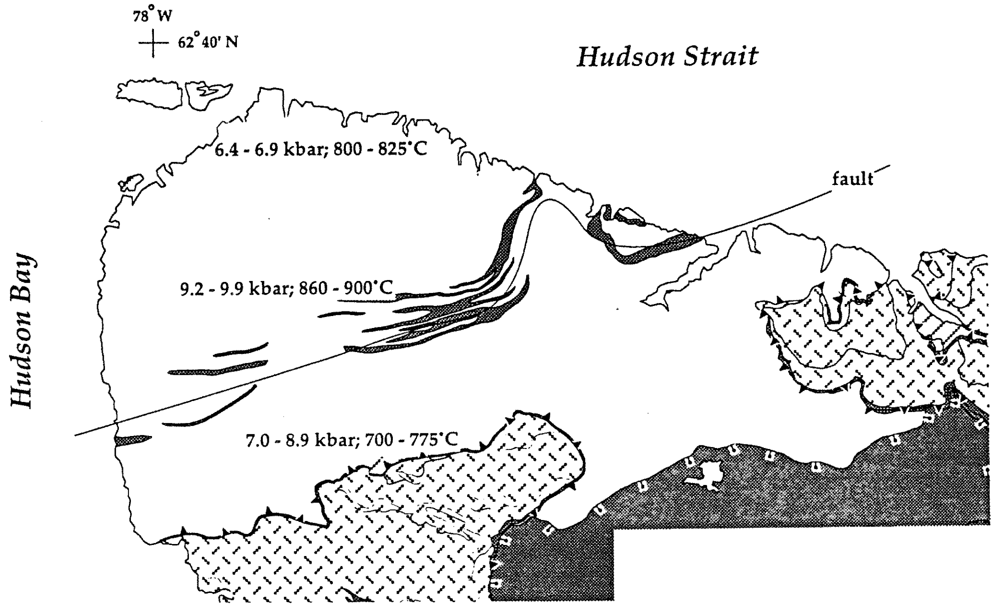


Fig. 1.14. Highly simplified map of the Narsajuaq terrane showing approximate location of proposed fault across the metamorphic discontinuity and corresponding ranges in P-T measurements for the different tectonic blocks Monday (1994).

terrane and the underlying Superior Province basement underwent retrogression as indicated by several observations (Lucas and St-Onge 1991, 1992; Monday 1994; St-Onge and Lucas 1995): (1) the granulite assemblages are preserved (albeit locally) in all units, even those (particularly in the south, near the contact with the Cape Smith Belt) that predominantly contain amphibolite assemblages; (2) retrograde mineral reaction textures are preserved in the amphibolite-facies rocks and include hornblende rimming clinopyroxene and orthopyroxene, epidote-actinolite-tremolite replacing hornblende, and chlorite replacing biotite; and (3) distinct mineral zones have been developed that have a consistent spatial relationship to the source of the fluids, the Cape Smith Belt.

Overthrusting of the relatively dry granulite-facies rocks of the Narsajuaq terrane onto the dehydrating continental margin sequence (Povungnituk and Chukotat groups) has been proposed as the mechanism responsible for the retrogression in the Narsajuaq terrane (Lucas and St-Onge 1992). Similarly, thrusting of the same units onto the Archean basement resulted in the hydration and amphibolitization of the Superior Province basement (St-Onge and Lucas 1995). Fluid flow from the Cape Smith Belt rocks was upward into the overlying Narsajuaq terrane rocks and downward into the underlying Superior Province basement, resulting in synchronous retrograde metamorphism in both units (St-Onge and Lucas 1995; M. St-Onge, personal communication 1995). U-Pb dating of coronitic titanite from the re-equilibrated basement rocks in the Ungava Orogen has yielded ages of ca. 1.79-1.81 Ga, interpreted as constraining the timing of the retrogression event (Scott and St-Onge, in press). These ages correspond to the timing of terminal collision between the Narsajuaq terrane and the underthrust Superior Province basement (St-Onge et al. 1992, Lucas and St-Onge 1992).

A second retrogression event in the Narsajuaq terrane is related to intrusion of the Late suite syenogranite pegmatite dykes (1758 Ma; Parrish 1989) which produced spectacular amphibolite facies retrogression halos (cm- to m-scale) adjacent to the dykes in areas where the host rocks were at granulite grade (St-Onge and Lucas 1990b; St-

Onge et al. 1992).

1.3.3 Structural history

The Narsajuaq terrane underwent four main phases of deformation as was previously mentioned in the section on regional geology. The oldest structure in the Narsajuaq terrane is related to D_1 and is manifested by a gneissic foliation defined by a compositional layering of tonalite-granodiorite, quartz diorite and granite (Fig. 4). Within individual layers the foliations are defined by aligned metamorphic minerals (hornblende + biotite ± clinopyroxene ± orthopyroxene) and/or deformed aggregates of plagioclase ± quartz ± k-feldspar (Lucas and St-Onge 1992). Syntectonic intrusion of the granitic veins is believed to have resulted in an enhancement of the compositional layering (Lucas and St-Onge 1995). An E-W trending, moderately plunging lineation is present in the D_1 foliation, and is primarily defined by quartz-feldspar rodding, although locally it is defined by orthopyroxene, clinopyroxene, or sillimanite (Lucas and St-Onge 1992). The alignment of the metamorphic minerals documents synchronous nature of the metamorphism and deformation. Although scarce, a few examples of kinematic indicators associated with D_1 have been noted. Dextral shear is indicated by asymmetric rotated feldspars, folded granite veins and rotated mafic boudins (Lucas and St-Onge 1992). These kinematic indicators and the moderately plunging lineation led Lucas and St-Onge (1992) to conclude that the Narsajuaq terrane underwent dextral transcurrent shear during D_1.

The D_1 deformation was heterogenous, however, throughout the entire terrane, as areas of high penetrative strain (indicated by the development of straight gneisses) are found within the same outcrop as rocks which record lower bulk strains (Fig. 15). In general, the oldest rocks in the Narsajuaq terrane record higher strains whereas the Younger plutons have undergone less bulk strain as evidenced by the absence or lack of straightening deformation in enclaves and the less foliated nature of the plutons. D_1 is constrained to have terminated prior to 1821 Ma, the age of the youngest undeformed pluton that cross-cuts the D_1 foliation (St-Onge et al. 1992; R. Parrish, personal com-

(a)

(b)

Fig. 1.15. Heterogenous strain in the Older suite units related to D_1 in the Narsajuaq terrane. In some areas of low strain (a), intrusion of syntectonic granitic veins (light) disrupts the gneissic layering of the diorite-tonalite unit (dark), whereas other regions have experienced much higher strains (b) resulting in a highly deformed gneissic rocks.

munication 1994).

Reworking of the D_1 foliation and the larger-scale gneissosity occurred during deformation associated with the D_2 event. Reorientation of the D_1 fabric (through 90° over a narrow zone (metre-scale) in some cases) into parallelism with the main D_2 fabric in the underlying Cape Smith Belt lithologies is the most obvious manifestation of the D_2 event in the Narsajuaq terrane (Lucas and St-Onge 1992). Faults that are inferred to be of D_2 age are found north of Sugluk Inlet based on the repetition of intrusive contacts between Narsajuaq plutons and Sugluk metasediments and the presence of high strain zones along the presumed faults (Lucas and St-Onge 1992). D_2 stretching lineations overprint the D_1 lineations and are more steeply plunging and defined by amphibolite facies minerals (hornblende - biotite) and/or quartz-feldspar rodding. The D_2 lineations are approximately parallel to those recorded within the Cape Smith Belt units (Lucas and St-Onge 1992). The D_2 event was responsible for the accretion of the Narsajuaq terrane southward over the Superior Province basement and the final amalgamation of the Cape Smith Belt (Lucas 1989; St-Onge and Lucas 1990a, Lucas and St-Onge 1992). This D_2 accretion event is constrained to post-date Older and Younger suite magmatism within the Narsajuaq terrane and pre-date intrusion of the Late suite syenogranite pegmatite dykes that stitch all the tectonic domains together.

Two subsequent deformation events, D_3 and D_4, were of a regional-scale and affected all elements of the Ungava Orogen. D_3 is manifest as the broad E-W Kovik antiform and generated the synclinorium in which the Cape Smith Belt is preserved (Fig. 2; Hoffman 1985). On a smaller scale D_3 produced asymmetrical east-trending folds (south verging) that are preserved in all tectonic units (St-Onge and Lucas 1990a; Lucas and St-Onge 1992). This folding episode post-dates the D_2 accretion event as all D_2 thrust faults are deformed by D_3 east-trending structures (St-Onge and Lucas 1990a; Lucas and St-Onge, 1992). Structural analysis of the orientations of conjugate sets of pegmatite dykes suggests that the dykes were intruded during a N-S orientated deformation event, corresponding to D_3 (Lucas and St-Onge 1992), thus constraining D_3

to earlier than or at 1758 Ma, the age of the dykes (Parrish 1989).

Refolding of all previous structures into NNW upright folds during D_4 gave rise to the map-scale dome and basin interference pattern and easily visible in the map trace of the faults separating the Narsajuaq terrane from the Superior Province basement (Fig. 3; Lucas and St-Onge 1992). It was this particular deformation episode that was responsible for generating the significant structural relief presently preserved in the Orogen (Lucas 1989; St-Onge and Lucas 1990a). An undeformed granitic pluton cutting a north-trending D_4 synform indicates that this youngest deformation episode predated its emplacement, dated at 1742 Ma (Lucas and St-Onge 1992; Dunphy et al. in press).

1.4. GEOCHEMISTRY

1.4.1 Analytical techniques

Samples for geochemical and isotopic analysis were chosen to be free of alteration, homogenous and representative of the lithology under study. For geochemical analysis most samples were crushed and powdered in an agate mill and shatterbox, although some were processed in a tungsten carbide shatterbox. Major and some trace elements were determined by X-ray fluorescence (XRF) using the methods of Norrish and Hutton (1969) and Schroeder et al. (1980) at the Université de Montréal and at McGill University. Rare earth elements (REE) and other trace elements were analyzed by instrumental neutron activation analysis (INAA) also at the Université de Montréal. A series of samples from the Cape Smith suite were analyzed at the Centre de Recherche Minéral du Québec for the Ministère des Ressources Naturelles du Québec. For these samples the major elements were determined by XRF, trace elements by inductively coupled plasma atomic emission spectrometry (ICP-AES) and REE's by INAA. More complete descriptions of the analytical methods are presented in Appendix 2. Representative analyses of the Narsajuaq terrane samples are presented in Table 2 and the complete geochemical data set of a total of 92 samples from the Narsajuaq terrane is found in Appendix 3. Sample locations are provided in Appendix 4. Routine precision for XRF data is better than 1% for the major elements and 5% for the trace elements.

Table 2. Representative geochemical analyses for the Narsajuaq terrane plutonic suites.

	Older Suite												Younger Suite									
Sample	D-41c	L-90a	L-90b	L-211	S-18a	M-103	D-72	D-163	D-211	D-14a	D-29	D-41a	D-26	D-28	S-106	K-69	D-20	D-24	D-43	L-118	D-283	L-205
Lithology	diorite	diorite	diorite	diorite	qtz dior	mzdior	tonalite	tonalite	grano	grano	tonalite	granite	diorite	diorite	diorite	diorite	mzdior	qtz mzdior	mzdior	tonalite	IVUJ gran	IVUJ gran
SiO2 (wt %)	52.45	53.83	49.58	47.34	54.30	56.34	67.67	62.66	65.19	73.81	74.85	75.44	52.15	52.63	45.38	50.83	55.85	60.38	57.73	66.10	69.98	70.27
TiO2	1.03	1.04	0.95	0.81	0.84	0.85	0.47	0.84	0.51	0.08	0.04	0.08	0.75	0.58	1.54	1.45	0.93	0.69	0.70	0.50	0.50	0.32
Al2O3	16.65	20.25	16.19	14.64	19.01	16.79	16.24	18.04	15.34	15.03	15.11	14.45	15.59	16.65	15.32	15.59	18.19	15.26	14.65	16.95	14.29	15.27
Fe2O3	10.13	7.85	10.67	12.48	8.15	6.95	3.71	5.54	5.38	0.97	0.56	1.09	8.73	8.27	14.86	12.59	6.80	5.56	5.73	3.77	3.32	2.69
MnO	0.16	0.09	0.18	0.23	0.16	0.09	0.06	0.05	0.08	0.01	0.01	0.01	0.18	0.15	0.24	0.18	0.10	0.09	0.09	0.03	0.02	0.02
MgO	4.52	2.96	7.33	8.97	3.00	3.90	1.71	1.76	2.18	0.01	0.02	0.04	6.68	6.57	7.83	5.68	3.80	3.47	4.50	1.21	0.41	0.51
CaO	8.55	6.01	9.31	11.19	6.49	5.06	3.23	5.07	3.75	1.81	1.29	1.67	8.96	9.18	10.33	8.46	6.09	4.98	6.41	3.27	1.26	2.02
Na2O	4.25	5.15	3.81	2.11	4.64	4.37	4.14	4.99	3.47	4.32	6.48	3.70	3.54	3.75	2.72	3.70	5.10	4.30	4.31	4.61	3.09	3.55
K2O	0.87	2.05	1.28	0.41	1.73	4.07	2.11	1.23	2.83	3.68	1.11	3.83	0.71	0.65	1.56	1.03	2.86	3.93	4.03	3.28	6.07	4.71
P2O5	0.28	0.33	0.24	0.10	0.28	0.65	0.15	0.29	0.12	0.02	0.01	0.04	0.32	0.15	0.23	0.23	0.58	0.49	0.85	0.16	0.14	0.14
LOI	0.28	0.28	0.40	1.68	0.85	0.86	0.64	0.47	0.39	0.33	0.27	0.37	1.13	0.97	0.52	0.88	0.60	0.33	0.36	0.64	0.37	0.25
TOTAL	99.17	99.84	99.94	99.96	99.45	99.93	100.13	100.94	99.24	100.07	99.75	100.72	98.74	99.55	100.53	100.62	100.90	99.48	99.36	100.52	99.45	99.75
Th (ppm)	1.7	0.9	0.7	1.4	1.5	nd	6.3	4.49	7.67	8.68	0.16	12.0	0.8	0.4	1.31	0.42	7.1	14.0	5.6	nd	1.21	24.02
U	2.4	0.7	0.2	1.17	0.9	nd	0.8	0.13	0.83	1.46	0.41	1.8	0.3	0.2	0.53	0.61	1.5	2.4	1.4	nd	0.36	0.75
Rb	6.7	37.0	15.2	5.5	40.0	84.7	61.2	13.6	90.1	76.9	80.0	80.7	5.6	8.8	18.5	10.0	34.6	93.2	88.9	60.3	117.8	142.4
Sr	660	1398	836	145	928	1652	641	899	478	405	576	389	1305	907	305	372	1910	1217	2145	568	402	462
Y	22.5	14.8	27.9	16.5	18.2	15.6	10.0	9.2	8.1	2.9	2.4	4.2	17.5	12.7	31.9	26.0	19.9	19.5	21.3	7.0	12.3	8.8
Zr	90	384	134	51	203	307	152	288	99	180	53	187	68	47	99	145	261	227	186	98	524	470
Nb	8.2	5.6	8.0	4.7	10.0	14.7	4.5	7.7	10.1	2.4	8.0	3.2	5.6	3.0	9.7	8.6	13.3	15.3	11.6	5.8	10.1	6.6
Ba	374	1489	542	124	1012	nd	824	1223	1506	951	955	1106	1045	445	543	249	2240	1549	2662	nd	2507	1649
Hf	2.40	9.50	3.70	1.36	5.70	nd	4.50	6.31	2.58	4.91	1.27	6.20	1.50	1.30	2.43	3.25	5.90	5.40	3.80	nd	12.74	13.44
Ta	2.1	1.3	1.0	0.24	1.6	nd	1.5	nd	0.35	nd	nd	2.5	0.7	0.8	0.36	0.27	2.1	1.5	1.7	nd	0.25	nd
Cr	115.8	70.1	230.6	nd	22.1	nd	33.6	10.2	nd	6.8	3.9	12.1	358.1	210.7	nd	nd	77.6	90.3	113.4	nd	nd	nd
Ni	45.1	29.7	145.8	nd	11.3	nd	10.3	10.2	nd	4.9	4.8	4.9	83.9	59.3	84.9	nd	47.0	43.3	70.1	21.0	6.4	nd
Zn	98.0	99.7	114.4	nd	110.3	nd	62.8	97.2	nd	19.4	19.5	15.7	102.4	79.4	150.0	nd	112.0	84.5	92.0	54.3	53.1	nd
Cu	14.6	6.7	105.1	nd	38.9	nd	4.8	24.2	nd	4.2	3.9	2.8	13.5	8.1	16.9	nd	8.7	52.8	89.3	6.1	5.1	nd
Co	35.3	22.6	44.3	48.3	35.2	nd	28.8	10.9	14.0	0.8	0.5	52.6	34.3	38.8	43.1	41.0	28.4	33.0	27.1	nd	3.3	3.9
Sc	27.21	16.44	27.41	34.19	15.41	nd	7.60	nd	11.18	nd	nd	0.90	22.69	23.80	39.81	25.63	11.08	10.82	11.05	nd	1.50	1.69
Cs	0.2	0.6	0.5	nd	1.0	nd	1.7	nd	0.9	nd	nd	1.2	0.2	0.5	0.35	0.47	0.5	1.7	1.3	nd	0.27	0.21
W	397	304	140	24	334	nd	355	nd	39	1	nd	602	141	174	13	21	477	257	355	nd	2	nd
Au	0.002	0.002	0.002	0.009	0.002	nd	0.001	0.011	0.007	0.001	0.002	0.005	0.004	0.001	nd	nd	0.007	0.003	0.003	nd	nd	0.01
La	8.90	12.70	15.77	13.36	20.16	nd	31.08	83.84	18.93	15.03	0.50	9.50	32.85	10.41	24.94	14.84	77.00	65.23	105.10	nd	32.49	75.80
Ce	21.40	26.70	38.83	25.50	46.09	nd	62.62	159.23	45.47	22.26	1.02	23.17	66.43	23.10	50.14	32.98	160.80	132.96	209.70	nd	52.85	117.17
Nd	15.00	17.60	24.48	9.66	23.70	nd	26.87	56.17	10.93	7.83	0.41	6.14	33.04	14.71	29.17	16.20	78.20	61.40	107.10	nd	17.87	32.87
Sm	3.38	3.37	5.06	2.76	4.04	nd	3.88	7.50	2.67	1.68	0.09	1.03	6.47	2.93	6.56	5.09	11.76	10.86	17.11	nd	3.98	4.49
Eu	1.07	1.20	1.32	0.63	1.30	nd	1.02	1.86	1.04	0.58	0.10	0.56	1.92	1.07	1.93	1.51	2.75	2.34	3.74	nd	1.68	1.13
Tb	0.60	0.49	0.81	0.31	0.55	nd	0.34	0.45	0.23	0.06	0.02	0.08	0.68	0.37	0.87	0.75	0.85	0.91	1.09	nd	0.37	0.39
Ho	0.84	0.99	1.22	nd	0.93	nd	0.52	0.42	0.22	0.03	0.03	0.42	0.84	0.37	1.60	1.39	1.08	0.87	1.05	nd	nd	nd
Yb	2.30	1.50	2.83	1.44	1.86	nd	0.97	0.82	1.17	0.38	0.30	0.62	1.75	1.34	2.99	2.43	1.40	1.81	1.70	nd	1.07	0.94
Lu	0.32	0.21	0.40	0.21	0.30	nd	0.18	0.14	0.17	0.06	0.03	0.10	0.27	0.19	0.43	0.39	0.16	0.23	0.20	nd	0.18	0.16
A/CNK	0.71	0.94	0.66	0.60	0.89	0.81	1.08	0.96	0.98	1.05	1.06	1.09	0.68	0.71	0.61	0.69	0.81	0.75	0.63	0.99	1.02	1.05
(La/Yb)n	2.8	6.1	4.0	6.7	7.8	nd	23.0	73.3	11.6	28.4	1.2	11.0	13.5	5.6	6.0	4.4	39.5	25.9	44.3	nd	21.8	57.8
(La/Sm)n	1.7	2.4	2.0	3.1	3.2	nd	5.2	7.2	4.6	5.8	3.6	6.0	3.3	2.3	2.5	1.9	4.2	3.9	4.0	nd	5.3	10.9

					Late Suite						Cape Smith Suite								
Sample	S-217	L-181	L-187	S-154a	L-67b	D-19	S-33	D-259	L-33	S-218	D-273	D-274	D2-2	D-271	D-275	D-284	D-42	D-240	D-278
Lithology	IVUJ gran	CNF gran	CNF gran	CNF gran	granite	granite	granite	syeno	syeno	syeno	diorite	qtz dior	diorite	tonalite	tonalite	tonalite	granite	grano	grano
SiO2 (wt %)	74.14	72.31	70.08	77.05	74.01	74.26	75.07	72.36	71.25	74.25	50.82	55.80	48.50	66.51	68.41	65.29	72.14	71.41	71.27
TiO2	0.13	0.04	0.60	0.18	0.14	0.10	0.14	0.11	0.10	0.08	1.14	1.05	1.67	0.42	0.36	0.77	0.09	0.20	0.10
Al2O3	13.95	16.63	14.25	12.89	14.37	14.77	12.70	14.65	16.44	14.64	15.92	17.54	16.30	17.13	14.54	16.36	15.33	15.89	16.40
Fe2O3	1.25	0.29	3.58	1.09	1.10	0.85	1.29	1.97	0.85	0.61	11.80	10.76	13.10	3.60	4.56	4.94	0.73	1.75	0.93
MnO	0.01	0.01	0.10	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.20	0.29	0.25	0.05	0.10	0.07	0.02	0.03	0.01
MgO	0.12	0.09	0.45	0.43	0.24	0.19	0.14	0.09	0.12	0.08	6.42	1.78	4.20	1.20	0.69	1.61	0.15	0.48	0.10
CaO	1.06	2.07	2.61	1.86	1.20	1.22	0.54	1.07	1.84	1.22	9.51	6.81	8.05	3.29	2.91	4.67	1.21	2.32	1.32
Na2O	3.61	4.94	3.38	3.47	3.45	3.83	2.55	3.47	4.36	3.42	3.31	3.47	4.23	4.68	3.74	4.84	6.02	5.06	6.68
K2O	4.87	3.44	4.39	3.16	4.61	4.52	5.75	5.32	3.98	5.35	0.74	0.86	0.53	1.37	2.30	0.56	3.05	2.28	2.74
P2O5	0.04	0.01	0.25	0.04	0.05	0.04	0.04	0.03	0.04	0.04	0.39	0.26	0.23	0.14	0.10	0.23	0.04	0.07	0.04
LOI	0.27	0.32	0.27	0.46	0.40	0.20	0.20	0.33	0.24	0.32	0.02	1.61	1.07	1.59	2.02	0.78	0.19	0.50	0.16
TOTAL	99.45	100.15	99.96	100.65	99.58	99.99	98.44	99.41	99.23	100.02	100.27	100.23	98.13	99.98	99.73	100.12	98.97	99.99	99.75
Th (ppm)	8.23	0.24	4.62	2.52	9.6	12.3	9.3	7.73	3.2	1.48	nd	2.0	0.6	nd	nd	3.2	0.4	2.5	nd
U	0.67	0.14	0.53	0.23	2.1	1.5	1.4	0.92	0.7	0.27	nd	0.9	0.5	nd	nd	0.6	0.3	0.5	nd
Rb	98.2	34.1	92.5	72.3	98.3	97.2	172.3	94.1	62.4	112.3	9.1	18.9	10.0	40.7	46.9	10.4	44.1	48.7	33.2
Sr	274	650	530	312	310	259	110	777	765	245	881	407	470	776	291	604	1006	687	1330
Y	1.9	1.0	73.1	4.3	2.7	3.2	7.5	1.6	2.8	3.0	22.9	32.1	30.0	5.7	29.3	12.5	2.6	4.1	2.2
Zr	123	17	490	105	133	106	139	283	76	51	87	110	110	151	168	285	31	109	56
Nb	4.2	1.3	32.7	3.7	8.3	6.5	8.9	2.5	2.3	4.8	7.9	8.9	10.0	6.6	11.3	16.1	7.8	8.2	8.8
Ba	773	2200	2913	1371	1615	1145	747	3408	2801	959	nd	268	116	nd	nd	446	2073	986	nd
Hf	3.18	0.21	13.53	2.70	4.10	3.40	5.10	5.08	1.90	1.94	nd	2.70	2.90	nd	nd	6.43	0.70	2.31	nd
Ta	nd	nd	nd	nd	6.6	3.1	3.2	0.06	2.0	0.08	nd	0.4	1.0	nd	nd	0.5	1.8	0.4	nd
Cr	nd	5.1		7.0	4.9	6.5	6.6	nd	9.0	nd	nd	nd	0.0	nd	nd	nd	15.8	nd	0.0
Ni	3.7	5.8	4.4	5.1	6.9	5.4	5.4	6.3	4.8	4.6	56.5	4.9	57.0	9.2	3.5	16.8	6.2	7.1	5.6
Zn	23.8	8.5	102.1	17.7	39.3	29.1	13.9	19.0	13.9	16.2	119.3	116.5	88.0	56.3	60.0	68.1	20.2	44.4	17.3
Cu	0.9	4.8	3.5	6.0	6.0	4.1	7.4	4.2	5.5	4.7	155.0	4.8	80.0	7.8	2.4	17.0	2.0	0.9	2.4
Co	0.6	0.5	2.1	2.0	58.8	35.1	47.3	1.0	28.1	0.9	nd	9.7	28.0	nd	nd	9.7	19.4	2.4	nd
Sc	1.11	nd	nd	nd	1.23	1.07	0.50	0.60	0.69	0.87	nd	20.85	30.00	0.00	nd	7.34	0.59	2.11	nd
Cs	0.2	nd	nd	nd	0.8	0.7	1.2	0.37	0.6	0.27	nd	0.4	0.3	nd	nd	0.3	0.2	0.5	nd
W	nd	nd	5	2	1675	751	893	nd	547	nd	nd	7	nd	nd	nd	3	384	nd	nd
Au	nd	0.001	0.002	0.011	0.001	0.001	0.002	nd	0.001	nd	nd	0.009	nd	nd	nd	0.008	0.002	nd	nd
La	23.75	17.30	109.64	24.53	11.20	23.90	16.50	35.75	16.15	15.91	nd	12.64	12.00	nd	nd	45.41	2.49	12.46	nd
Ce	45.08	20.44	210.04	31.10	41.20	41.10	29.48	67.15	25.10	22.25	nd	27.76	29.00	nd	nd	80.34	5.87	23.53	nd
Nd	10.35	4.22	102.88	9.68	10.20	13.90	11.72	18.73	10.69	8.13	nd	19.48	19.00	nd	nd	29.26	3.82	4.62	nd
Sm	1.62	0.39	19.70	1.39	1.65	1.76	1.51	2.95	1.21	1.29	nd	4.99	5.10	nd	nd	4.47	0.56	1.68	nd
Eu	0.66	0.83	4.44	0.97	0.57	0.55	0.38	0.77	0.54	1.37	nd	1.75	1.60	nd	nd	1.58	0.17	0.49	nd
Tb	0.06	0.02	2.16	0.10	0.17	0.11	0.20	0.09	0.12	0.07	nd	0.82	1.00	nd	nd	0.35	0.10	0.12	nd
Ho	0.54	0.09	1.50	0.30	0.19	0.15	0.55	nd	0.23	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Yb	nd	0.20	7.01	0.35	0.40	0.50	0.75	0.28	0.18	0.30	nd	3.24	3.40	nd	nd	1.30	0.35	0.55	nd
Lu	0.03	0.03	0.98	0.05	0.13	0.08	0.13	0.06	0.09	0.05	nd	0.44	0.48	nd	nd	0.20	0.09	0.08	nd
A/CNK	1.06	1.07	0.95	1.03	1.12	1.10	1.11	1.09	1.11	1.07	0.68	0.92	0.74	1.13	1.04	0.96	1.00	1.06	1.00
(La/Yb)n	nd	62.0	11.2	50.3	20.1	34.3	15.8	91.6	64.4	38.0	nd	2.8	2.5	nd	nd	25.1	5.1	16.3	nd
(La/Sm)n	9.5	28.6	3.6	11.4	4.4	8.8	7.1	7.8	8.6	8.0	nd	1.6	1.5	nd	nd	6.6	2.9	4.8	nd

Analytical precision of other trace and rare earth elements (analyzed by INAA) are as follows: Co, Sc, Ta, Hf, La, Ce, Sm, Eu, and Yb are better than 5% whereas Nd, Tb and Lu are 5-10%.

Analytical techniques followed for the preparation of samples for isotopic analysis were modified from Hart and Brooks (1977) and Zindler et al. (1979). Sm-Nd isotopic analysis was performed on a Finnigan MAT 261 mass spectrometer in static mode at the Geological Survey of Canada in Ottawa. All data are corrected to give LaJolla $^{143}Nd/^{144}Nd = 0.511860$, normalized to $^{146}Nd/^{144}Nd = 0.7219$. Analysis of BCR-1 gave $^{143}Nd/^{144}Nd = 0.512652$ +/- 0.000011 (2SE), in agreement with the accepted value of 0.512643 +/- 0.000019 (Carlson 1984). Replicate reproducibility is approximately 0.006% for $^{143}Nd/^{144}Nd$ and 0.4% for $^{147}Sm/^{144}Nd$ and is generally better than 0.5 epsilon units at the crystallization age. Present day values of the chondritic uniform reservoir (CHUR) used for epsilon Nd calculations are $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$ (Faure 1986). The decay constant for ^{147}Sm is 6.54×10^{-12} (Faure 1986). Sr isotopic measurements were undertaken on a National Bureau of Standards (NBS) single collector (12 inch radius) mass spectrometer at the Université de Montréal, and on a VG Sector 54 mass spectrometer at the Université du Québec à Montréal, Montréal, Québec. Analysis of the Sr standard E&A yielded $^{87}Sr/^{86}Sr = 0.70804$ +/- 0.00003 (2SE), while NBS 987 gave 0.710270 +/- 0.000018 (2SE), normalized to $^{86}Sr/^{88}Sr = 0.1194$. 1.42×10^{-11} is the decay constant used for ^{87}Rb (Faure 1986). Isotopic data is presented in Table 3.

1.4.2 Integrity of the geochemical data

Before a discussion of the geochemical characteristics of the Narsajuaq terrane can begin, the geochemical integrity of the samples must be addressed. Given the present mineral assemblages in the rocks suggesting retrogression of granulite-facies to amphibolite- and greenschist-facies conditions, it is possible that some changes in the geochemical signature of the samples could have occurred as a result of these metamorphic events. A brief overview of previous studies on the effects of metamorph-

Table 3. Nd and Sr isotopic data for the Narsajuaq terrane plutonic suites.

SAMPLE	AGE (Ma)	Nd (ppm)	Sm (ppm)	147Sm 144Nd	143Nd 144Nd	E (Nd) i	Rb (ppm)	Sr (ppm)	87Rb 86Sr	87Sr 86Sr	87Sr 86Sr i
Older Suite											
L-90a	1863	16.325	3.529	0.13066	0.511992	3.2	37.0	1398	0.08	0.70453	0.70248
S-18a	1863	21.316	4.233	0.12003	0.511900	4.0					
L-211	1863	14.921	3.052	0.12360	0.511420	-6.3					
D-211	1863	14.481	2.623	0.10945	0.511230	-6.6	90.1	478	0.55	0.71944	0.70480
D-72	1863	26.943	4.095	0.09185	0.510806	-10.7					
S-216	1863	3.227	0.474	0.08880	0.511053	-5.1					
D-41a	1848	5.680	1.122	0.11935	0.511166	-10.4					
D-29	1848	0.542	0.115	0.12858	0.511755	-1.1					
D-14a	1848	9.233	1.519	0.09940	0.511165	-5.7	76.9	405	0.55	0.71812	0.70350
L-90b	1845	22.687	5.177	0.13793	0.512109	3.6					
D-41c	1844	13.803	3.412	0.14944	0.512241	3.5					
Younger Suite											
D-20	1836	75.372	12.215	0.09795	0.511532	1.7					
S-106	1835	29.550	6.373	0.13035	0.511625	-4.1					
K-69	1835	16.638	4.868	0.17684	0.511605	-15.6					
D-28	1835	13.209	2.892	0.13232	0.512061	3.9	8.8	907	0.03	0.70272	0.70198
D-24	1834	59.358	10.586	0.10779	0.511639	1.4					
D-43	1834	110.642	17.961	0.09811	0.511535	1.7	88.9	2145	0.12	0.70569	0.70253
S-154a	1803	9.759	1.294	0.08009	0.510363	-17.5					
L-181	1803	4.554	0.367	0.04869	0.509938	-18.5	34.1	650	0.15	0.70742	0.70348
S-217	1800	13.740	1.777	0.07816	0.510947	-5.6					
D-283	1800	21.401	3.770	0.10645	0.511255	-6.2					
L-205	1800	39.331	4.816	0.07399	0.510935	-4.9	142.4	462	0.89	0.72950	0.70636
Late Suite											
S-218	1758	7.982	1.190	0.09007	0.511193	-4.2					
L-33	1758	10.365	1.374	0.08009	0.511206	-1.6					
D-259	1758	24.716	2.977	0.07278	0.511005	-3.9	94.1	777	0.35	0.71250	0.70363
L-96	1742	8.062	1.583	0.11864	0.511248	-9.7	106.5	195	1.59	0.74404	0.70432
S-44	1742	6.410	1.167	0.11002	0.511295	-6.9	110.6	272	1.18	0.73443	0.70489
D-19	1742	12.502	1.761	0.08513	0.511137	-4.4	97.2	259	1.09	0.73128	0.70402
L-68	1742	12.404	1.833	0.08931	0.511157	-4.9	99.9	250	1.16	0.73409	0.70506
L67b	1742						98.3	310	0.92	0.72777	0.70475
Cape Smith											
D-274	1898	18.143	4.817	0.16050	0.512305	2.3					
S-195	1860	12.854	2.876	0.13523	0.512031	2.9	2.0	167	0.03	0.70289	0.70197
D-42	1848	2.972	0.571	0.11603	0.511732	1.5					
D-240	1845	9.377	1.579	0.10177	0.511598	2.2	48.7	687	0.21	0.70782	0.70237
D-284	1840	30.736	4.676	0.09194	0.511532	3.2	10.4	604	0.05	0.70329	0.70197

ism on whole rock chemistry is presented followed by a discussion of the geochemical integrity of the samples used in the present study.

Previous studies of granulite-grade rocks have documented important geochemical features in these rocks, primarily low concentrations of the large-ion lithophile elements (LILE's; e.g., Weaver and Tarney 1981; Rudnick and Presper 1990; Truscott and Shaw 1990) compared to unmetamorphosed rocks of similar bulk composition. The depletions are variable, however, and in some cases enrichment of these elements has been documented (e.g., Gray 1977; Weaver and Tarney 1983; Rudnick et al. 1985). Plutons that have been intruded at depth, under granulite-grade conditions, do not follow a prograde metamorphic path and may not undergo geochemical changes that are typical of rocks formed near or at the surface of the earth and subsequently metamorphosed at granulite-facies conditions. Lucassen and Franz (1992) used the term "magmatic arc metamorphism" to describe this process of direct metamorphism of an igneous protolith with almost complete recrystallization into granulite, with reference to similar rocks from the Jurassic Andean arc in northern Chile. These authors interpreted the high-grade metamorphism to be isochemical and as such, the geochemical data for the granulite-grade rocks could be used to give information about the source of the rocks. The rocks of the Narsajuaq terrane were also intruded at granulite-facies conditions; the mineral assemblages indicative of granulite-grade metamorphism are primary igneous assemblages, and metamorphic zircons have ages similar to the timing of the waning of plutonism. Therefore, the geochemical signature of these high grade rocks likely reflects the original geochemical signature of the magmatic system from which they were derived.

Subsequent retrograde metamorphism in the Ungava Orogen, however, could have had a significant effect on the geochemical signature of the rocks, and this must be addressed. Previous studies of retrograde metamorphism emphasized the role of fluids and their effects on the bulk composition of the rocks (e.g. Beach 1980; Winchester and Max 1984). These studies documented variable changes in the major and minor element chemistry of the analyzed samples and noted a significant relationship between the metasomatic effects of the retrogression and structural conduits for the fluids. A recent study of the retrograde basement rocks in the Ungava Orogen by St-Onge and Lucas (1995) has also documented structural controls on the extent of the retrogression. The basement rocks of the Superior Province contain mineral assemblages that document conditions ranging from amphibolite- to granulite-facies conditions with a distinct zonation of metamorphic minerals that is spatially related to the Cape Smith Belt (Lucas and St-Onge 1992; St-Onge and Lucas 1995). Samples located closest to major structures, such as the basal shear zone between the Cape Smith Belt and the underlying Superior Province basement, contain the most hydrated mineral assemblages (muscovite-biotite), whereas in general those samples located further away from the basement-cover contact retain granulite-facies mineral assemblages. Geochemical analyses of samples from the same lithological unit ranging from primary granulite-facies to variably hydrated assemblages document slight metasomatic changes, predominantly in samples from the most hydrated zone (muscovite-biotite). The metasomatic effects included gains in SiO_2, Na_2O, K_2O and Ba and losses in FeO_t, CaO, MgO, TiO_2, MnO, Y, V and Cr (St-Onge and Lucas 1995). A similar study by Lucassen and Franz (1992), however, noted that while slight changes in the geochemical composition of retrograded granulite- to amphibolite-facies plutons are recorded, the overall average changes in the major and minor elements are small. They concluded that the geochemical signature of these retrogressed amphibolite-facies rocks could be considered robust.

Such detailed studies comparing granulite-grade and amphibolite-grade samples are not yet possible on the rocks of the Narsajuaq terrane due to the lack of appropriate samples of similar lithological composition but differing degrees of hydration. More detailed sampling and metamorphic and geochemical studies are necessary for such a study. However, for two plutons (monzodiorite and diorite), a comparison of the geochemistry of samples from a higher grade to a more retrogressed composition can be made. In the case of the monzodiorite pluton the comparison is between samples of upper amphibolite-facies (clinopyroxene - hornblende ± biotite) versus lower amphibolite-facies (hornblende - biotite) composition, whereas the diorite pluton contains lower amphibolite-facies (hornblende ± biotite) and greenschist-facies (epidote) mineralogies. The geochemical compositions of samples at different metamorphic conditions from the dioritic pluton show considerable overlap although K, Rb, Zr, Hf and the heavy REE's are slightly greater in one sample at higher metamorphic grade (Fig. 16a). For the monzodiorite pluton the LILE's (particularly Rb) and the light REE and middle REE are somewhat more depleted in the samples that are more retrogressed (hornblende - biotite) although as in the diorite pluton, there is significant overlap between the higher and lower grade compositions (Fig. 16b). In general the average compositions of the higher-grade and retrograde samples overlap, therefore suggesting that the retrogression did not significantly change the geochemical composition of the rocks. In addition, drawing upon the results of the study of St-Onge and Lucas (1995), it is likely that samples only from the most hydrated and retrogressed zone (muscovite-biotite) in the Narsajuaq terrane could have a metasomatic overprint similar to that documented in the basement rocks and would not be representative of the original composition of the rocks (M. St-Onge, personal communication 1995). Therefore, samples from the Narsajuaq terrane characterized by the hydrated assemblage muscovite-biotite were not selected for geochemical analysis.

Samples that were chosen for geochemical analysis are relatively fresh, with only minor alteration of plagioclase as evidenced by seritization on some grains, although in some samples sericite is quite pervasive. As was discussed in a previous section, primary igneous mineralogies are preserved (clinopyroxene - hornblende - biotite), with some retrograde mineralogies present as reaction-rim overgrowths (mainly clinopyroxene to hornblende, hornblende to epidote and biotite to chlorite) or sometimes completely replacing the original mafic mineral. Calcium carbonate is almost completely absent except as vein-fill in small rare micro-cracks in a few samples. Igneous textures are not often preserved, however, due to the strong deformation experienced by these rocks. See Appendix 1 for a complete summary of petrographic observations.

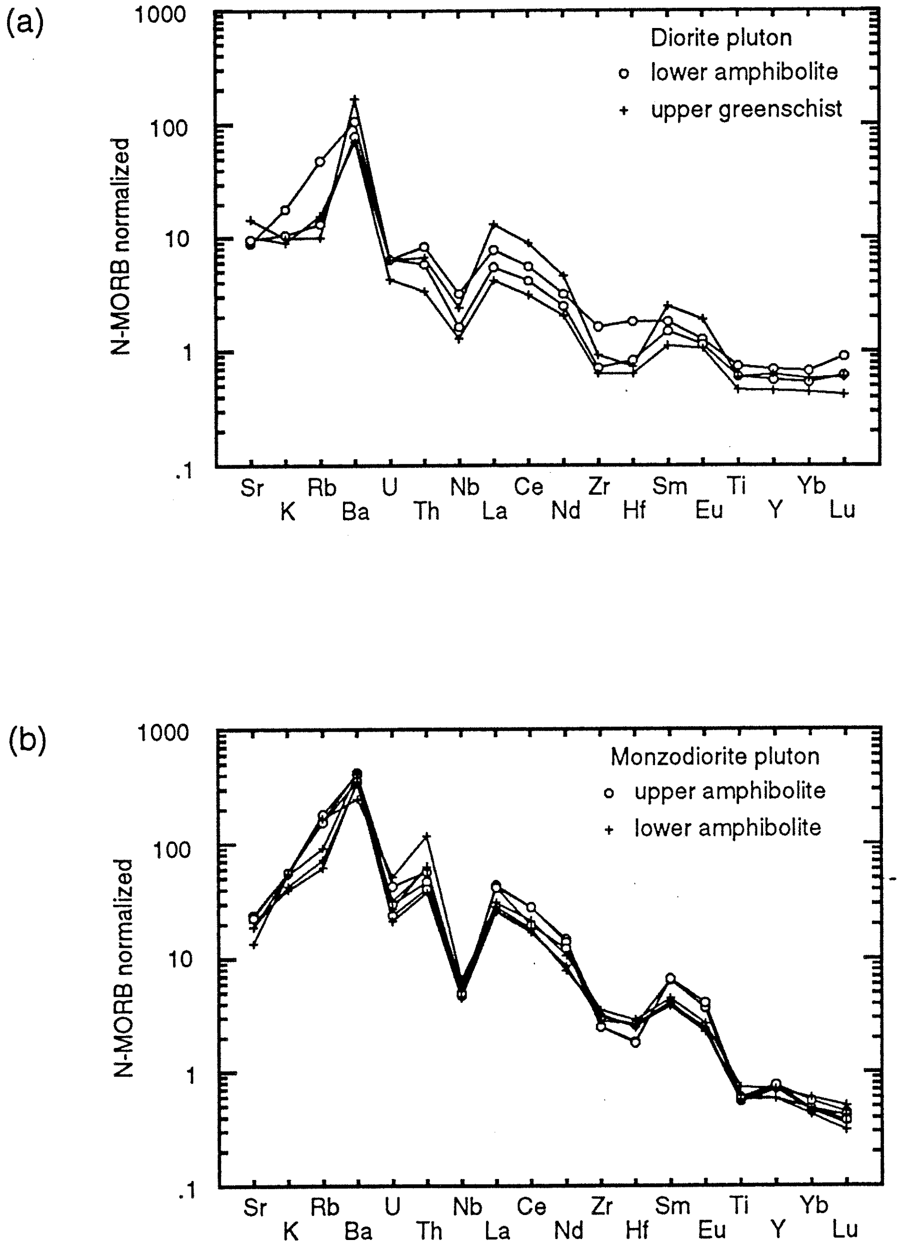


Fig. 1.16. Trace element variation diagrams (N-MORB normalized (Sun and McDonough 1989)) for samples at different metamorphic grades from two different plutons: (a) diorite and (b) monzodiorite. While some element mobility may be noted for certain elements (i.e., Rb, Sr), there is considerable overlap in the composition of the samples suggesting that retrograde metamorphism did not significantly change the geochemical signature of the rocks.

Low volatile contents were measured in the samples with LOI generally < 2.6 wt% although one sample (S195) has much higher values (>6 wt%) and was rejected. This sample also has lower than average concentrations of the LILE's compared to other samples of similar bulk composition and contains relatively high normative corundum (2.4 wt%) which is minor or generally lacking in most samples. One other analyzed sample was subsequently rejected as it contains a very low concentration of Na_2O (D64) suggesting some alteration and element mobility.

The geochemical integrity of the samples that were analyzed can be checked through an assessment of the consistency of the geochemical trends, particularly for the mobile elements, K, Na, Rb, Sr, Ba, Th and U. No consistent correlation of the volatile content with the concentration of the mobile elements was noted. Dioritic samples have a range of LOI from 0-2.6 wt% and are used for the purpose of illustration (Fig. 17). Regardless of the LOI content, approximately the same relative variation in the concentration of the mobile elements is observed (e.g., K_2O in samples with low LOI (< 1 wt%) varies from 0.3 to 2.4 wt% while samples having higher LOI have K_2O concentrations that vary from 0.4 to 2.4 wt%). Another series of diagrams plotting an immobile and then a mobile element versus a ratio of two incompatible, immobile elements document consistent variations for the different lithologies (diorite, tonalite, granite) of the Narsajuaq terrane. For example the relative variations observed in Zr or Ce versus Ce/Y are similar to the variations seen in K_2O or Sr (Fig. 18). The various lithological groups define consistent trends and maintain their relative positions within the X-Y coordinates. In addition, the coherent trends of the major and trace element data as seen in Harker and spider diagrams suggests that the geochemical signature is relatively robust as alteration and mobilization of elements would disturb these trends resulting in erratic and large variations. It is considered unlikely that geological processes could have acted on all the rocks throughout the entire field area in a similar manner, thereby shifting all the elements in the same way in order to produce the coherent geochemical trends.

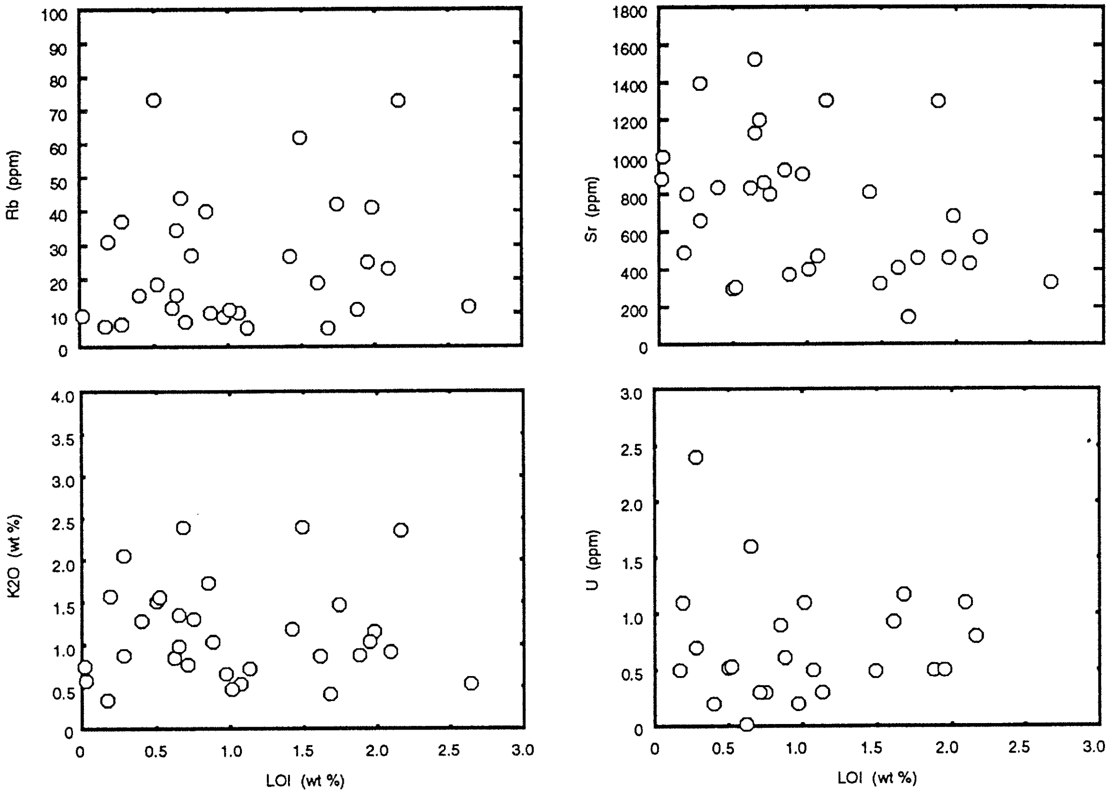


Fig. 1.17. Geochemical diagrams illustrating the lack of correlation between volatile content (LOI wt %) and mobile element concentrations (Rb, Sr, K_2O, U) for diorite compositions.

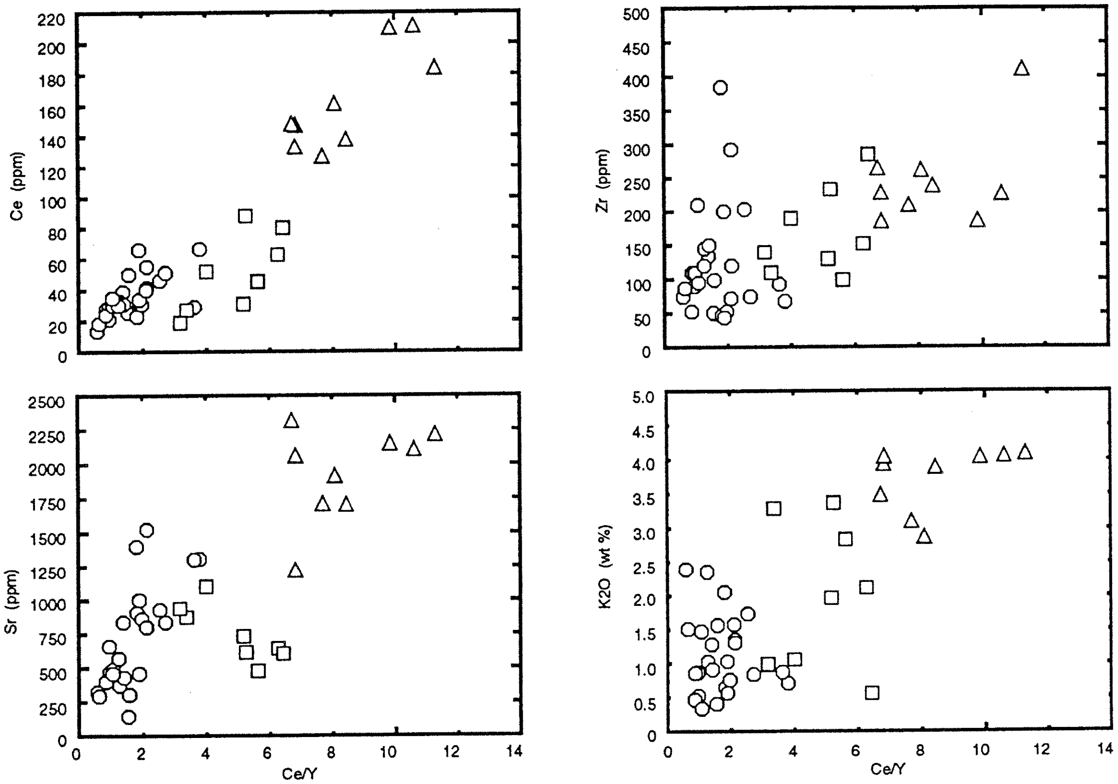


Fig. 1.18. Immobile (Ce, Zr) and mobile (Sr, K_2O) element concentrations for diorite (circles), tonalite (squares) and monzodiorite (triangles) compositions plotted against the ratio of two immobile elements (Ce/Y). The relative position and grouping of the samples for the mobile element diagrams is similar to that for the immobile elements suggesting that these samples have not been affected by processes that have significantly modified their geochemical composition.

In summary, with the exception of a few samples as noted, the geochemical signature of the samples used in this study are considered representative of the original composition of the magma from which they were derived, and only magmatic processes such as assimilation, fractionation or partial melting have produced the observed geochemical variations. The retrogression event which affected both the Narsajuaq terrane and the Superior Province basement did not significantly affect the geochemical composition of the rocks except those immediately adjacent to major thrust faults which are characterized by muscovite-biotite assemblages. Such samples are not considered in this study.

1.4.3 Major and trace element geochemistry

The plutons of the Narsajuaq terrane range in composition from diorite to syenogranite and cover a SiO_2 range from 45-81 wt%. They are predominantly calc-alkaline (CA, Fig. 19), although a few samples have more tholeiitic (TH) compositions (primarily samples from the Cape Smith suite). The plutons have K_2O contents similar to the medium- to high -K andesites of Gill (1981; Fig. 19). Systematic trends of decreasing CaO, Fe_2O_3, MgO, TiO_2 and P_2O_5 with increasing SiO_2 are noted (Fig. 20).

1.4.3.1 *Older suite*

Diorites and quartz diorites of the Older suite are Fe and Mg rich (Fig. 20; Fe_2O_3 = 7.0-15.8 wt% and MgO as high as 9 wt%) and contain moderate concentrations of Cr (11-231 ppm) and Ni (2-146 ppm) (Fig. 21). Most samples plot in the calc-alkaline field on Irving and Baragar's (1971) AFM diagram although 3 samples fall in the tholeiitic field (Fig. 19). It is not possible to systematically compare the TH and CA samples as they are spread over a wide compositional range and show much variation in their composition. The diorites have moderate to high Al contents (Fig. 20; Al_2O_3 = 14.6-20.3 wt% at 46-58 wt% SiO_2) as well as a moderate range of Na_2O/K_2O ratios (1.2-5.1) and very low Rb/Sr ratios (0.01-0.14) with the exception of two samples having higher values (L174 = 0.19 and M34 = 0.25). Variable concentrations of the large ion lithophile elements (LILE's) as well as Zr and Hf are preserved in the rocks (Fig. 21;

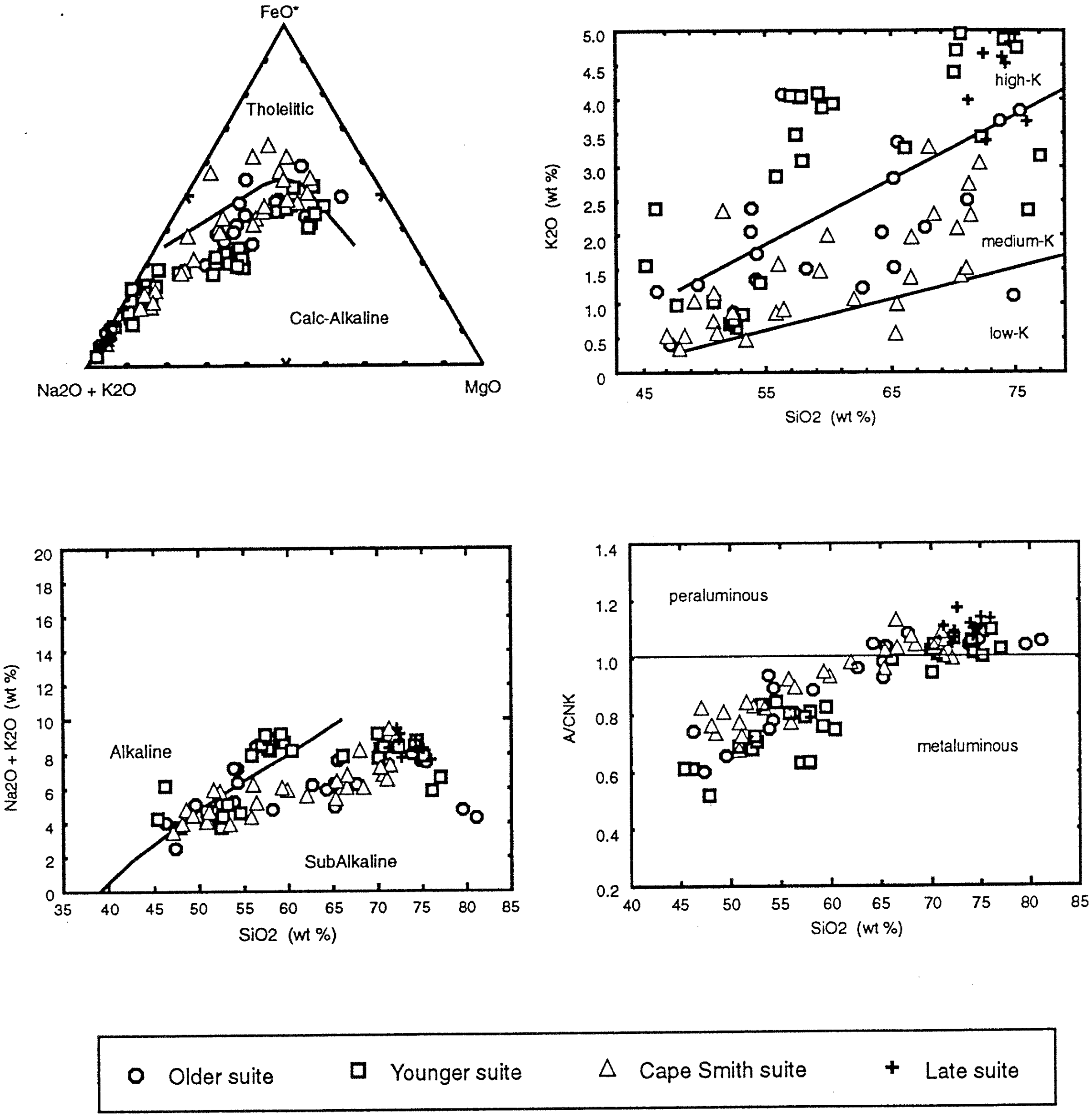


Fig. 1.19. Major element geochemical classification diagrams for the Narsajuaq terrane suites. The same legend for the samples is used for subsequent diagrams, except where noted.

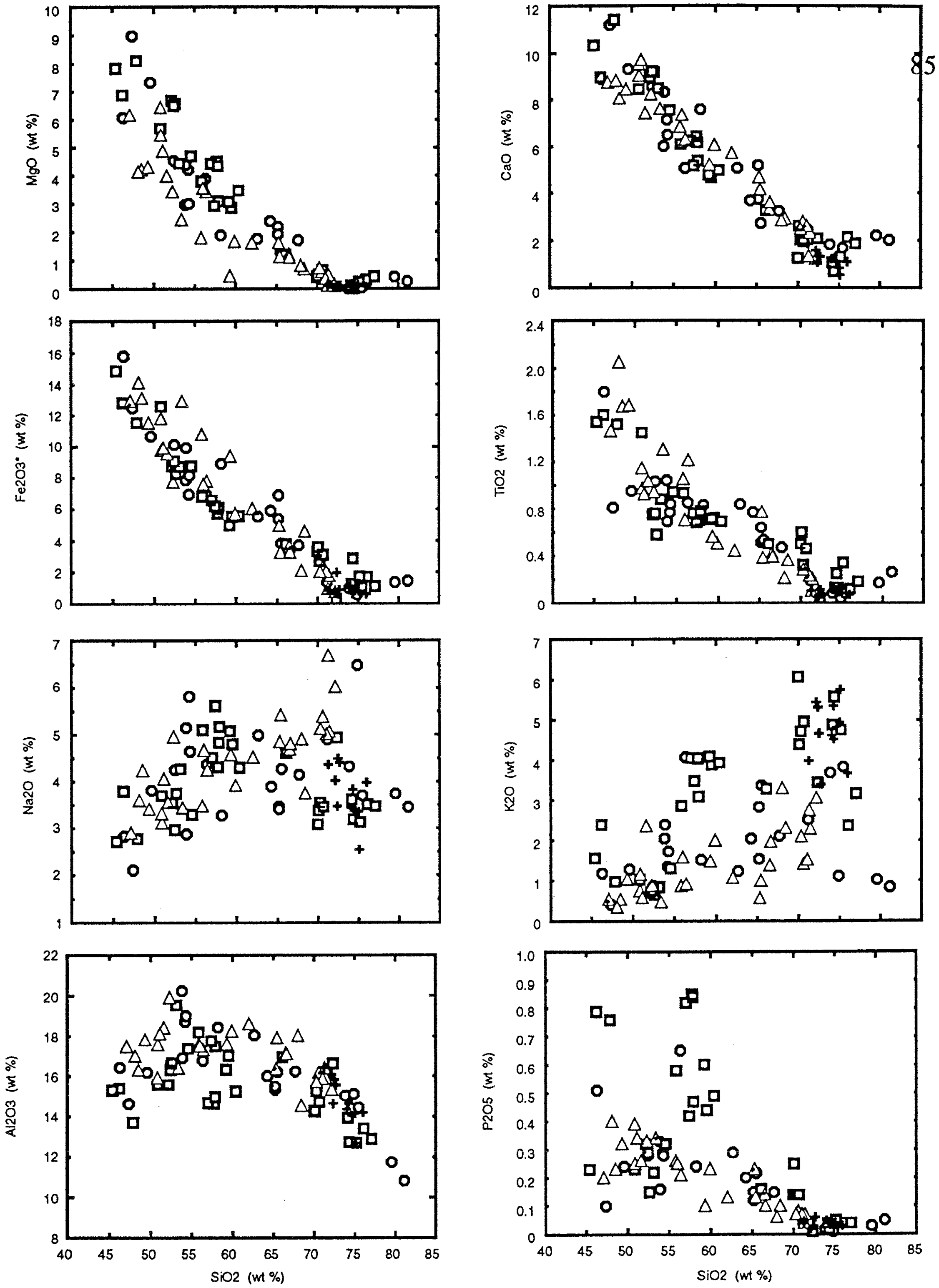


Fig. 1.20. Major element Harker variation diagrams for the plutons of the Narsajuaq terrane. Legend as in Figure 19.

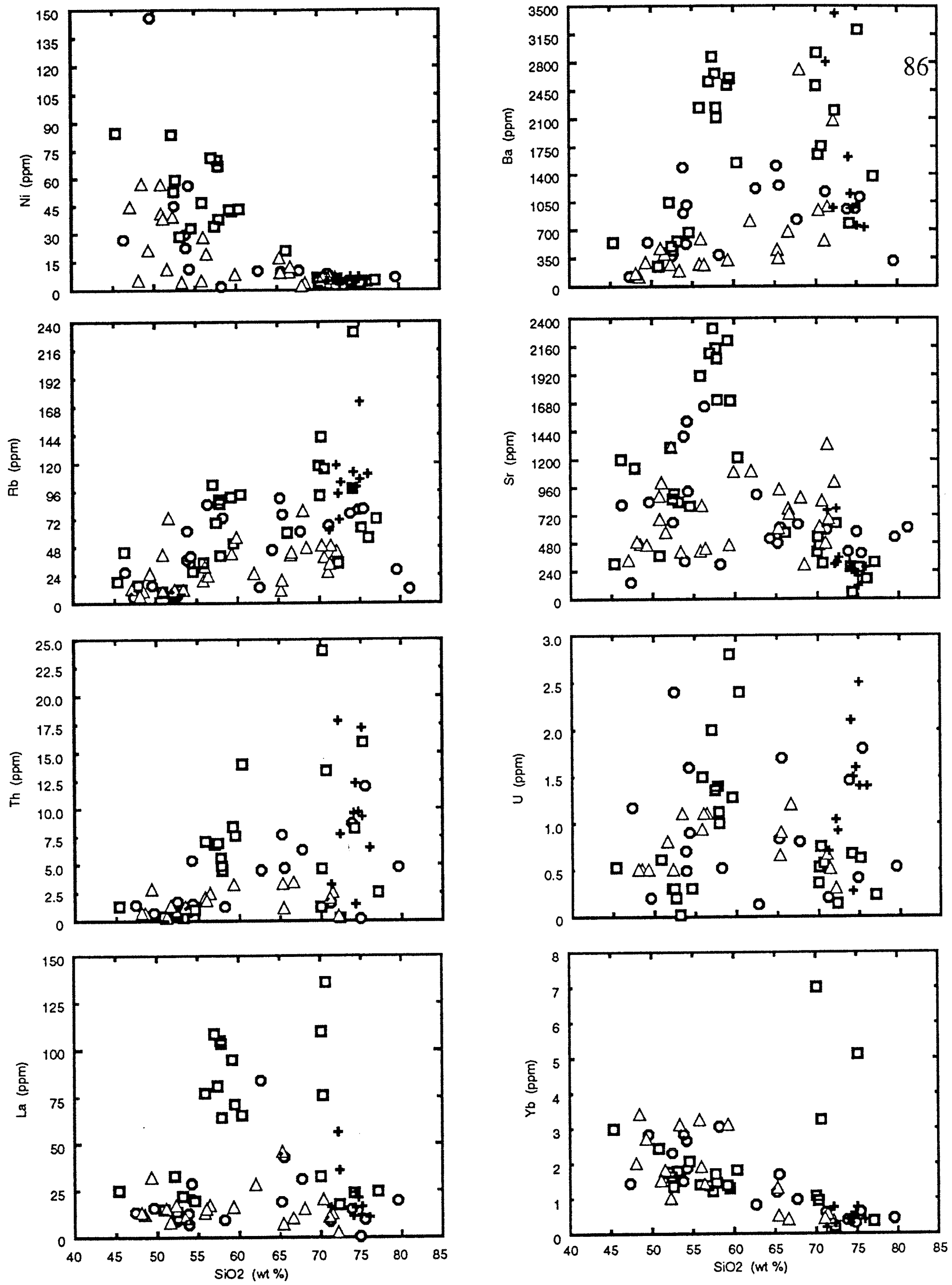


Fig. 1.21. Trace element Harker variation diagrams for the Narsajuaq terrane plutonic suites. Legend as in Figure 19.

e.g., Ba concentrations range from 124-1489 ppm, U = 0.2-2.4 ppm, Sr = 145-1524 ppm, Zr = 47-384 ppm). The light REE's (Fig. 22a) are slightly fractionated (La_n/Sm_n = 1.7-3.2, La_n/Yb_n = 1.8-7.8) with flat heavy REE patterns and negligible to slightly negative Eu anomalies (Eu/Eu* = 0.73-1.09). When plotted on a trace element variation diagram small negative Nb (La_n/Nb_n = 1.1-2.8) and Ti anomalies are evident in these rocks.

Only one sample of monzodioritic composition is present in the Older suite (M103) and as it is geochemically identical to rocks of similar composition in the Younger suite, it is included in the discussion on monzodiorites in the following section.

Interlayered with the diorites are rocks of tonalitic composition (tonalite-granodiorite) containing 62-68 wt% SiO_2 and moderate concentrations of the major elements (e.g., Al_2O_3 = 15.3-18.0 wt%, MgO = 1.3-2.4 wt%, K_2O = 1.2-3.4 wt%; Fig. 20). Ni and Cr contents are low in these rocks with Ni concentrations near 10 ppm and Cr ranging from 10-34 ppm. The tonalites/granodiorites have moderate LILE concentrations (Fig. 21; Th = 4.5-7.7 ppm, Ba = 824-1506 ppm, Sr = 478-899 ppm) with U being the most variable (0.1-1.7 ppm). Low to moderate Rb/Sr (0.02-0.12) and Na_2O/K_2O (1.2-4.1) ratios are recorded in these rocks. The REE (Fig. 22b) are variably fractionated (La_n/Yb_n = 11.6-73.3), with enriched light REE's and depleted, flat heavy REE. Eu anomalies are negligible. Prominent negative Nb (La_n/Nb_n = 1.9-10.9) and lesser Ti anomalies are preserved in these plutons.

The pervasive granitic (*sensu lato*) veins and sills that are associated with the Older suite diorites-tonalites are Si rich (SiO_2 = 71-81 wt%) and are slightly peraluminous (A/CNK = 1.04-1.09). Moderate to low Al contents (10.8-16.1 wt% Al_2O_3) and very low concentrations of MgO (<0.4 wt%), Fe_2O_3 (<1.4 wt%) and TiO_2 (<0.3 wt%) are typical of these granites (Fig. 20). Na_2O/K_2O ratios are variable, ranging from 1.0-5.8 with the higher values measured in veins of tonalitic composition reflecting the low K contents of the tonalites. These granitic veins have trace element concentrations (LILE's) that are similar to the tonalites although Nb, Y and the REE

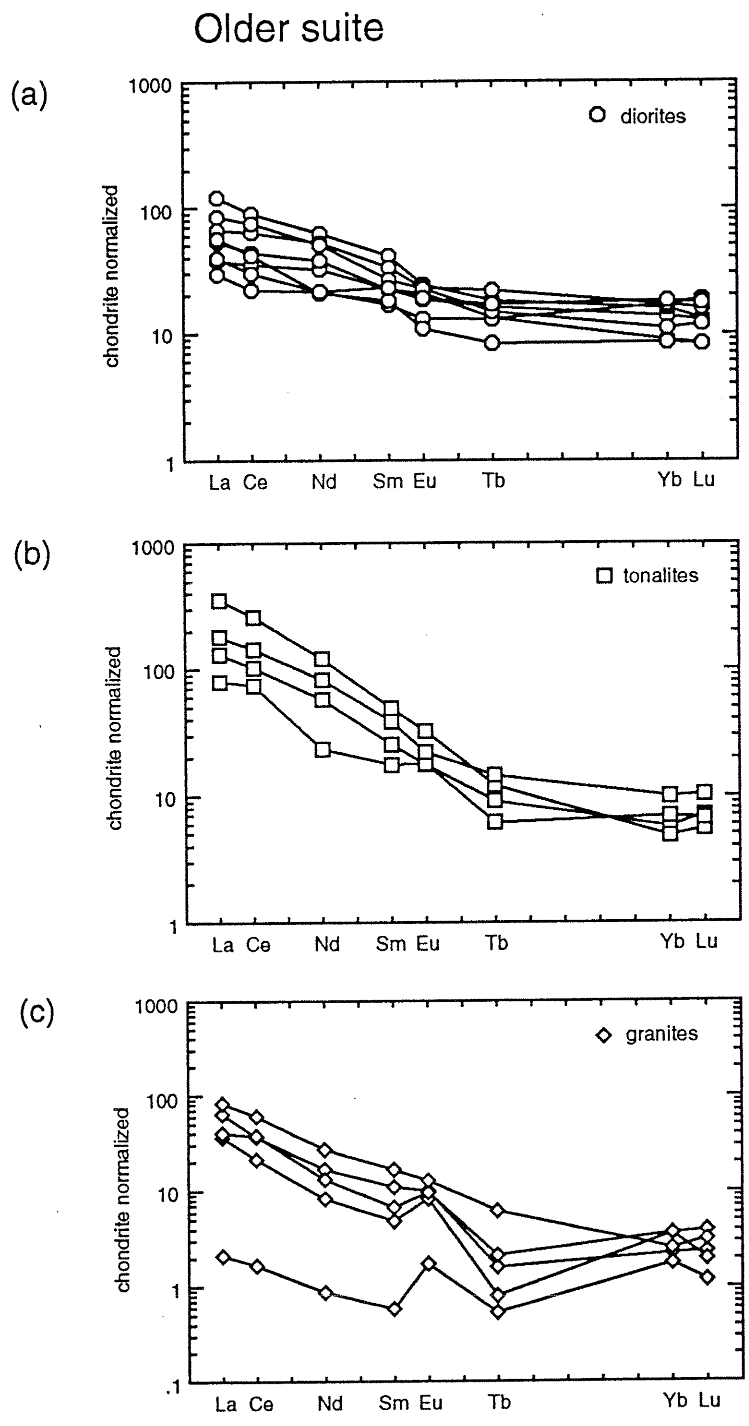


Fig. 1.22. REE diagrams (chondrite normalizing values from Sun and McDonough, 1989) for the various compositions of the Older suite: (a) diorites; (b) tonalites; (c) granites.

contents are lower in the veins (Fig. 21, 22c). Negative Ti and Nb ($La_n/Nb_n = 3.0\text{-}6.3$) anomalies are apparent on the trace element variation diagrams. One sample however, does not have such negative Nb anomalies (D-29) and is also unusual with regards to its overall REE content which is significantly lower than the other granitic veins and it has a large positive Eu anomaly. The abundances of certain other trace elements (e.g. Ba, U and Sr) in D-29 are similar to the other vein samples whereas Th and Zr are lower and Nb higher. All samples have fractionated REE patterns (Fig. 22c) with La_n/Yb_n values ranging from 10.0-32.6 (although D-29's peculiar REE pattern results in a La_n/Yb_n value of 1.2). Several of these veins have significant positive Eu anomalies and have overall concave-up REE patterns suggestive of plagioclase accumulation. Of the two non-cumulate samples, one (L27) has fractionated heavy REE's whereas the other (D14a) has a flatter heavy REE pattern. The variable enrichment of the LILE's over the REE's are reflected in high ratios of Ba/La (16-1910) and Th/Nb (up to 3.8).

1.4.3.2 *Younger suite*

Younger suite plutons also range in composition from diorite to granite (SiO_2 = 45-77 wt%) with mafic and felsic compositions most common. Although intermediate compositions (tonalite-granodiorite) are present within this suite, they were not extensively sampled for geochemical studies as they are a minor component of the suite. In general, Younger suite plutons are geochemically similar to those of the Older suite, as can be seen from the overlap of the geochemical signatures in many of the diagrams, although the former appears to be slightly more potassic (Fig. 19). The Younger suite plutons define a calc-alkaline trend on the AFM diagram with no tholeiitic compositions present (Fig. 19).

The Younger suite diorites and quartz diorites have moderate to high concentrations of MgO (4.4-8.1 wt%), Fe_2O_3 (8.3-14.9 wt%) and Al_2O_3 (13.7-19.5 wt%) at 45-55 wt% SiO_2 (Fig. 20). On the basis of TiO_2 contents two groups of samples are delineated (Fig. 20), one with moderate values of 0.6-0.9 wt% and a second group having high TiO_2 concentrations of 1.5-1.6 wt%, although such groupings are not recognized for other elements. The Younger suite diorites have similar average abundances of trace elements as the Older suite diorites (Fig. 21, 22a, 23a) although the overall range in concentrations of U, Zr and Hf are somewhat lower in the former (e.g., Zr = 47-207 ppm with one high value at 394 ppm, U <0.6 ppm) while Sr is slightly higher. The diorites have REE patterns that are slightly more fractionated than the Older suite diorites (Fig. 23a) for both the light and heavy REE's ($La_n/Yb_n = 4.4-13.5$, $La_n/Sm_n = 1.9-3.3$) and have a slightly greater average abundance of the light REE, although considerable overlap exists in the overall ranges of the REE between the two suites. The Younger suite diorites also have negligible Eu anomalies similar to the Older suite diorites.

Monzodiorites and quartz monzodiorites of the Younger suite are found in two intrusions south of Sugluk Inlet (Fig. 3), although as previously mentioned, one sample from the Older suite has been analyzed and as it possesses similar geochemical characteristics as the Younger suite monzodiorites and hence is included in this discussion. These samples are K- and P-rich (Fig. 20; K_2O = 2.8-4.1 wt%, P_2O_5 = 0.42-0.85 wt%) with a restricted SiO_2 range (55-60 wt%) and are alkaline according to the alkali classification of Irving and Baragar (1971; Fig. 19). Al contents are slightly lower (Al_2O_3 = 14.7-18.2) compared to other samples at equivalent Si-content (Fig. 20). The monzodiorites have intermediate Cr (59-138 ppm) and Ni (34-71 ppm) contents and moderate to high concentrations of LILE's such as Ba (1549-2867 ppm), Sr (1217-2313 ppm) and Th (4.5-14.0 ppm) (Fig. 21). The monzodiorites generally have trace element concentrations that are more restricted than other compositions although this may be a function of the general close proximity of the samples compared to samples of other compositions which may come from several 10's of km apart. Pronounced negative Nb anomalies (La/Nb = 4.3-9.9) are evident as is the fractionated nature of the REE patterns ($La_n/Yb_n = 25.9-56.8$; Fig. 23b). These REE patterns are very distinct due to the complete fractionation of all the REE's including the heavy REE, which contrasts with the generally unfractionated to slightly fractionated heavy REE patterns of all other samples. In addition the overall abundance of the light REE's is much greater in the

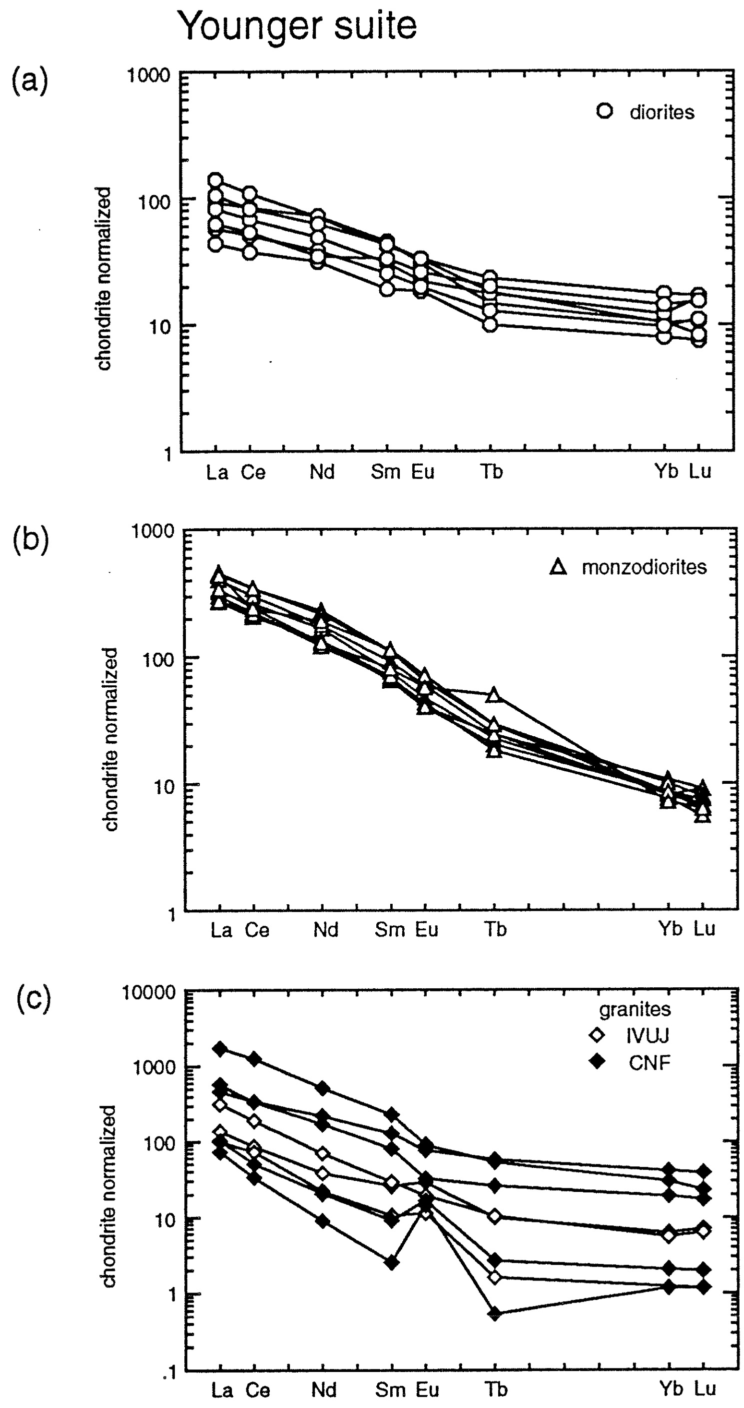


Fig. 1.23. Chondrite normalized REE diagrams for the Younger suite: (a) diorites; (b) monzodiorites; (c) granites.

monzodiorites compared to other mafic compositions, with chondrite-normalized La values ranging from 270 to 458 (Older and Younger suite diorites have La_n <137).

Only one sample (L118) of intermediate composition (tonalite) was analyzed from the Younger suite and its major and trace element composition overlaps almost completely with the Older suite tonalites with the exception of Ni which is slightly higher in L118 at 21 ppm.

Geochemical analyses of Younger suite granites are from samples of the huge granitic sheets at Ivujivik (IVUJ) and Cap de Nouvelle-France (CNF) (Fig. 3). Other granitic and granodioritic plutons from the Younger suite were identified during the course of mapping and have a different appearance and character (more deformed and foliated) compared to these granitic sheets, however no such samples are available for geochemical analysis. The IVUJ and CNF plutons are granitic to syenogranitic in composition, with high SiO_2 = 70-77 wt% and K_2O = 2.4-6.1 wt% and low Na_2O (3.1-4.9 wt%), Al_2O_3 (12.7-16.6 wt%) (Fig. 20). These granites are subalkaline and mildly peraluminous (A/CNK = 0.95-1.10; Fig. 19). LILE contents (Fig. 21) in both plutons are very similar with variable but high concentrations of Ba (773-3200 ppm), Rb (34-231 ppm) and Th (0.2-24.0 ppm), although Sr and U concentrations are rather low (53-650 and 0.14-0.75 ppm respectively). Similar to previously discussed samples, the CNF and IVUJ granites also have significant negative Nb anomalies (La_n/Nb_n = 3.2-17.0) even though anomalously high (but variable) Nb and Y concentrations were measured (Nb = 1.3-32.7, Y = 1.0-96.6). Large positive Eu anomalies in two samples of the CNF granite (S154a and L181) and a concave-up shape to the REE patterns of these samples suggests plagioclase accumulation (Fig. 23c). Other samples from this pluton have negligible Eu anomalies and fractionated light REE's (La_n/Sm_n = 3.6-7.6) with flat heavy REE patterns. A slightly more fractionated REE pattern is recorded in the IVUJ granites (Fig. 23c; La_n/Sm_n =5.3-10.9) although the overall abundance of all the REE are lower (e.g. La_n = 100-320 in the IVUJ compared to values of 463-1738 in the non-cumulate CNF samples). These abundances are slightly greater than the non-cumulate Older suite granitic veins.

1.4.3.3 *Cape Smith suite*

Plutons of a range of compositions from diorite to granite are intrusive into the lithologies of the Cape Smith Belt. Several samples from a total of 14 different plutons were analyzed in order to characterize this suite and provide a basis for comparison with plutons of the Older and Younger suites. The Cape Smith suite plutons can be divided into two groups on the basis of Fe-enrichment - tholeiitic (TH) and calc-alkaline(CA) series (Fig. 19), however, of the 28 analyzed samples only 6 have tholeiitic affinities. Similar to all the previously discussed suites, the plutons of Cape Smith suite are subalkaline and metaluminous to mildly peraluminous (A/CNK = 0.68-1.13; Fig. 19).

Diorites and quartz diorites (Si_2O = 47-59 wt%) of the Cape Smith suite have major element abundances that are similar to the Older and Younger suite diorites although slightly lower concentrations of MgO (0.5-6.4 wt%) and K_2O (0.3-2.4 wt%) and higher Fe_2O_3 (7.6-14.1) and Al_2O_3 (15.9-19.9) are found in the former (at equivalent Si_2O) (Fig. 20). Low to moderate Ni (4.0-57.0 ppm), Sr (330-1300 ppm) and Rb (6.0-73.0 ppm) and low Th (0.6-3.2) and Ba (116-581 ppm) are recorded in these diorites, with the TH samples having the lowest abundances (Fig. 21). The spectrum of trace element concentrations in the Cape Smith suite diorites are similar to the diorites from the other suites (as shown by the considerable overlap between the various suites on a trace element variation diagram) although the average abundances for many of the trace elements from the Cape Smith suite are less than the other suites. Slightly fractionated REE's characterize these diorites (La_n/Yb_n = 2.3-12.2, La_n/Sm_n = 1.1-2.9) with the TH series samples being the least fractionated (Fig. 24a). In addition, subtle fractionation of the heavy REE's are noted, contrasting to the flatter heavy REE patterns for the other diorite suites, although the overall abundances of all the REE's are similar. Only one sample (D5-2) has an important positive Eu anomaly (Eu/Eu* = 1.6) suggesting that it may be a cumulate.

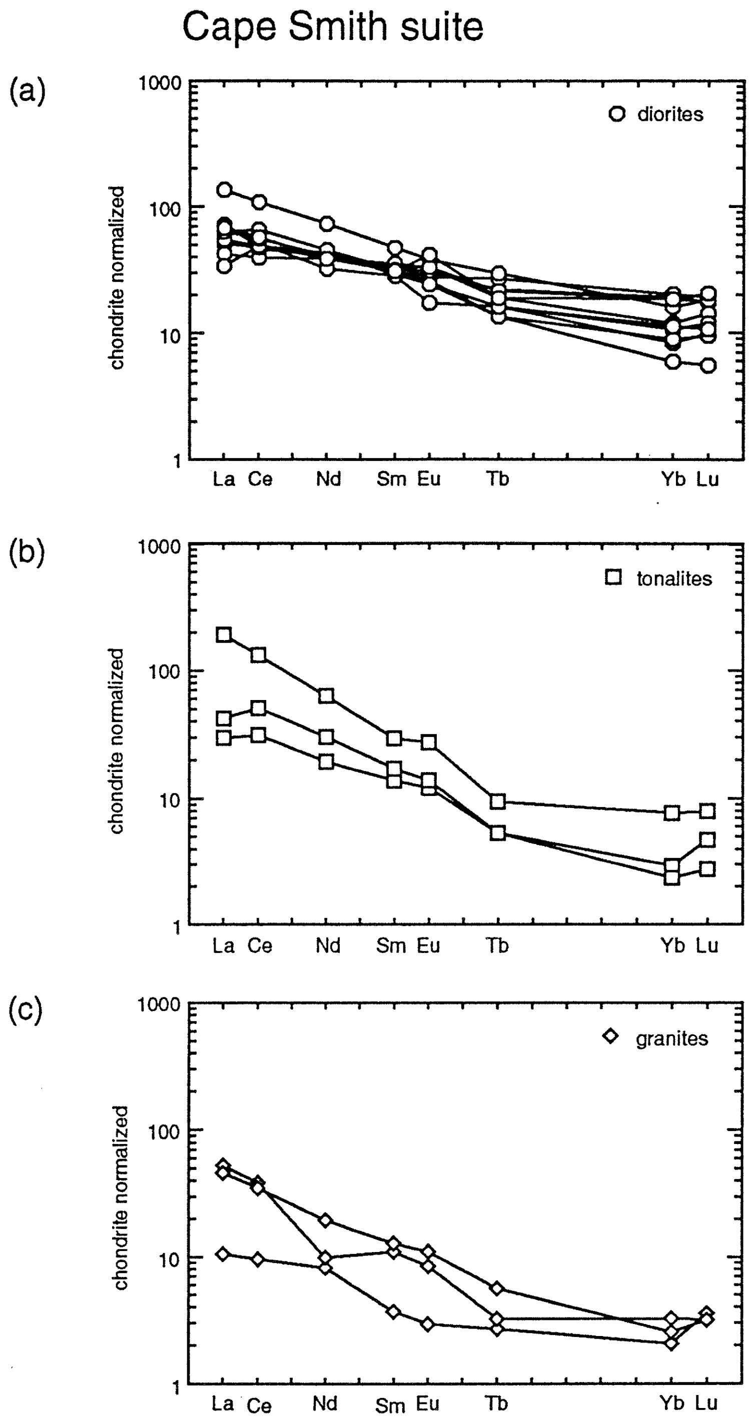


Fig. 1.24. REE diagrams for the Cape Smith suite: (a) diorites; (b) tonalites; (c) granites.

The Cape Smith suite tonalites (59-68 wt% SiO_2) have intermediate CaO, Na_2O and Al_2O_3 values but average MgO and K_2O contents that are slightly lower than previously described tonalites (Fig. 20). Low Ni (2-16.8 ppm) as well as moderate to low LILE's (e.g., Rb = 10.4-79.0 ppm, Th < 3.2 ppm, Ba = 343-806 with one high value at 2700 ppm) characterize these samples (Fig. 21). While the overall range in trace element abundances in the Cape Smith suite tonalites is similar to those measured in the Older suite tonalites, the average concentration of the trace elements in the former is slightly less. Average REE contents in the Cape Smith suite tonalites (Fig. 24b) are also slightly lower than the Older suite tonalites but they both have a similar pattern of fractionated and enriched light REE's (La_n/Yb_n = 10.0-25.1 and La_n = 30-192 for the Cape Smith suite tonalites).

The most felsic samples in the Cape Smith suite are granites and granodiorites (SiO_2 = 70-72 wt%) that have low major element contents with the exception of Na which is relatively high (Na_2O = 5.0-6.7 wt%; Fig. 20). These samples have low Cr and Ni contents as well as low Rb (26.2-49.0 ppm), Th (0.4-2.5 ppm) and U (0.3-0.7 ppm) (Fig. 21). Ba is generally low (559-986 ppm) although one sample has a higher concentration (2073 ppm), and Sr contents are moderate (476-1330 ppm). The LILE concentrations in the Cape Smith suite granites shows some overlap with those values measured for the Older and Younger suite granites but the average concentration of each element is lower in the Cape Smith suite with the exception of Sr, which is higher (Fig. 21). Fractionation of the REE's in the granites of the Cape Smith suite (Fig. 24c) results in La_n/Yb_n ratios of 5.1-18.2. The heavy REE's are flat to slightly fractionated, but it is primarily enrichment of the light REE's that give rise to the fractionated patterns (La_n = 10.5-52.5). These REE contents are similar to the Older suite granites (although D42 of the Cape Smith suite has significantly lower REE concentrations) and are less than the measured values for the Younger suite granites.

1.4.3.4 *Late suite*

The Late suite consists of granites and syenogranites (SiO_2 = 71-76 wt%) rich in

K (K_2O = 3.4-5.8 wt%) but containing very low concentrations of the other major elements (Fig. 20). The granites have very low Na_2O/K_2O ratios (<1.3) and are all mildly peraluminous (A/CNK = 1.05-1.17). High but variable LILE concentrations are measured (e.g., Ba = 725-3408 ppm, Rb = 62.4-172.3; Fig. 21) and low REE contents. The REE's (Fig. 25) are variably fractionated (La_n/Yb_n ratios = 15.8-91.6) with the syenogranites showing the greatest amount of fractionation. Only one sample (S218) has a significant Eu anomaly. The LILE contents of samples from this suite are most similar to the Younger suite granites but the REE contents of the latter range to much higher concentrations. The Late suite REE's more closely resemble the Older suite granites.

1.5. ISOTOPE GEOCHEMISTRY

A sub-set of samples was chosen for isotopic analysis in order to constrain the source(s) for these rocks which also helps constrain possible geotectonic models for the petrogenesis and evolution of the Narsajuaq terrane. Samples were chosen with the goal of attaining a regional scale coverage for the various lithologies. Correlations based on well constrained field relationships has permitted us to extrapolate the U-Pb (zircon) ages from analyzed samples (a total of 28 ages for the Narsajuaq terrane and associated plutonics in the Cape Smith Belt; Table 1) to other Narsajuaq terrane rocks throughout the region and it is these ages that have been used in the isotopic calculations. A variation of 50 Ma in the inferred age for a given sample will cause a change of 0.4 - 0.7 ε_{Nd} units which is only slightly greater than the analytical error estimated for these samples. Given the overall large range in Nd isotopic composition in the rocks of the Narsajuaq terrane (see below) this variation is considered minor. Isotope data are presented in Table 3. Prior to discussing the isotopic data for the Narsajuaq terrane, a brief discussion of the isotopic composition of the main potential magma sources (depleted mantle and continental crust) for the plutons is presented.

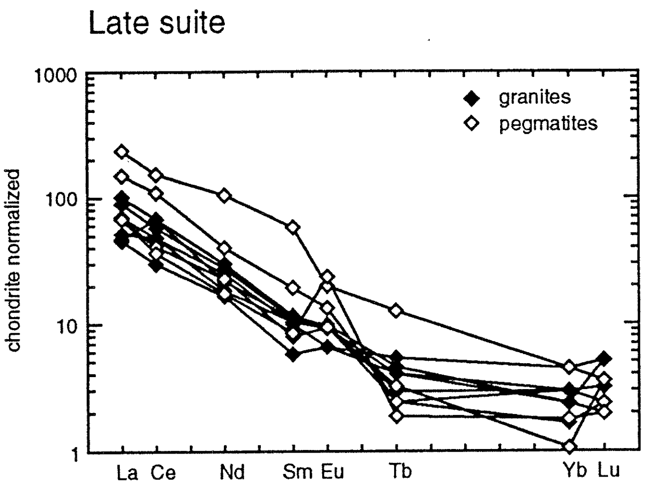


Fig. 1.25. REE diagram for the Late suite granites.

1.5.1 Depleted mantle

The rocks of the Ungava Orogen possess a wide range of isotopic signatures from juvenile or mantle dominated to highly enriched and contaminated. Several authors have presented results and interpretations of isotopic data (Sr, Nd and Pb) for the rocks of this region, predominantly for the mafic and ultramafic rocks of the Cape Smith Belt (Zindler 1982; Chauvel et al. 1987; Smith and Ludden 1989; Hegner and Bevier 1991; Gaonac'h et al. 1992). A recent synthesis of the geochemistry and isotope chemistry of the mafic rocks of the Ungava Orogen defined the compositions of various mantle reservoirs (depleted mantle, enriched mantle, arc source) present in the region at 2.0 Ga (Dunphy et al. 1995). The depleted mantle reservoir for this region of northern Quebec at this time is delimited by $\varepsilon_{Nd} \approx +4.5$ to $+5.5$ (Fig. 26; Dunphy et al. 1995). This estimate defines a mantle evolution growth curve that is slightly higher (more primitive) than that suggested by Stern et al. (1994) for the mantle of the Minto block of the Superior province located to the south of the present study area but agrees well with estimates for the contemporaneous depleted mantle of the western segment (Flin Flon) of the Trans-Hudson Orogen (Stern et al. 1995). In the following section, the depleted mantle, as well as the Nd depleted mantle extraction ages (T_{DM}), refer to this estimate which has a present day average $^{143}Nd/^{144}Nd = 0.513134$ and an assumed $^{147}Sm/^{144}Nd = 0.215$ ($\varepsilon_{Nd} = +9.7$).

1.5.2 Superior Province

Isotopic analyses of the Superior Province basement rocks in the Ungava Orogen are scant (Harvey 1995; J. Dunphy, unpublished data) but nonetheless the available data permit us to establish a range in composition for the rocks. Initial ε_{Nd} values (recalculated to 1.85 Ga) for the plutonic units of the Superior Province range from -8.1 to -18.1 with an average value at -11.7 (Fig. 26). The oldest sample from the Superior Province in the Ungava Orogen (ca. 3.1-3.6 Ga; R. Parrish, personal communication 1994) defines the lower limit of this range. The average composition of this segment of the Superior Province is similar to that calculated by Stern et al. (1994) for the Minto block of the Superior Province (located immediately south of the Cape Smith Belt; Fig.

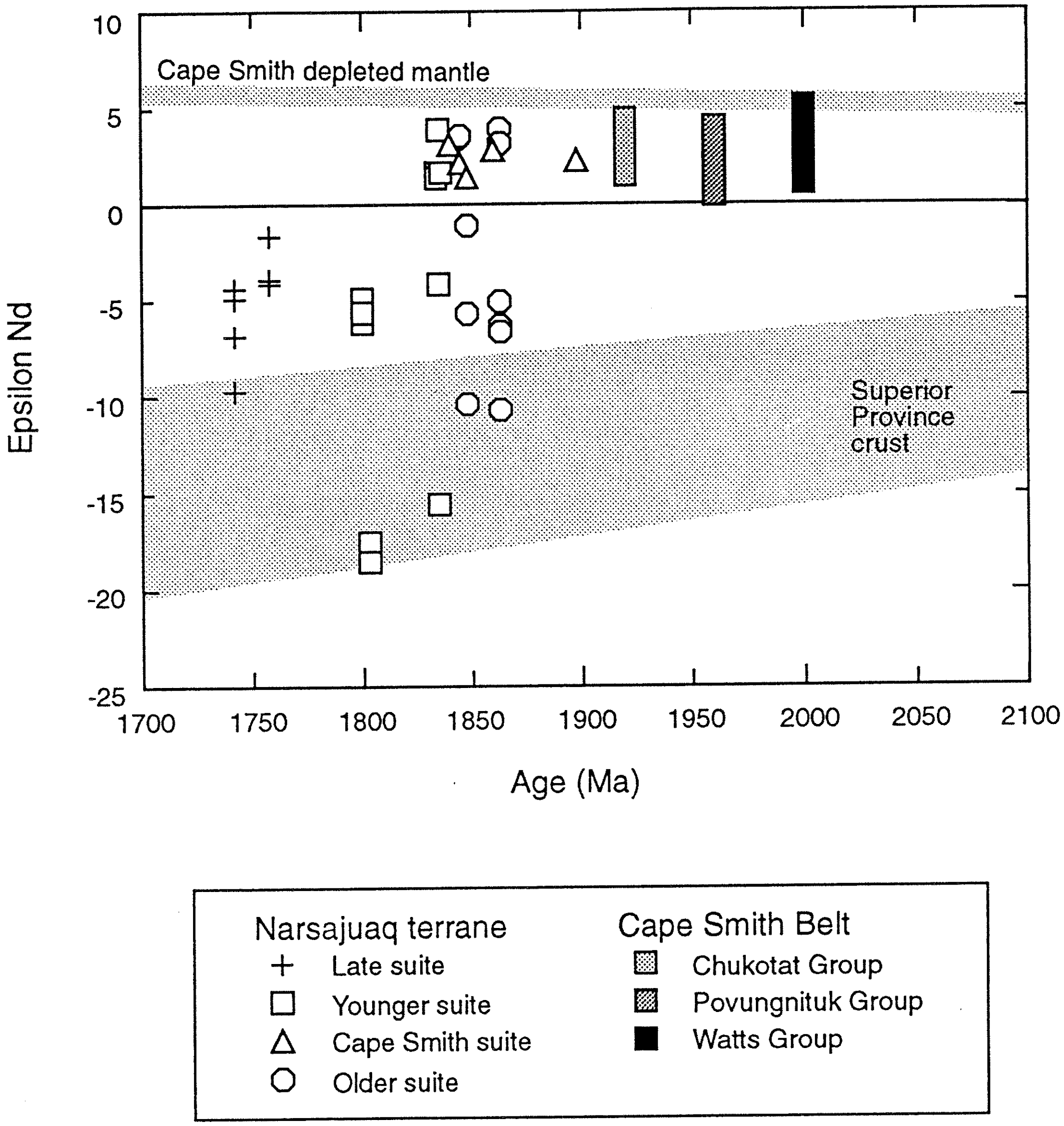


Fig. 1.26. Epsilon Nd (ε_{Nd}) versus age for the various suites of the Narsajuaq terrane. Fields representing the estimated range in composition for the Cape Smith depleted mantle (Dunphy et al. 1995) and the Superior Province basement (Harvey 1995) are presented. Also shown for reference are fields for the units of the Cape Smith Belt (Zindler 1982; Smith and Ludden 1989; Hegner and Bevier 1991; Gaonac'h et al. 1992).

2) which has an average ε_{Nd} value (1.85 Ga) of -10.1. Some very old crust was identified within the Minto block (>3.1 Ga) which has a very enriched Nd isotopic composition of -22.4 at 1.85 Ga. No Sr isotope analyses are available for the portion of the Superior Province basement in the Ungava Orogen; however, samples from the Minto block have been analyzed by Stern et al. (1994). The average initial Sr isotopic composition ($^{87}Sr/^{86}Sr$) for the 2.7 Ga old Minto block at 1.85 Ga is 0.7126, while an older 3.1 Ga sample has an initial $^{87}Sr/^{86}Sr$ at 1.85 Ga of 0.7072.

1.5.3 Narsajuaq terrane

1.5.3.1 *Older suite*

A series of samples from the Older suite (1863-1844 Ma) have a range of initial ε_{Nd} values from +4.0 to -10.7 (Fig. 26) with model Nd mantle extraction ages (T_{DM}) varying from 1.97 - 3.11 Ga. Samples of dioritic composition are the most primitive (+3.5 to +4.0) although one sample (L211) has an enriched Nd isotopic composition with ε_{Nd} (1863 Ma) = -6.3. This sample also has the highest T_{DM} age of all the diorites, at 2.84 Ga with the other samples clustering around 2.0 Ga. Several tonalite samples from the Older suite have low initial ε_{Nd} values (-5.1 to -10.7) and correspondingly old T_{DM} ages (2.50 - 2.86 Ga). One of these tonalites (S216), which was mapped as part of the Early Proterozoic suite however, has a preliminary U-Pb age that is Archean (R. Parrish, personal communication 1994). However the Nd isotopic composition of this sample at 2.9 Ga is +9.5 (!) which indicates that either the Nd isotopic analysis is incorrect or that the U-Pb age is not the age of crystallization. Given the preliminary nature of the age data, we prefer the latter interpretation and suggest that this sample is indeed part of the Early Proterozoic magmatic suite but contains inherited Archean zircon. The granitic vein samples show a large variation in Nd isotopic composition with initial ε_{Nd} = -1.1 to -10.4.

Sr isotopic analyses were undertaken on a representative sample from each lithology of the Older suite (Table 3). The diorite sample has the lowest initial $^{87}Sr/^{86}Sr$ ratio at 0.7025 (1863 Ma) while the tonalite has the highest ratio (0.7048). A sample of the granitic vein material has an intermediate initial $^{87}Sr/^{86}Sr$ value of 0.7035.

1.5.3.2 *Younger suite*

Younger suite plutons show larger variations in their Nd isotopic compositions compared to the Older suite with initial ε_{Nd} ranging from +3.9 to -18.5 (Fig. 26). Only one diorite sample (D28) has a relatively primitive Nd isotopic composition ($\varepsilon_{Nd} = +3.9$ at 1835 Ma) while two other diorite samples have initial ε_{Nd} values that are negative (-4.1 and -15.6). The monzodiorite samples cluster at values of +1.4 to +1.7 whereas the granitic plutons have much lower values (IVUJ = -4.9 to -6.2; CNF = -17.5 to -18.5). T_{DM} ages for the Younger suite rocks are identical to the Older suite with the exception of one diorite sample (K69) whose T_{DM} age is 6.01 Ga! This obviously incorrect model age is a result of the near chondritic $^{147}Sm/^{144}Nd$ ratio of the sample (0.1768).

A representative sample from each Younger suite lithology was chosen for Sr isotopic analysis. Initial $^{87}Sr/^{86}Sr$ values of 0.7020 were measured for a diorite sample and 0.7025 for a monzodiorite. The Late suite granites have higher ratios with the CNF granite sample having an initial $^{87}Sr/^{86}Sr$ ratio of 0.7035 while a sample from the IVUJ granite has a value of 0.7064 (Table 3).

1.5.3.3 *Cape Smith suite*

All of the Cape Smith suite samples that were analyzed for their Nd isotopic composition produced positive ε_{Nd} values ranging from +1.5 to +3.2 (Fig. 26). One of the samples chosen for isotopic analysis was the tonalite S195, which was excluded from the previous discussion on the geochemical composition of the Narsajuaq terrane as it was deemed to have been altered. This sample was analyzed for its isotopic composition nonetheless, as only one other sample of a similar composition was available. Previous studies on altered and metamorphosed rocks have demonstrated that such processes may not cause any significant change in the Nd isotopic composition of the rocks (e.g., Stern et al. 1995). Tonalite samples have the highest (most positive) ε_{Nd} values whereas the granites are the most isotopically enriched as they have the lowest (most negative) ε_{Nd}

values. The Nd isotopic composition of the diorite sample is intermediate between the tonalite and granite samples. The T_{DM} ages for these plutons are relatively young at 1.98 - 2.31 Ga.

Two Cape Smith suite tonalites and one granodiorite were also analyzed for their Sr isotopic composition which was found to show little variation with $^{87}Sr/^{86}Sr_i$ = 0.7020-0.7024 (Table 3).

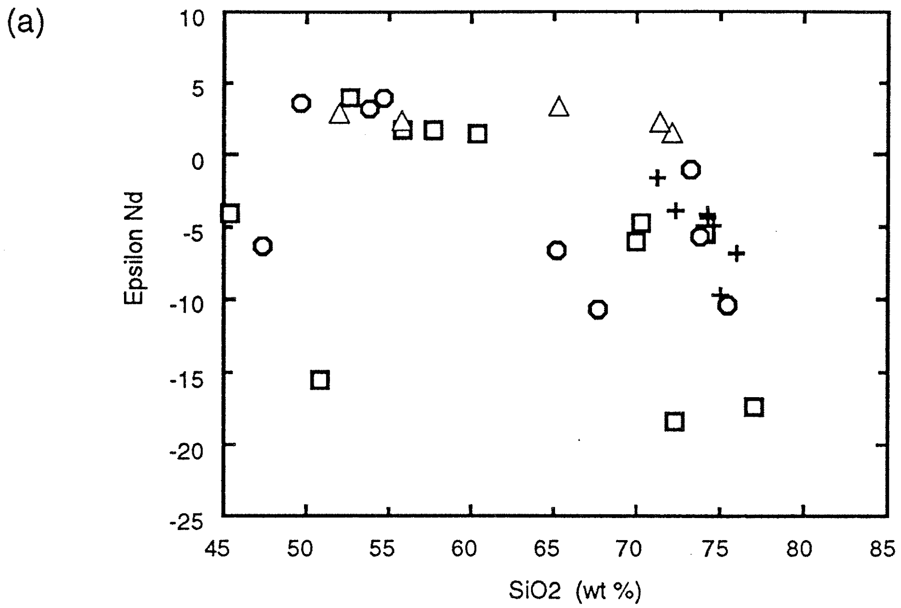
1.5.3.4 *Late suite*

The Late suite granites have a moderate spread in their isotopic compositions with initial ε_{Nd} (Fig. 26) ranging from -1.6 to -4.2 for the syenogranite pegmatites and -4.4 to -9.7 for the granitic plutons. Only one pegmatite was analyzed for its Sr isotopic composition which was measured at 0.7036. The initial $^{87}Sr/^{86}Sr$ ratio for the granitic pluton varies slightly from 0.7040-0.7051 (Table 3). T_{DM} ages for the Late suite are intermediate (2.17 - 2.96 Ga).

1.5.4 Brief discussion of isotope results

The samples of the Narsajuaq terrane plutons define a wide range of isotopic compositions ranging from near mantle values to that of continental crust. Several samples of dioritic composition have ε_{Nd} values that are only slightly lower than that of the contemporaneous depleted mantle (Fig. 26) and were likely derived from such a source with minimal subsequent contamination. Several other samples are more enriched than average continental crust (Superior Province crust) and have compositions that are closer to that of a longer-term enriched crust (represented by the lower limit of the field for the Superior Province crust on Figure 26) which is present in the area as evident from very old zircons (up to ≈ 3.1-3.6 Ga; R. Parrish, personal communication 1994).

The samples that were analyzed for their isotopic composition show a rough overall correlation of ε_{Nd} with composition (Fig. 27a), with most of the more mafic (dioritic) rocks having ε_{Nd} values above that of a chondritic reservoir (CHUR) whereas



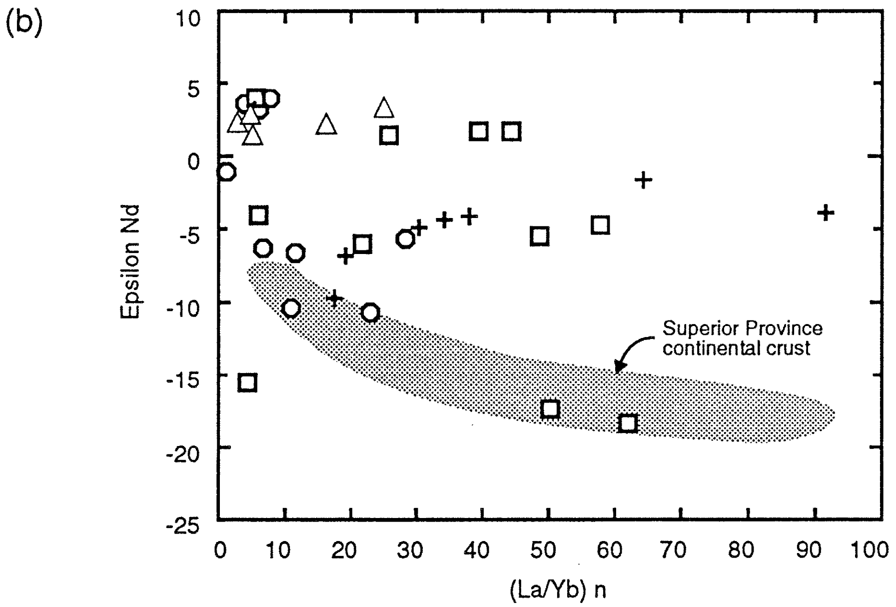


Fig. 1.27. Variation of (a) SiO_2 and (b) $(LaYb)_n$ with Epsilon Nd. The decreasing (more negative) ε_{Nd} values with increasing SiO_2 and $(LaYb)_n$ suggests that assimilation and concurrent fractionation (AFC) may have played an important role in the petrogenesis and evolution of some suites. Data for the Superior Province crust from Harvey (1995). Symbols as in Figure 26.

the more felsic (tonalitic-granitic) rocks are isotopically enriched ($\varepsilon_{Nd} < 0$). Such isotopic variations in other suites have been attributed to a process of assimilation and concurrent fractional crystallization (AFC; DePaolo 1981). Some exceptions to the trend are noted however, such as S-106, L-211 and K-69 which have ε_{Nd} values of -4.1, -6.3 and -15.5 respectively at approx. 45-50 wt% SiO_2. A general AFC trend is also reflected in the increasing $(La/Yb)_n$ ratios (reflecting greater fractionation) with decreasing (more negative) ε_{Nd} for certain samples from the various suites (Fig. 27b). With regards to geographic distribution a very rough correlation is observed of decreasing ε_{Nd} values (more negative) with increasing distance to the north. However, more isotopic analyses, particularly of the mafic compositions, are necessary to fully document this trend.

The combined results of the Nd and Sr isotopic analyses for all the groups do not show a single systematic correlation of ε_{Nd} vs. $^{87}Sr/^{86}Sr$ (t=1.85 Ga). Rather, they define a fan-shaped array intermediate between inferred mantle compositions and continental crustal compositions (Fig. 28). As was indicated in the ε_{Nd} vs. age diagram (Fig. 26), a range of older crustal compositions is required to adequately explain the isotopic composition of all the Narsajuaq terrane samples. Partial melting of such crust and/or bulk assimilation or assimilation and fractional crystallization of this crust by a mantle source could produce most of the range in compositions noted. It is likely however, that more than one mantle source composition was involved in the petrogenesis of these magmas, given the heterogenous nature of the mantle in general, and the previously documented variations in the isotopic and geochemical composition of the Ungava mantle by Dunphy et al. 1995. In addition, it is possible that crustal rocks of even older age (> 3.1 Ga) and different composition could have been present during the petrogenesis of these rocks, although no such crust has yet been identified. Therefore, the various possible mantle and crustal compositions could produce a web of assimilation and AFC lines on the diagram, accounting for all isotopic compositions. The isotope data and its significance will be further discussed in the following section.

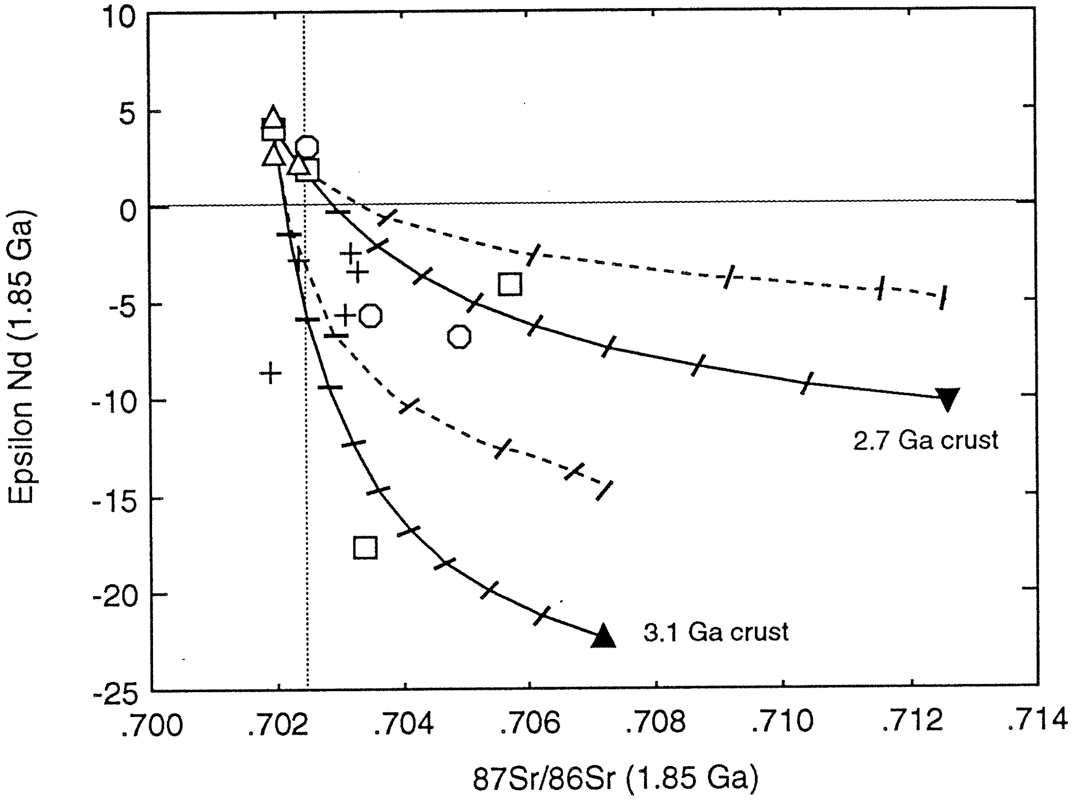


Fig. 1.28. Epsilon Nd versus $^{87}Sr/^{86}Sr$ (recalculated to 1.85 Ga) for samples from the Narsajuaq terrane (open symbols, legend in Figure 26) and Superior Province crust (2.7 and 3.1 Ga crust; Stern et al. 1994). Solid curves represent bulk assimilation of the given crustal composition by a parental magma having a composition similar to the most primitive sample while dashed curves represent AFC with the same end-members. Ticks on curves represent 10% increments of assimilation (maximum to 50% for AFC curves).

1.6. PETROGENESIS AND DISCUSSION

1.6.1 General observations and comparison to other magmatic arcs

Calc-alkaline volcanic and plutonic rocks have long been recognized as the products of arc magmatism occurring at convergent plate boundaries. In modern environments the plutonic suites are far inferior in volume to the volcanic products which are dominated by andesitic compositions (Gill 1981). However in older suites plutonic rocks become more abundant due to tectonic uplift and erosion processes which remove the overlying volcanic units and expose the plutonic roots that lie beneath. Examples of such plutonic suites can be found all along the western margin of the North and South American continents (Fig. 29; Brown 1982) where east-dipping subduction beneath the continents has been occurring since at least the Mesozoic (e.g., Brown 1982; Pitcher 1982). Some other examples can be found on some of the Indonesian Islands such Papua New Guinea (Mason and McDonald 1978), the Aleutians (Perfit et al. 1980; Malhburg Kay et al. 1990) and the Caribbean (Kesler et al. 1975). A great compositional diversity ranging from gabbro, diorite, tonalite, granodiorite and granite (and their extrusive equivalents of basalt, basaltic andesite, andesite, dacite and rhyolite) is found in these suites. In many cases the plutonic bodies coalesce to form huge batholiths consisting of hundreds or even thousands of plutons spanning many tens of millions of years.

The Cretaceous Peninsular Ranges Batholith (PRB; Fig. 29) is a well studied example, consisting of hundreds of plutons, 1-50 km in diameter, with a total length of over 1000 km (Silver and Chappell 1988). The PRB was emplaced over a period of approximately 60 Ma representing two main periods of magmatic activity, with rocks of 140-105 Ma in the western part of the batholith and the eastern PRB showing a systematic younging of ages from west to east of 105 to 80 Ma (Silver and Chappell 1988). The authors interpreted the distribution of the ages to indicate the presence of an older static magmatic arc followed by a younger east-migrating arc. The batholith intruded apparently coeval metavolcanic and metasedimentary rocks in the west and a thicker package of metasedimentary rocks (potentially older basement) in its eastern

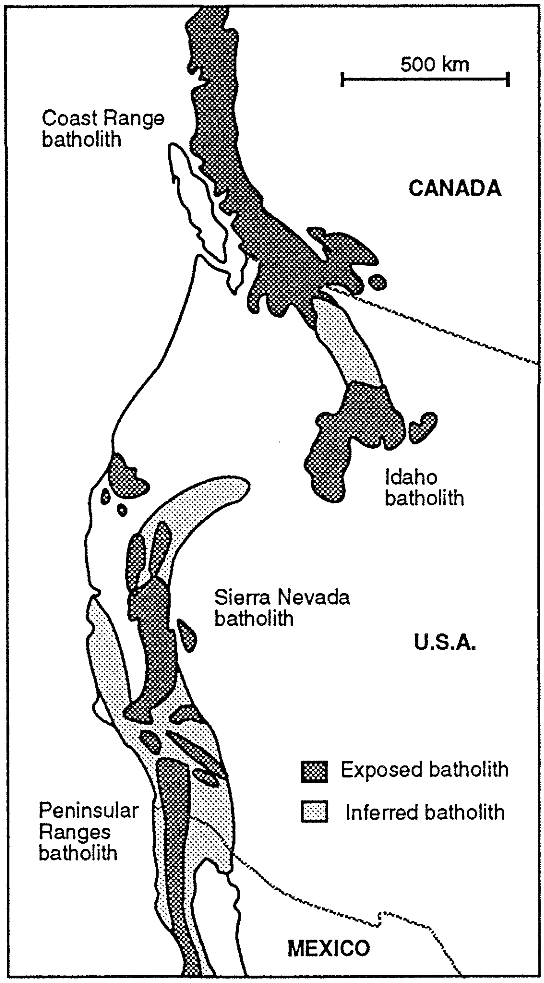


Fig. 1.29. Major batholiths of western North America related to eastward subduction under the continent since the Mesozoic. Modified from Brown (1982).

segment (Silver and Chappell 1988). Although a range of plutonic rock compositions are present in the PRB, tonalites and low-K granodiorites are the most common (Larsen 1948, cited in Silver and Chappell 1988). On the basis of geochemical data, primarily REE abundances, Gromet and Silver (1987) divided the PRB into three zones (western, central and eastern). Other parameters also show marked asymmetry across the width of the batholith, such as the previously mentioned age data, the initial $^{87}Sr/^{86}Sr$ ratios and $\delta^{18}O$ (increasing from west to east; Silver and Chappell 1988) and Nd isotopic composition (ε_{Nd} values decreasing eastward; DePaolo 1980, 1981). Gromet and Silver (1987) interpreted the geochemical differences between the different zones as related to different source compositions rather than subsequent contamination or fractionation processes. The Coast Range Batholith (Roddick 1983; Barker and Arth 1984; Barker et al. 1986) and the Sierra Nevada Batholith (Bateman and Chappell 1979; Bateman 1983) are two other well known examples of Cretaceous arc magmatism found within North America (Fig. 29). The various plutons and batholiths along the western margin of South America have also received wide attention, and there have been a number of special volumes and synthesis papers dealing with the particulars of the geology and geochemistry of these rocks as well as those elsewhere along the circum-Pacific region (e.g., Circum-Pacific Plutonic Terranes (Roddick 1983); The Nature and Origin of Cordilleran Magmatism (Anderson 1990); Plutonism from Antarctica to Alaska (Mahlburg Kay and Rapela 1990); Andean Magmatism and Its Tectonic Setting (Harmon and Rapela 1991)). Many similarities have been reported for these other plutons/batholiths as has been observed within the PRB.

As was described in a previous section, the plutonic rocks of the Narsajuaq terrane span a range of compositions from dioritic to granitic with intermediate compositions most abundant (more than 50% of the exposed surface consists of the broadly tonalitic Older suite with only approximately 15% mafic compositions present in the entire terrane), similar to the PRB. The compositional polarity noted in the PRB of increasing SiO_2 content from west to east (reflected in the greater abundance of more felsic alkali-rich granitoids and lesser amounts of mafic plutons) is also observed within the Narsajuaq terrane. With increasing distance north from the southern margin of the terrane, the overall bulk composition of the Narsajuaq terrane becomes more felsic and mafic diorite plutons are almost non-existent. Although the overall size of the plutonic units in the terrane is much smaller than the massive Mesozoic batholiths of California, the fact that the southern boundary of the terrane is a major fault and the northern, eastern and western boundaries are undefined due to the limit of the land mass, suggest that the actual size of the terrane is much larger than presently exposed. The range in ages for the main phase of the Narsajuaq terrane plutonism from 1898 to 1821 Ma represents a slightly longer period of magmatism (77 Ma) than has been documented for the PRB (60 Ma).

For comparative purposes a geochemical database containing major and trace element analyses of a variety of compositions (gabbro-diorite to granite) from various magmatic arc plutons of the western North and South American continents and parts of Antarctica was compiled. The names and locations of the plutons and a complete reference list can be found in Table 4. The Narsajuaq terrane plutons are predominantly calc-alkaline but some tholeiitic compositions have been noted, mainly confined to the Cape Smith suite, although no clear tholeiitic trend can be defined. Significant overlap occurs between the Narsajuaq terrane suites and the granitoids of the PRB, Sierras, Chile, etc. from the reference suites (Fig. 30a). Peacock's (1931) alkali-lime index (SiO_2 content in wt% when $CaO=Na_2O+K_2O$) for the Narsajuaq terrane suites (Fig. 30b) lies within the range defined for alkali-calcic to calc-alkaline suites at ≈ 54-65, similar to the plutonic suites of the reference data set. Brown (1982) also defined trends of increasing arc maturity based on the Peacock index, with the least mature arcs having SiO_2 values of approximately 68, to the most mature with value of ≈56. The relatively wide range in K_2O contents measured in the Narsajuaq terrane plutons is similar to that observed in the Phanerozoic plutons of the reference suites, with all rocks falling into the field defined by Gill (1981) for medium- to high-K andesites (Fig. 30c). In general, the major element composition of the plutonic suites of Narsajuaq terrane is quite similar to those from North and South America, although the alkaline monzodiorite rocks are unique to

Table 4. Data sources and locations for the Phanerozoic geochemical database.

Location	**Reference**
Alaska-British Columbia (Coast Range)	Barker et al. (1986)
California (Penninsular Range)	Gromet and Silver (1987)
	Silver and Chappell (1988)
California (Sierra Nevada)	Bateman and Chappell (1979)
Peru	Le Bel et al. (1985)
Chile	Lopez-Escobar et al. (1979)
	Baldwin and Pearce (1982)
	Rogers and Hawkesworth (1989)
	Pitchowiak et al. (1990)
Antarctica	Saunders et al. (1980)
	Harrison and Piercy (1990)

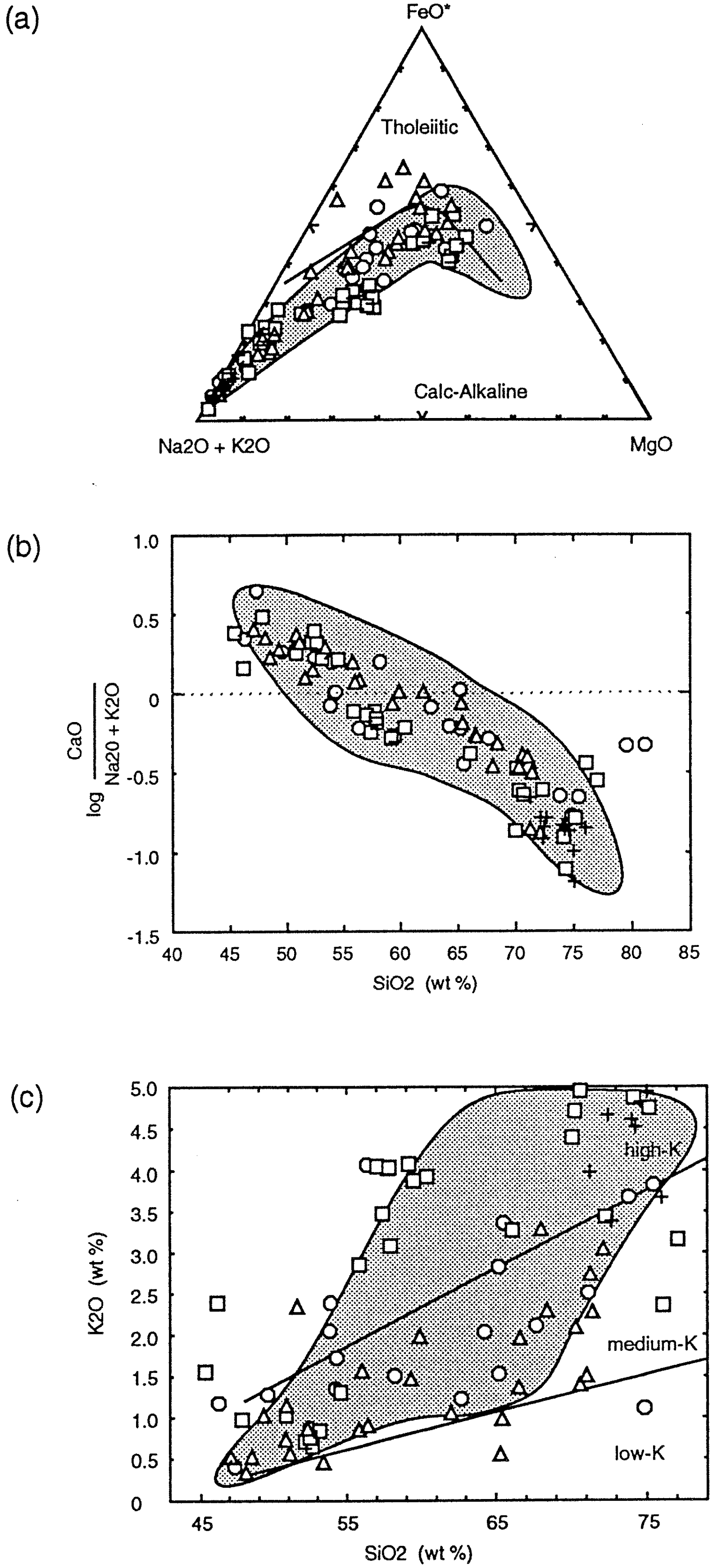


Fig. 1.30. Major element classification diagrams for the Narsajuaq terrane (open symbols, legend in Figure 19) and the reference suite (shaded field; references in table 3): (a) AFM diagram (Irving and Baragar 1971); (b) Peacock diagram (Peacock 1931); (c) K_2O vs. SiO_2 diagram (Gill 1981).

the Narsajuaq terrane (Younger suite). Na_2O is slightly higher on average in the Narsajuaq terrane rocks, which also have higher and more variable Na_2O/K_2O ratios compared to the reference data set.

There are also some slight differences in the trace element composition of the Narsajuaq terrane plutons and the Phanerozoic plutons of the magmatic arcs from the reference suite. Rb contents have an overall larger range and Sr concentrations are lower in the reference magmatic arc suites (Fig. 31) which results in a more variable Rb/Sr ratio for these suites (the majority of samples range up to 1.0) compared to the Narsajuaq terrane which generally has values < 0.5. K/Rb contents of the Narsajuaq plutons are higher (starting at ≈ 200) and more variable than the plutons of the reference data set which range from ≈150 to maximum of 600 (Fig. 31). The most extreme differences are in the radiogenic elements of Th and U which are much higher in the Mesozoic magmatic arcs than in the Narsajuaq terrane plutons (Fig. 31). Th contents up to 45 ppm are observed in the plutons of the reference suites whereas 25 ppm is the maximum for the Narsajuaq terrane, and typical U concentrations at 65 wt% SiO_2 are approximately 10 ppm for the North and South America plutons (but range to 13 ppm) compared to <2 ppm for the Narsajuaq granitoids. The behaviour of Y and Yb are also noticeably different in the two groups, with Y and Yb contents remaining relatively constant in the reference suite over a wide range of SiO_2 contents but showing marked decrease in the Narsajuaq terrane suites (Fig. 31). The variation in REE content of tonalite compositions in the PRB has been thoroughly investigated by Gromet and Silver (1987) who documented a change from relatively flat REE patterns for samples in the western segment to much more fractionated and light REE enriched patterns for the eastern samples (Fig. 32a). Many of the samples in the reference data set from South America have REE patterns and overall abundances that are similar to the western PRB samples whereas light REE enriched samples from the Coast Range batholith of British Columbia and Alaska more closely resemble the eastern PRB (Fig. 32b). Tonalite compositions from the Narsajuaq terrane closely resemble the eastern PRB and Coast Range samples both in terms of REE abundances and the shape of the REE patterns

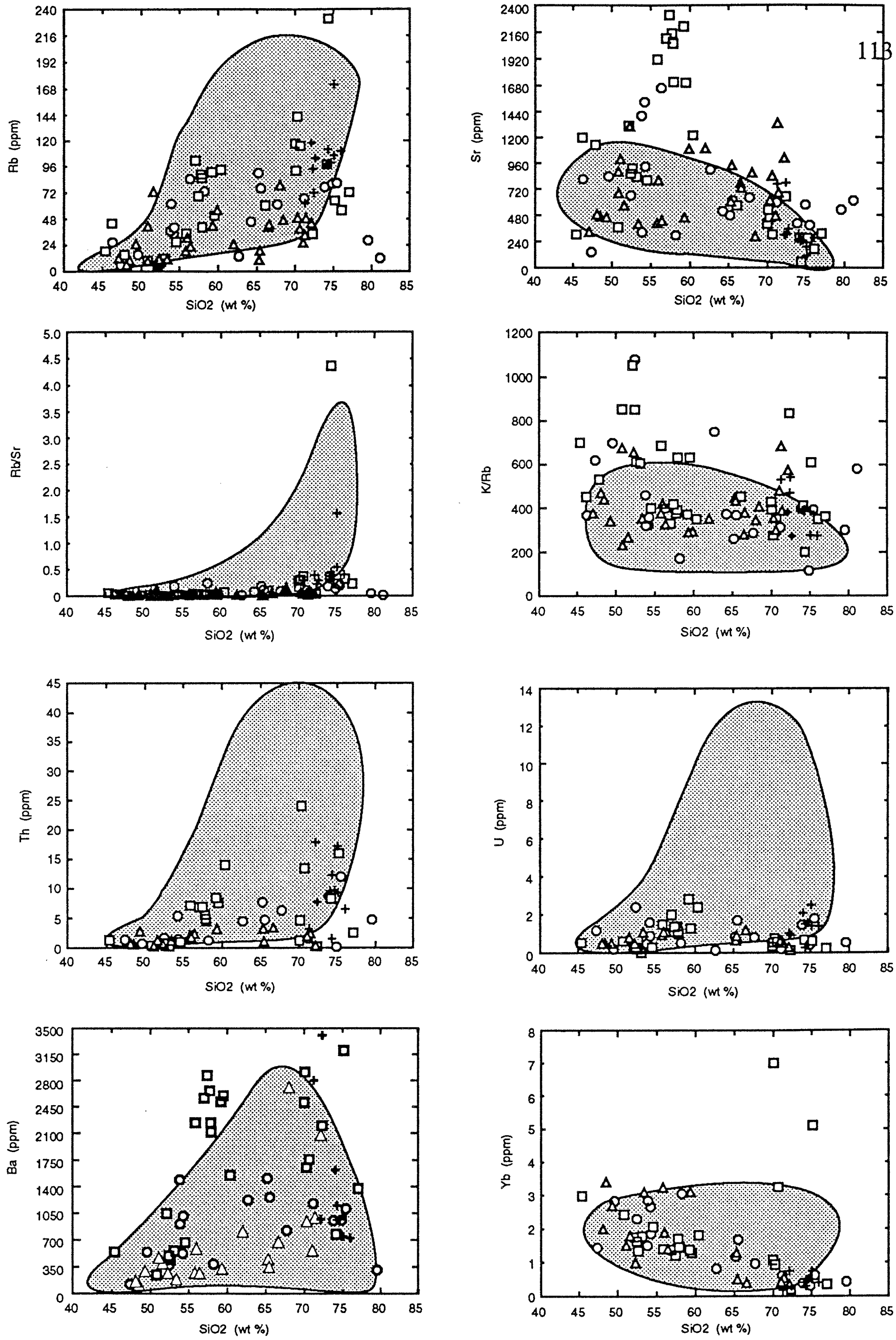


Fig. 1.31. Trace element Harker variation diagrams for the Narsajuaq terrane (open symbols, legend in Figure 19) and the reference suite (shaded field; references in Table 4).

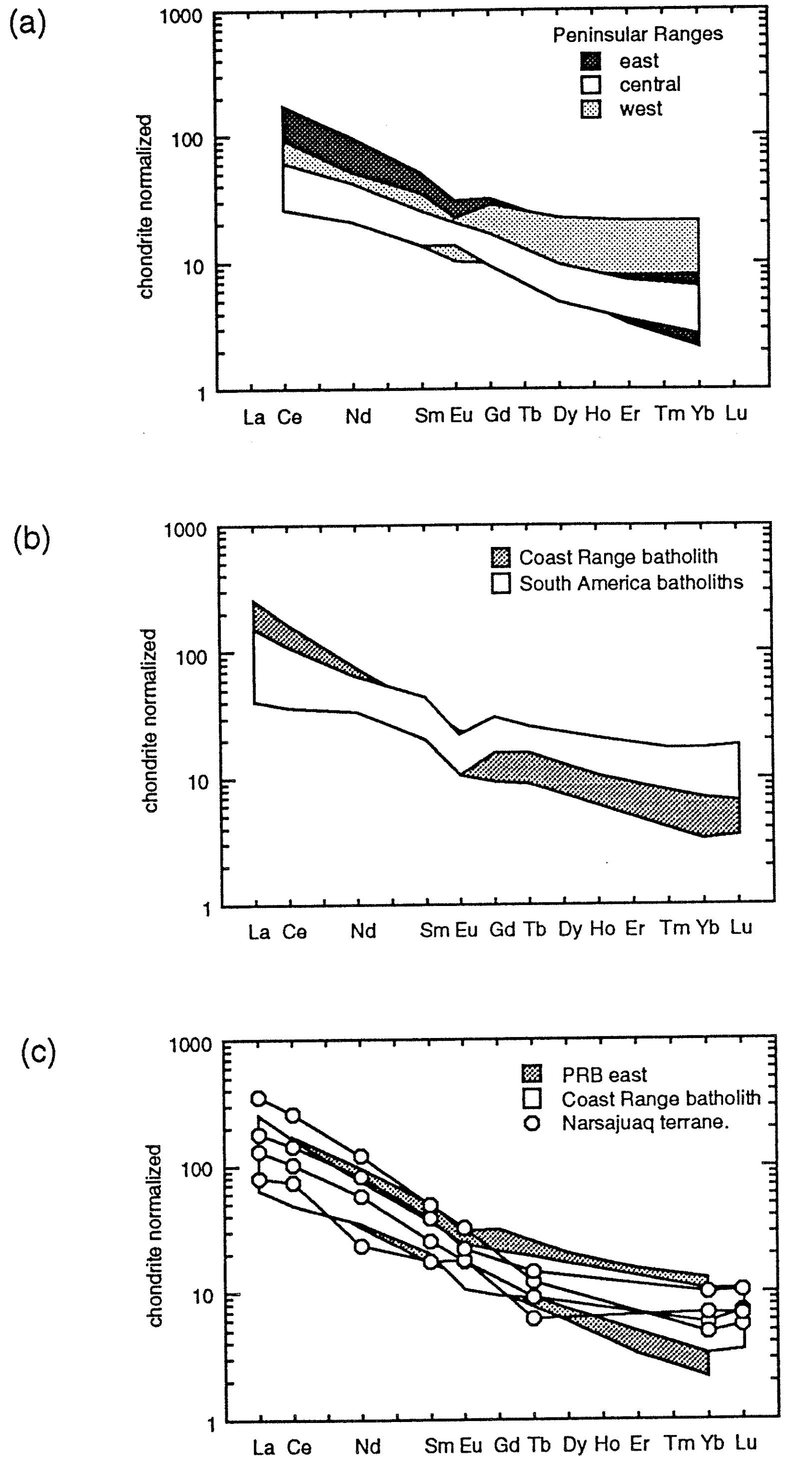


Fig. 1.32. Chondrite normalized (Sun and McDonough 1989) REE diagrams for intermediate compositions from (a) the Peninsular Ranges batholith; (b) the Coast Range and various South American batholiths; and (c) the Narsajuaq terrane. References for the Phanerozoic data in Table 4.

(Fig. 32c). A detailed evaluation of the spatial distribution of the various geochemical parameters for the Narsajuaq terrane is not possible at present given the regional-scale sampling. This present study aims to provide an overall regional-scale characterization of this previously unknown terrane, after which more detailed studies similar to those by Gromet and Silver (1987) may be undertaken.

The general geotectonic classification schemes such as those established by Pearce et al. (1984) for granitic rocks also illustrates the similarity of the geochemical composition of the Narsajuaq terrane plutons to modern volcanic arc granites (Fig. 33a and b). Pearce (1983) also created similar diagrams for differentiating volcanic arc basalts formed in an oceanic arc environment versus a continental margin. The mafic plutons of the Narsajuaq terrane plot mainly within the continental arc field although some samples do straddle the boundary and plot within the oceanic arc field (Fig. 33c and d; only samples of dioritic composition are plotted).

In summary, many similarities exist between the plutons of the Narsajuaq terrane and Phanerozoic plutonic rocks that were formed in magmatic arc settings. The range in compositions of the plutons, their individual sizes and the geochemical characteristics (except for Th and U) of the Narsajuaq terrane rocks are comparable to the various plutons and batholiths found along the western margin of North and South America. It is therefore suggested that the Narsajuaq terrane plutons may have had a similar origin as these granitoids and hence are interpreted as the products of an Early Proterozoic magmatic arc.

1.6.2 Petrogenesis of arc magmas

In the previous section we documented the similarities and differences between the plutons of the Narsajuaq terrane and those formed in magmatic arcs. The petrogenesis of volcanic and plutonic rocks in such an environment can involve several different sources and many magmatic processes. Sources as diverse as the subducted oceanic lithosphere and associated sediments, the mantle wedge lying between the sub-

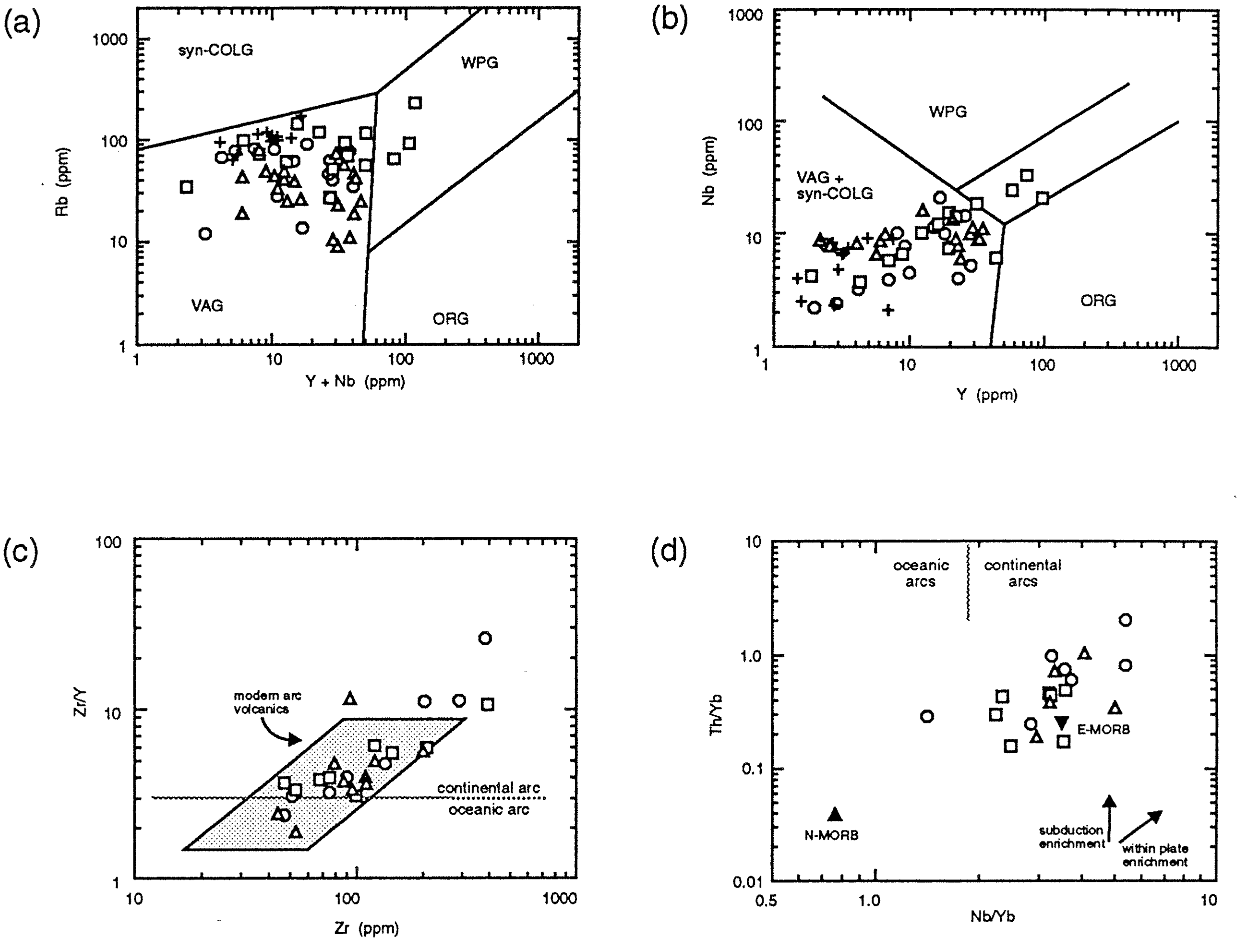


Fig. 1.33. Geochemical tectonic classification diagrams for Narsajuaq terrane suites illustrating the arc affinity of the plutons: (a) and (b) granitic rocks with > 5% modal quartz (from Pearce et al. 1984; VAG = volcanic arc granites, syn-COLG = syn-collisional granites, WPG = within plate granites, ORG = ocean ridge granites); (c) and (d) mafic Narsajuaq terrane rocks (from Pearce 1983). Symbols as in Figure 19. Shaded field for modern arc volcanics in (c) from Pearce (1983); N-MORB and E-MORB values from Sun and McDonough (1989).

ducting slab and the overlying plate, and continental crust can all contribute to the generation of arc magmas. Superimposed on the geochemical variations of the sources are the variations induced by the processes of magma generation such as partial melting, fractional crystallization and combined assimilation and fractionation (AFC-type processes).

Primary arc magmas can be generated by partial melting of the subducted slab (altered oceanic crust +/- sediments) and/or the mantle wedge (lithospheric mantle). Partial melting of the slab during subduction appears to have been more common in the Archean due to higher geothermal gradients (Martin 1986) although modern examples have been documented in areas where subduction of young hot crust is occurring (Drummond and Defant 1990). Melting of the subducting slab under eclogite to garnet amphibolite conditions will produce high-Al tonalite-trondhjemite-granite (TTG) magmas having high $(La/Yb)_n$ and Sr/Y ratios and low concentrations of the heavy REE, Y, Nb, K/Rb and Rb/Sr (Drummond and Defant 1990) as the garnet (and hornblende to a lesser extent) will retain these elements during partial melting due to their relatively high Kd's. Partial melting of the basaltic slab can occur at relatively low temperatures and pressures under water-saturated conditions, but only under anhydrous conditions can primary andesite magmas be produced requiring much higher temperatures (Fig. 34; Gill 1981). At 1400°C and 30 kb Johnston (1986) produced andesitic liquids by 30% melting of eclogite under anhydrous conditions. Some inconsistencies arise, however, when one compares the experimental results with natural andesite compositions. The REE patterns of natural andesites suggest that garnet does not play a role in its petrogenesis (unless all garnet is consumed) as most andesites have fairly flat heavy REE patterns (Gill 1981). In addition, most natural andesites have between 1-4 wt% H_2O, inconsistent with dry melting conditions (Gill 1981). Partial melting of anhydrous ultramafic mantle compositions will generally produce liquids of basaltic or picritic compositions, but fusion of wet lherzolite may generate andesitic magmas (generally basaltic andesite), albeit water saturated (15 wt% H2O at 40 km) and would not reach the surface unmodified (Wilson 1989).

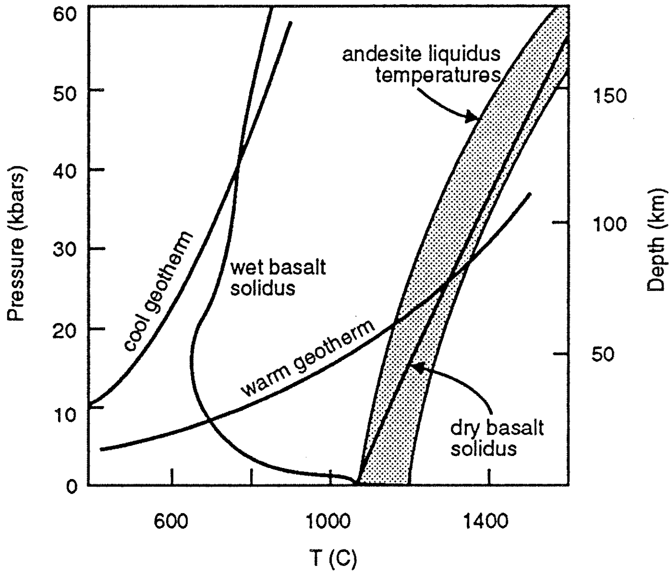


Fig. 1.34. P-T diagram showing the solidi for basalt under hydrous and anhydrous conditions with a shaded field representing the range of andesite liquidus temperatures. Also shown are two hypothetical geotherms corresponding to subduction of an old, cold oceanic slab (cool geotherm) and a younger, warmer subducting slab (warm geotherm). Modified from Gill (1981).

A more commonly applicable model for the generation of arc magmas involves hydration of the mantle wedge by subduction-related fluids and melts which would lower its solidus temperature and induce partial melting. In geological settings where lower geothermal gradients are present or subduction of an old and cold slab is occurring, the subducted slab will dehydrate before it can reach conditions amenable to melting, thereby releasing a LILE-enriched fluid/melt into the overlying mantle (Martin 1986; Tatsumi et al. 1986; Drummond and Defant 1990). Under higher geothermal gradients and where young, warm oceanic crust is subducting, the slab itself may partially melt to some extent and these melts rise into the overlying mantle wedge (Wyllie 1984). Experimental studies by Sekine and Wyllie (1982) on granite-peridotite systems document how such fluids can lower the solidus temperature of the mantle component. It is a matter of debate as to the exact nature of these fluids (they may be hydrous melts or siliceous fluids derived from the altered basaltic crust and sedimentary layer (Rapp et al. 1991)) but in either case, they are variably enriched in the LILE's and light REE's (Gill 1981). Fluxing of this fluid/melt into the mantle wedge will transfer some or all of these elements into the latter, and upon melting will incorporate them into the resulting partial melts. High concentrations of K, Rb, Th, U, light REE's, as well as marked depletions in the HFSE's (high field strength elements; e.g., Nb, Ta, Ti), are typical of volcanic and plutonic rocks derived in such a manner (e.g., Tatsumi et al. 1986; Davidson 1987; Ellam and Hawkesworth 1988; McCulloch and Gamble 1991; Hawkesworth et al. 1994). It is this distinct trace element signature that distinguishes magmas generated in subduction zone environments from all other settings (e.g., Pearce 1983; Pearce et al. 1984; Brown et al. 1984).

In oceanic island arc settings very little crustal contamination can occur, except via sediment subduction and/or intrusion of the arc magmas into, and reworking of, the emerging arc and associated accretionary wedge. In continental arcs, the presence of a thick continental crust and its lithospheric mantle adds increased complexity and difficulty in unravelling the petrogenetic origin of the magmas. It is generally believed that continental arc magmas are generated in the same manner as island arc magmas, but during ascent the primary magmas are modified due to interaction with the continental crust, which can be up to 70 km thick. In addition, the continental crust itself may be an important source for magmas. Depending on the age and composition of this crust, and the extent of contamination and/or crustal melting, the geochemical and isotopic effects may or may not be pronounced. Ancient differentiated crust has a significantly different composition compared to the potential arc magmas that could interact with it and will impart this signature to the magmas via assimilation. Higher concentrations of the LILE's and light REE's, as well as slightly higher contents of the HFSE's, are measured in magmas that have been generated in continental arcs compared to equivalent compositions from island arc settings (Pearce 1983; Brown et al. 1984). In addition, older continental crust will also impart a distinct isotopic signature to the magmas that interact with it, producing rocks with high $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ and low $^{143}Nd/^{144}Nd$, for example. However, if the crustal contaminant is young and the isotope tracer systems have not had the time to develop distinct compositions, then the isotopic effects of contamination may not be detectable.

Processes which modify primary magma compositions such as fractionation and combined assimilation-fractional crystallization (AFC) add to the geochemical variations acquired from the source rocks. In fact, the rarity of primitive magma compositions in arcs (e.g., Gill 1981; Nye and Reid 1986), particularly continental margin arcs, points to the important and consistent role of such processes. In continental margin settings the thick crust acts as a density filter, inhibiting the ascent of primitive magmas which are more dense than the overlying crust, forcing them to fractionate in order to rise to higher crustal levels (MASH process; Hildreth and Moorbath 1988) or inducing melting of the crust itself, producing more siliceous and less buoyant magmas that can ascent to higher levels (Huppert and Sparks 1988). Gill (1981) attributed fractionation of plagioclase + orthopyroxene/olivine + augite + magnetite as the most common process that generates andesitic liquids from basaltic parents, although various other phases including garnet, hornblende and apatite are also noted as contributing to the diversity of the magmas produced.

A generalized schematic model incorporating all of the above sources and processes that are involved in the petrogenesis of arc magmas is presented in Figure 35. Due to the large variety of potential sources and petrogenetic processes, magmatic arcs, and in particular continental arcs, are one of the most complex zones for magma genesis on Earth.

1.6.3 Petrogenesis of the Narsajuaq terrane magmas

Determining the exact origin of the Narsajuaq terrane plutons may not be possible, given the large range in composition and age of the rocks within any given suite. In addition, samples taken for geochemical and isotopic analysis were collected from plutons several to tens of km apart and may not be directly related petrogenetically to one another. Some of the scatter seen in the geochemical diagrams may be due to this regional scale of sampling. Detailed calculations of partial melting, crystal fractionation and assimilation-fractional crystallization processes for the Narsajuaq terrane plutons are pointless at the present time given these above limitations. Nonetheless some general calculations, commentaries and conclusions concerning the origin and evolution of these rocks can be made and a model for their petrogenesis proposed. Table 5 presents a summary of the geological, geochemical and isotopic characteristics of the various plutonic suites which synthesizes all the available data for the Narsajuaq terrane. This table may aid the reader in the following discussion of the petrogenesis of the magmas.

A basic assumption in most petrological studies is that the samples under question are representative of the original magma sources and that they have not been subsequently modified by mineral accumulation or alteration (i.e., they represent liquid compositions). In the case of fresh volcanic rocks this may be a relatively straight forward matter but in intrusive rocks, particularly evolved compositions, this is more difficult to evaluate. The samples of the Narsajuaq terrane plutons do not contain obvious cumulate textures and generally do not have geochemical signatures indicative of mineral accumulation (e.g., positive Eu anomalies) although a few exceptions have

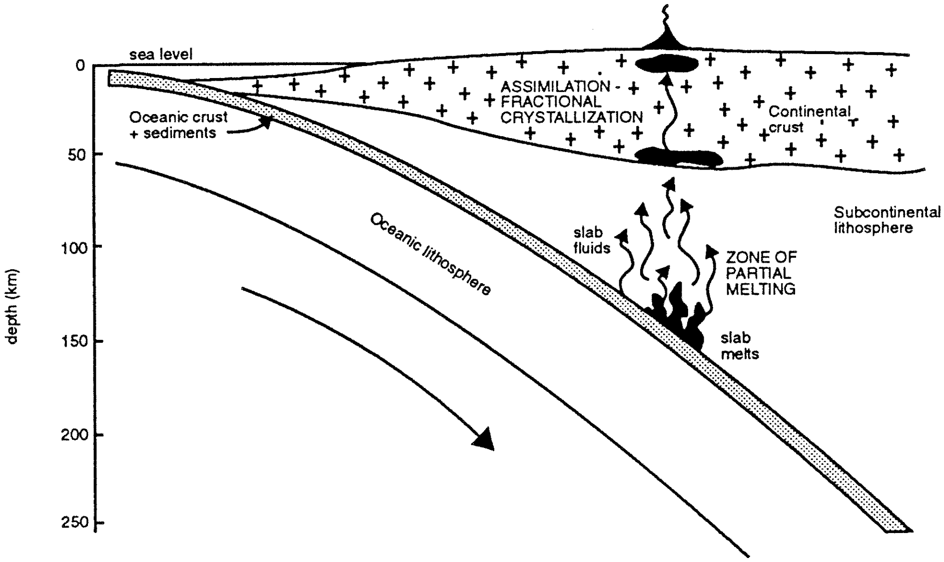


Fig. 1.35. General schematic diagram illustrating the petrogenetic processes and possible sources involved in arc magmatism along an active continental margin.

Table 5. Summary of geological characteristics of the Narsajuaq terrane suites

	OLDER SUITE	YOUNGER SUITE	CAPE SMITH SUITE	LATE SUITE
AGE	1863-1844 Ma	1836-1800 Ma	1898-1839 Ma	1758-1742 Ma
COMPOSITION	diorite, tonalite, granite; tonalite predominates; rare peridotite sills	diorite, monzodiorite, tonalite, granite; granite predominates	diorite, tonalite, granite; mafic and intermediate compositions predominate	granite, syenogranite
OUTCROP STYLE	layered sequence: diorite-tonalite-granite; main base unit of terrane	discret, homogeneous km-size bodies; intrusive into Older suite	discrete plutons intrusive into Cape Smith Belt units	small plutons and pegmatitic dykes
DEFORMATION	well foliated, important compositional layering	range in deformational style from well foliated (generally older plutons) to massive	well foliated to massive	generally massive
GEOCHEMICAL SIGNATURE	calc-alkaline, medium-K, moderate to pronounced negative Nb, Ti anomalies, moderate to high LILE, fractionated LREE (La_n/Yb_n = 1.8-73.3)	calc-alkaline (alkaline monzodiorites), medium- to high- K, moderate to pronounced negative Nb, Ti anomalies, moderate to high LILE, fractionated LREE (La_n/Yb_n = 4.4-62.0)	calc-alkaline and tholeiitic, medium-K, small negative Nb, Ti anomalies, low to moderate LILE, slightly fractionated LREE (La_n/Yb_n = 2.3-25.1)	peraluminous, high-K, moderate to pronounced negative Nb, Ti anomalies, moderate to high LILE, fractionated LREE (La_n/Yb_n = 15.8-91.6)
$\varepsilon_{Nd\ (initial)}$	+4.0 to -10.7	+3.9 to -18.5	+3.2 to +1.5	-1.6 to -9.7
$^{87}Sr/^{86}Sr_{\ (initial)}$	0.7025 to 0.7048	0.7020 to 0.7064	0.7020 to 0.7024	0.7036 to 0.7051
MAGMA SOURCES	mutiple sources: slightly contaminated (seds, intra-crustal) mantle magmas to mixed mantle/crustal melts	mutiple sources: variably contaminated (seds, intra-crustal) depleted mantle and enriched mantle magmas; older Narsajuaq terrane crust; Archean Superior Province crust (average 2.7 Ga and older 3.1 Ga)	minor contamination (sed. subduction) of depleted mantle magmas; fractionation to produce intermediate and felsic components; and/or partial melting basaltic crust	crustal sources: Narsajuaq terrane crust; average 2.7 Ga and older 3.1 Ga Superior Province crust

been noted. The amount of internal differentiation within any individual pluton is limited (no compositionally zoned plutons have been mapped) although some textural variations are noted, particularly in the huge, late granitic sheets of the Younger suite. In addition the plutons are geochemically homogeneous (at the scale of sampling) and have overall coherent geochemical trends. The effects of alteration have already been addressed in the geochemistry section, and it was concluded that the geochemical analyses of the Narsajuaq terrane plutons are generally robust and have not been affected by significant alteration. All these factors indicate that the chosen samples do indeed represent magmas that have not been greatly modified by mineral accumulation or alteration processes and can be said to approximate melt compositions.

The samples of the Narsajuaq terrane plutons all represent fractionated, non-primary mantle melts, as they contain concentrations of Mg, Ni and other compatible elements that are too low to have been directly derived from an unmodified primary mantle melt (Fig. 36; Elthon 1989). The most primitive non-cumulate rocks of the Narsajuaq terrane are the diorites, and it is possible that such compositions represent parental magmas for the rest of the series. Only one sample (L90b) has even moderate Ni concentrations of 146 ppm at 7.3 wt% MgO, which is below the 200-250 ppm estimated by Elthon (1989) for primary melts at this MgO content. In this context it is difficult to infer what the original primary source composition may have been, especially given the heterogenous nature of the mantle and overlying lithosphere. With these constraints in mind, we can nevertheless present some probable scenarios and rule out certain other possibilities.

Crustal contamination may significantly modify the composition of a resulting magma, as was noted in the previous section. Typical upper continental crust contains high concentrations of the LILE's and light REE's whereas lower crust is depleted in such elements (Taylor and McLennan 1985). Bulk assimilation of such crust should be readily identifiable in the resulting geochemical signature of the magmas due to their pronounced chemical differences. Assimilation - fractional crystallization (AFC), how-

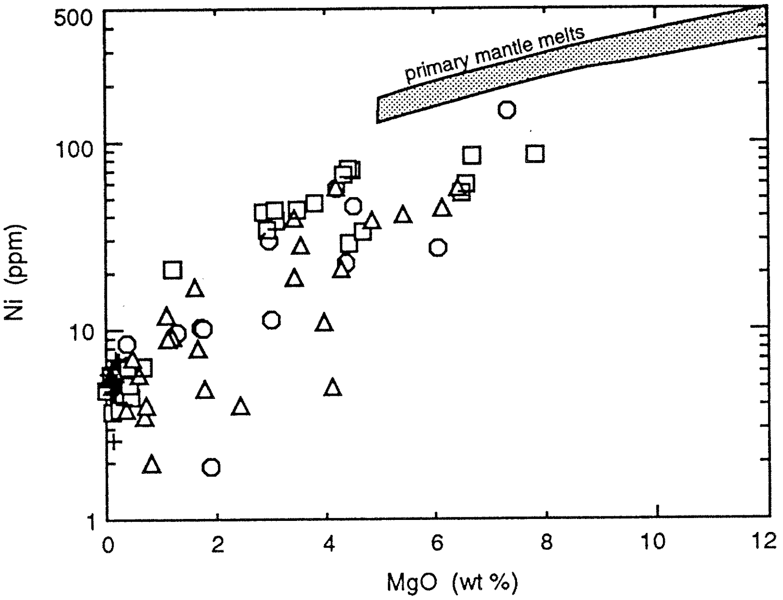


Fig. 1.36. MgO versus Ni for the Narsajuaq terrane plutons demonstrating their non-primary nature. The shaded field for primary mantle melts is from Elthon (1989).

ever, is a more complex process in which the interplay between the assimilation and fractionation may or may not significantly enrich the element in question depending upon the minerals involved and their Kd's for the particular elements. The most obvious crustal contaminant within the Ungava Orogen is the Archean Superior Province basement rocks. The average composition of the Superior Province rocks (Fig. 37) exposed in the eastern part of the orogen (Harvey 1995) is very similar to the estimate for average upper continental crust by Taylor and McLennan (1985) although lower Rb, Nb, Ti, Y and the heavy REE are found in the former. It is this average composition of the Superior Province crust that is used in subsequent geochemical calculations to evaluate crustal contamination. The isotopic composition of this crust (average 2.7 Ga and old 3.1 Ga crust) was discussed in a previous section. In most samples we cannot constrain which of the two Superior Province crusts could have been the contaminant; hence, mixing calculations have been done using both possible contaminants. Only in the case of samples with isotopic compositions more enriched than 2.7 Ga crust can we ascertain that the contaminant most likely was the older 3.1 Ga crust. Detailed geochronology on the Narsajuaq terrane samples to identify the possible presence of older zircons and their ages could help identify which crustal sources were present (as was done for the Lac Duquet pluton from the Late suite; Dunphy et al. in press) and could be a possible avenue to pursue. Another possible crustal contaminant that could be important in the petrogenesis of the younger plutons from the Narsajuaq terrane (the youngest granite of the Younger suite and the Late suite granites) is the emerging Narsajuaq terrane crust itself. This possibility is addressed in later sections when the petrogenesis of these granites is discussed.

1.6.3.1 *Cape Smith suite*

Discussion of the petrogenesis of the Narsajuaq terrane plutons will begin with the Cape Smith suite, as these samples all have juvenile isotopic compositions indicative of little older crustal contamination. The similarity in bulk composition and the spatial and temporal relationship of the plutons of the Cape Smith suite to the other suites of the Narsajuaq terrane suggests that they are related. Determining the origin of the rel-

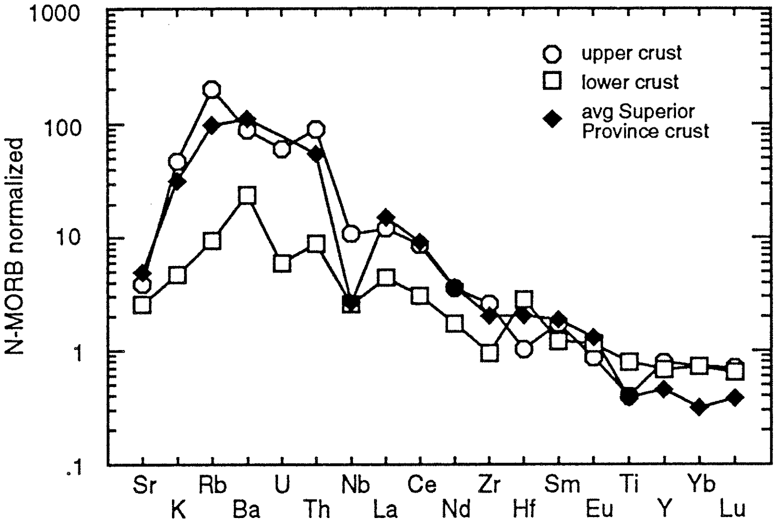
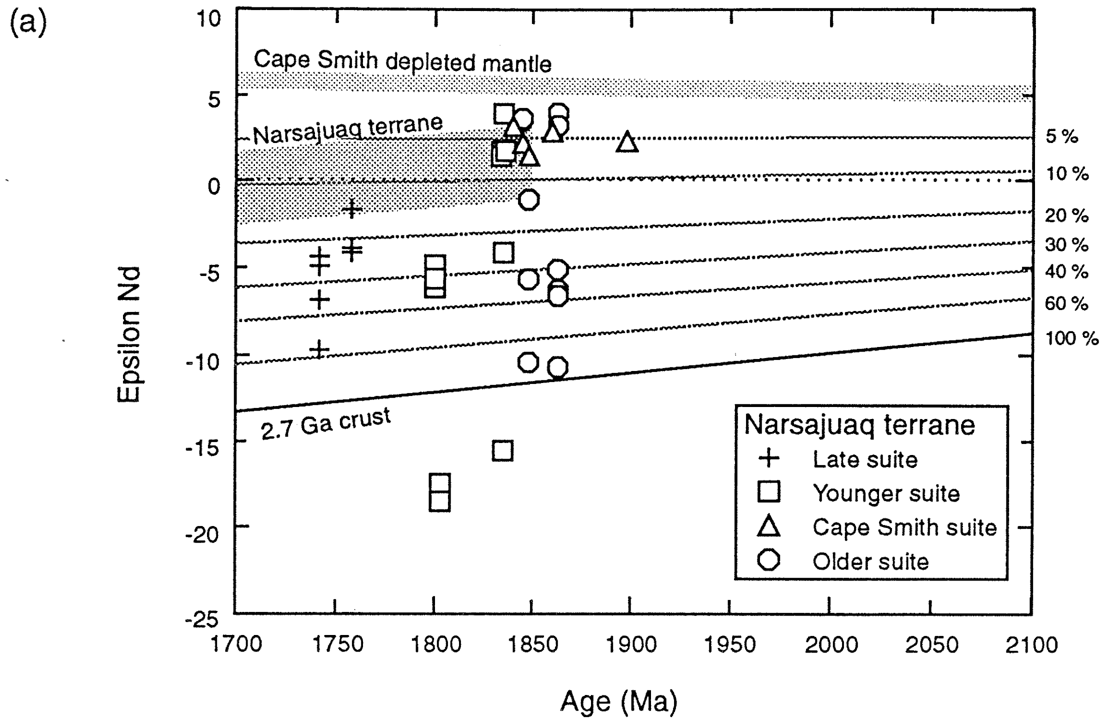


Fig. 1.37. N-MORB normalized (Sun and McDonough 1989) trace element variation diagram for typical crustal compositions. Upper and lower continental crust from Taylor and McLennan (1985), data for average Superior Province crust from Harvey (1995).

atively uncontaminated Cape Smith suite can therefore provide a basis for comparison with the rest of the Narsajuaq terrane plutons.

The Cape Smith suite plutons were intruded over a period of approximately 60 Ma (Table 1) and do not show any systematic change in composition over this time (i.e., they do not become progressively more felsic with time). The long time span of magmatic activity suggests that the plutons are not all directly related to one another; we can nevertheless extract meaningful petrological information from them. Initial ε_{Nd} values range from +3.2 to +1.5 for various compositions, consistent with minor contamination (4-8%) of depleted mantle magmas by average 2.7 Ga Superior Province crust or sediments derived from such crust (Fig. 38a). If the contaminant contains older material (e.g., 3.1 Ga crust) then only 2-4% contaminant is necessary (Fig. 38b). Sr isotopic analyses on several samples confirm these results. Given that these samples are intrusive into the predominantly mafic units of the Cape Smith Belt it is unlikely that the contaminant was the continental crust itself; crustally derived sediments are a more plausible source. The moderate enrichment of the LILE's and light REE's in the Cape Smith plutons relative to typical depleted mantle magmas may be due to such contamination.

The most primitive rocks in this suite are diorites with MgO contents of 6.1-6.4 wt % and Ni =57 ppm, with only slightly fractionated REE and relatively low concentrations of the LILE's (Fig. 39), and it is possible that the tonalite and granite plutons were derived from such parental magmas. A slight enrichment in LILE's and light REE's is noted even in these most primitive samples however, suggesting a source control for these elements. Fractionation of hornblende, clinopyroxene and plagioclase (the original igneous minerals in the diorites) from such a source would only slightly decrease the REE content of the subsequent magmas. However the relatively extreme decrease of the heavy REE with increasing differentiation (e.g., Yb decreases from 3.4 ppm in the most primitive diorite to 1.3-0.4 ppm in the tonalite and granite samples; Fig. 24) suggests that a mineral having a high Kd for these elements was involved. A minor



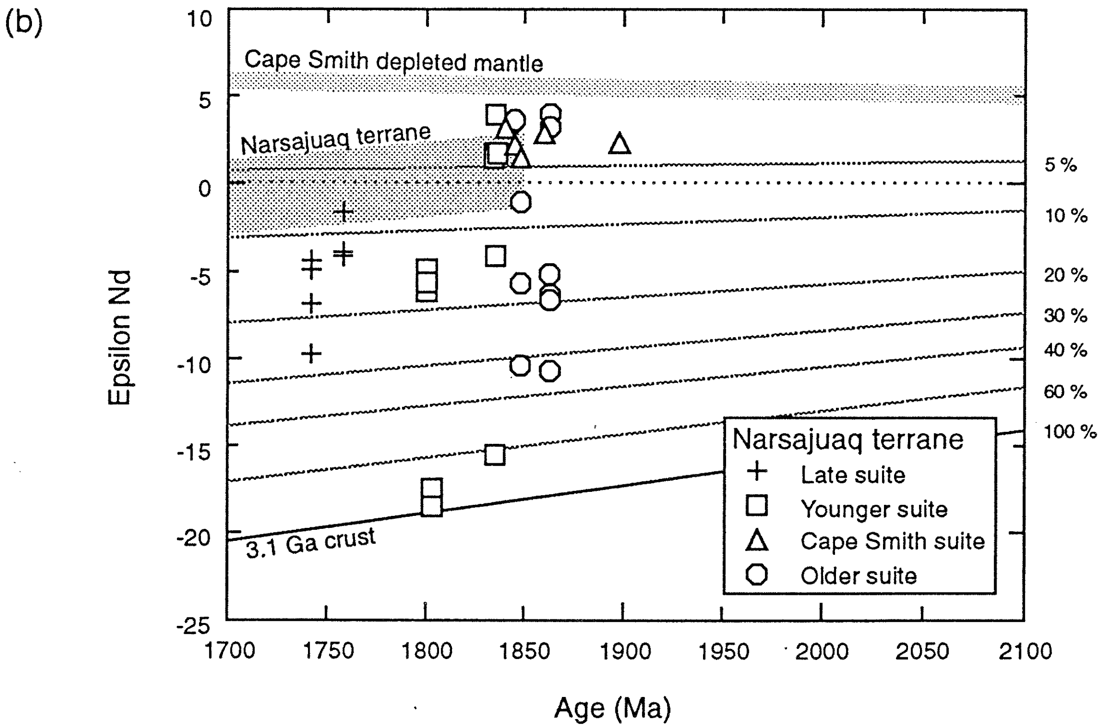


Fig. 1.38. Epsilon Nd versus age for the Narsajuaq terrane samples with mixing lines superimposed for bulk assimilation of (a) average 2.7 Ga Superior Province crust and (b) 3.1 Ga Superior Province crust. Numbers to the right of the mixing lines indicate % of Superior Province contaminant. Mixing is between average Cape Smith depleted mantle composition (Dunphy et al. 1995) and Superior Province crustal compositions (Harvey 1995). Also shown from 1.85 Ga to 1.70 Ga is a field for the composition of Narsajauq terrane rocks which could have represented an important relatively juvenile component in some plutons (CNF/IVUJ granites and Late suite plutons).

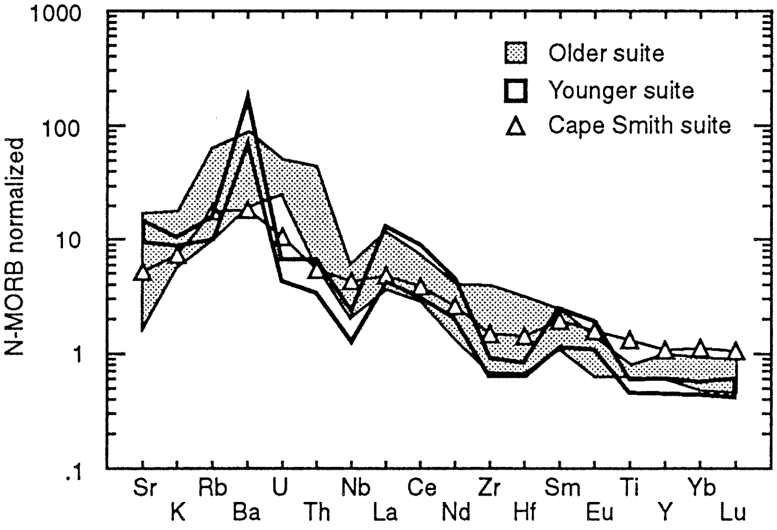


Fig. 1.39. N-MORB normalized (Sun and McDonough 1989) trace element variation diagram for the most primitive samples from the Narsajuaq terrane.

amount of garnet (<5%) in the fractionating assemblage would cause a decrease in the heavy REE content of the resulting magma. However, this mineral would also result in significant fractionation of the heavy REE, which is not observed. Fractionation of minor amounts of accessory phases such as apatite, titanite and zircon provides a better mechanism to reduce the REE contents, as these phases contain relatively high concentrations of these elements and, apart from zircon, do not fractionate the REE to any great amount. Plagioclase fractionation would result in decreases in Al_2O_3 and Sr contents and a larger negative Eu anomaly going from dioritic to granitic compositions due to their compatible nature in plagioclase. Al_2O_3 concentrations decrease only slightly while Sr and Eu show no discernable trends suggesting that plagioclase was not an important fractionating phase for these samples, assuming they represent liquid compositions. It is possible, however, that these samples may represent crystal-mush assemblages, thereby accounting for the lack of geochemical trends for Sr and Eu. One sample (D5-2) has a positive Eu anomaly (Fig. 24a) potentially related to plagioclase accumulation and may represent the residual fractionating assemblage which has an overall REE content that is similar to the diorite samples. With the exception of Ba which correlates positively with SiO_2, most other LILE's do not vary in any systematic manner (Fig. 21), and the overall range in the concentrations of the LILE's for the tonalite and granite samples overlap those of the diorites. Numerical calculations indicate that fractionation of an amphibolite to granulite assemblage (hornblende-clinopyroxene-plagioclase-accessory phases) from a parental melt of a composition similar to the most depleted diorite in the suite can account for some of the range in the concentration of the LILE's in the resulting melts, although several parents of slightly different compositions are necessary to explain the petrogenesis of all the Cape Smith suite samples (Fig. 40). A minor amount of contamination by subducted sediments having the same composition as average Superior Province crust and concurrent fractionational crystallization (AFC) can also account for some of the compositional range of the Cape Smith suite.

One tonalite sample (D284) has a slightly higher light REE content and a more

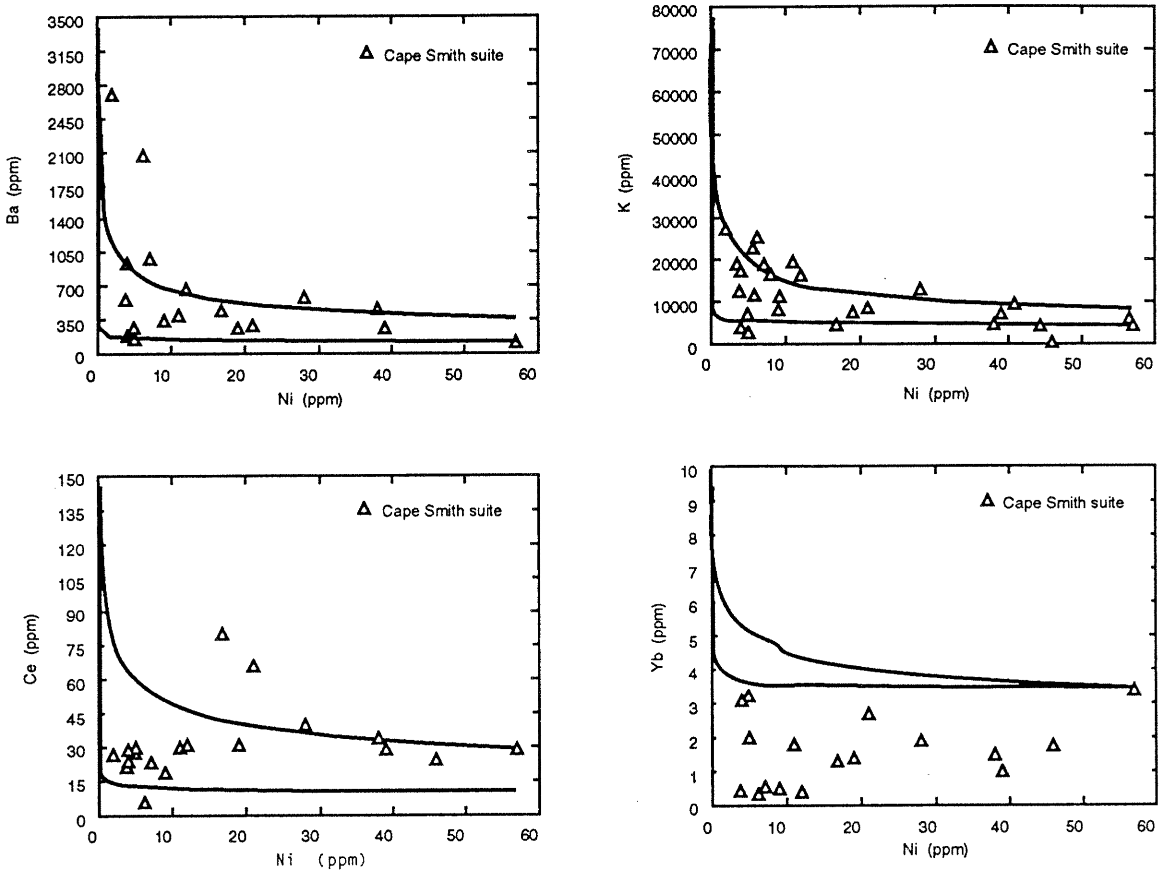


Fig. 1.40. Binary trace element diagrams for the Cape Smith suite with lines representing the range in compositions that would be produced via fractionation of an amphibolite to granulite assemblage from parental magma having a composition similar to the most primitive sample in the suite.

fractionated light REE pattern relative (Fig. 24b) to the most primitive diorite which could be the result of a slightly greater amount of interaction with light REE enriched crustal material (i.e., more contamination), yet this sample has the most depleted isotopic signature (initial ε_{Nd} = +3.2) of all the analyzed plutons from this suite. A plausible explanation for the origin of this sample is that it was derived from a source that was more light REE enriched than the above proposed parental magma. An enriched isotopic reservoir was postulated to have been present in the Ungava Orogen at 2.0 Ga (Dunphy et al. 1995) and may have served as a source for some arc magmas. However, local source enrichment due to the release of LILE- and light REE-enriched fluids from the subducting slab could also have been a viable mechanism. The greater light REE enrichment for sample D284 is not correlated with its LILE contents, as this sample has generally low concentrations of the LILE's relative to the other tonalite samples, although it does contain the highest concentration of Nb and has the highest Nb/Y ratio, similar to typical E-MORB or OIB (Sun and McDonough 1989). This sample clearly illustrates the difficulty in trying to constrain the origin of the Narsajuaq terrane samples, as it is almost impossible in a regional-scale study to find solutions for the petrogenesis of all samples. It is most likely that several different parental melts were produced from a variably light REE and LILE enriched source which subsequently fractionated to produce the range in compositions observed in the suite.

1.6.3.2 *Older suite*

The Older suite plutons were intruded over a more restricted period of time (20 Ma; Table 1) compared to the Cape Smith suite. They have a wide geographic distribution, and therefore the samples chosen for study are probably not directly related although they may have had similar sources. Field relationships indicate that the diorite and tonalite are related to each other, as they are often found intimately mixed in a layer-cake fashion (along with syntectonic granite veins), and although the geochronology indicates that the diorite samples are younger than the tonalite (two diorites are dated at 1844-1845 Ma while a tonalite is 1863 Ma), there are examples of tonalite cross-cutting diorite. Hence there appears to have been several pluses of magmatic activity during the 20 Ma period of magmatism.

A field representing the range in compositions for the most primitive samples from the Older suite are presented on a trace element variation diagram (Fig. 39), and although some of these samples are more primitive than those from the Cape Smith suite (higher MgO, Ni and ε_{Nd}), most samples have slightly higher trace element concentrations which suggests that the source compositions were somewhat different. One primitive Older suite sample (L90b) has 7.3 wt% MgO, 146 ppm Ni, relatively flat REE's ($La_n/Yb_n = 4.0$) and an ε_{Nd} value of +3.6. This sample has been dated at 1845 Ma (R. Parrish, personal communication 1994) and although the older samples evidently could not have been derived from it, a similar composition(s) could have been parental to the suite as a whole. A progressive enrichment of the light REE and a decrease in the concentration of the heavy REE with increasing differentiation from diorite to tonalite compositions is noted (Fig. 22). Concurrent contamination by older continental crust and fractionation of hornblende, plagioclase, clinopyroxene, orthopyroxene and accessory phases (AFC-type process) could cause the slight decrease in heavy REE and enrich the light REE's and also account for the observed LILE concentrations and isotopic signature of the derivative plutons. As was the case for the Cape Smith suite, simple fractionation or AFC from the single most primitive sample cannot explain the entire range of compositions present in the Older suite plutons. Similar to modern arc environments, a range of parental compositions is necessary.

The Older suite plutons span a range of ε_{Nd} values with several diorite samples clustering close to depleted mantle values, indicating little contamination by crustal material, while one diorite and all the tonalite and granite samples have enriched signatures and contain a significant crustal component (Fig. 38) which ranges from 8-90% depending on the contaminant: old 3.1 Ga crust or average 2.7 Ga crust. The Sr isotope values correlate well with the Nd results and confirm that one diorite sample analyzed contains little crustal material while the tonalite and granite samples have higher initial Sr isotopic compositions closer to that of older (Archean) continental crust

(Fig. 28). It is interesting to note that the diorite sample L211 which indicates significant Nd isotopic contamination (≈20-35%, depending on the age of the contaminant) does not have higher light REE contents and has noticeably lower Rb, Ba, Sr and K contents relative to the uncontaminated diorites. Contamination by lower continental crust material (which contains lower concentrations of these elements relative to upper crust; Fig. 37) or derivation from a source that is more depleted than L90b could produce such a signature in this diorite.

Several samples of the granitic veins from the Older suite consist primarily of plagioclase + quartz ± K-feldspar and have positive Eu anomalies indicative of plagioclase accumulation (Fig. 22c). These samples could represent residual crystal-mush assemblages that had fractionated from contaminated tonalitic parents. Two other granitic vein samples (L27 and D14a) may represent liquid compositions derived via this fractionation, which would explain the overall lower trace element abundances of these granites compared to the tonalites. All the granitic vein samples have enriched Nd isotopic signatures (ε_{Nd} = -1.1 to - 10.4). One tonalite (D72) and one granite (D41a) have very low initial ε_{Nd} values (-10.7 and -10.4, respectively) which is almost as enriched as average 2.7 Ga Superior Province crust at the time of their formation (Fig. 38a). These samples may have been derived entirely by moderate to large degrees of partial melting of this pre-existing crust or could have attained this isotopic signature through ≈35-40% contamination of a juvenile magma with the older 3.1 Ga crustal component (Fig. 38b). Another possibility to explain the isotope signature of the tonalite sample (D72) is that it is actually Archean in age (it has not been dated) which would give an ε_{Nd} value of +0.7 at 2.7Ga, although it was correlated to other Older suite rocks during field work investigations. In any case the sample suggests that old Archean material was present and involved in the petrogenesis of the Older suite plutons. The granite sample is Early Proterozoic and has been dated at 1848 Ma (R. Parrish, personal communication 1994).

1.6.3.3 *Younger suite*

The majority of the Younger suite plutons range in age from 1836-1821 Ma, with a large volume late granitic event ca. 1800 Ma (Table 1). Similar to the other suites, all compositions from diorite to granite are present although intermediate compositions are not prevalent. The most primitive Younger suite diorites have a more restricted range in composition than the Older suite diorites, and contain lower abundances of most elements with the exception of Ba which is higher in the former (Fig. 39). Of the three diorite samples analyzed for their isotopic composition, one has a composition close to that of the depleted mantle whereas the two others indicate that they have interacted with continental crust (Fig. 38). Comparison of only these three samples (Fig. 41) shows that the contaminated samples have overall higher trace element abundances (except for Sr and Ba), consistent with crustal contamination. However the contaminated samples have relatively high MgO and Ni contents, similar to the most primitive compositions in the suite, as well as high Fe_2O_3 (12.6-14.9 wt%), Nb and Y contents. The geochemical signature of these samples indicates a source control for the enrichement of certain trace elements. Derivation of these samples from a primary enriched mantle source followed by secondary crustal contamination may explain the geochemical signature of these samples. The isotopic composition of these diorite samples is consistent with 10% and 60% contamination by 3.1 Ga Superior Province crust (S106 and K69, respectively) of a melt derived from an enriched mantle composition ($\varepsilon_{Nd} = +2.5$ at 2.0 Ga; Dunphy et al. 1995).

The Younger suite also contains a very distinct series of rocks of monzodioritic composition having a particular major and trace element composition. The monzodiorites are K-rich and have exceptionally fractionated REE patterns and enriched LILE concentrations that are strikingly different from any other series of rocks (Fig. 23b, 21). They are relatively primitive isotopically with ε_{Nd} values of +1.7 to +1.4, which indicates a maximum of 8% contamination by average 2.7 Ga continental crust (Fig. 38). Moderate values of MgO and Ni and other compatible elements indicates that the monzodiorites are not primitive melts but have had at least a two-stage history. The

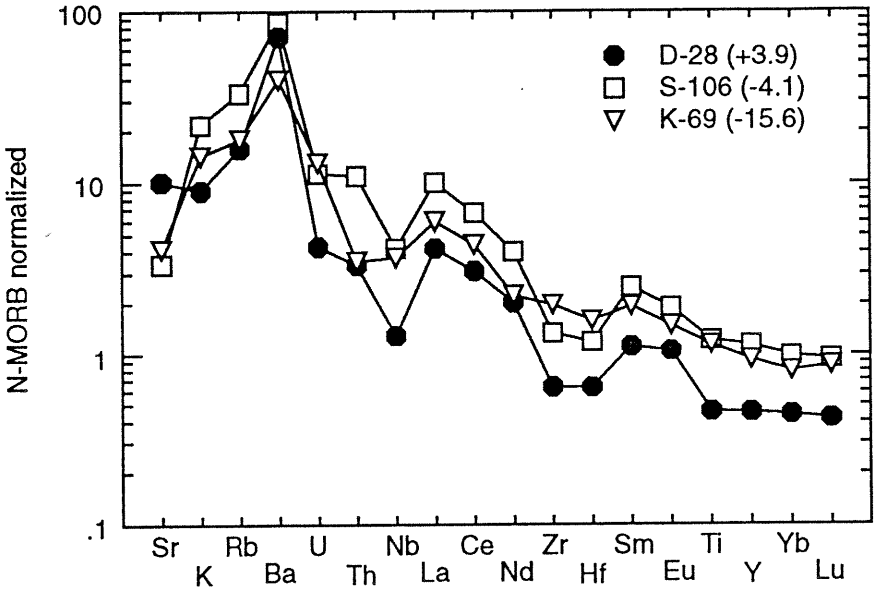


Fig. 1.41. Trace element variation diagram for three younger suite diorites which have a range in Nd isotopic composition. Epsilon Nd (1835 Ma) indicated next to sample name in brackets.

monzodiorites cannot be related by fractional crystallization to the diorites as derivative melts do not attain the requisite high concentration of the LILE's and light REE's at the desired Ni contents. Small to moderate degrees of partial melting (rather than fractional crystallization) of a LILE and light REE enriched mantle source, having a composition similar to that proposed by Stern et al. (1989) for the source of Archean monzodiorites (sanukitoids) in the Superior Province, could produce the range in compositions observed in the Narsajuaq terrane monzodiorites. The enrichment of the mantle souce could be directly linked to the subduction process whereby dehydration of the downgoing slab results in the release of a LILE- and light REE-enriched fluid/melt that rises and interacts with the overlying mantle.

The Younger suite granites sampled for petrological study come from the large relatively undeformed granitic sheets that intruded late (1803-1800 Ma) in the development of the Narsajuaq terrane. Although the samples from the Ivujivik and Cap Nouvelle France areas have many similarities in terms of field appearance, timing, relative deformation and bulk composition, some geochemical and isotopic differences are noted. Two samples from the CNF granite sheets have relatively low REE contents and positive Eu anomalies (Fig. 23c) and have been interpreted as products of mineral accumulation. The remainder of the CNF samples have significantly higher REE contents relative to the IVUJ samples although the LILE's in the samples from the two regions overlap with the CNF samples having the largest range. The isotopic signatures for the granites from the two regions are also significantly different. The IVUJ samples have ε_{Nd} values of -4.9 to -6.2 indicative of 25-35% contamination of depleted, juvenile magmas by average 2.7 Ga crust or only 15-18% contamination by older 3.1 Ga crust (Fig. 38). The CNF granites on the other hand can only be derived from the older crustal component as they have ε_{Nd} values of -17.5 to -18.5, much lower than average 2.7 Ga crust (Fig. 38). The Sr isotope data for the two granite series are opposite to the Nd data as the CNF granite has a relatively low initial $^{87}Sr/^{86}Sr$ value of 0.7035 whereas the IVUJ granite has a high value of 0.7064.

The Nd isotope data for the IVUJ granites suggests that both a juvenile and evolved component were involved in their petrogenesis. Assimilation of continental crust by a depleted mantle magma with simultaneous fractional crystallization (AFC) could explain the isotopic composition of some of these samples, however no mafic magmas corresponding to the age of these plutons (≈ 1800 Ma) have been identified within the Younger suite, suggesting that such mafic magmas, if they existed, may have pooled at the base of the continental crust where the density contrast between it and the felsic crust inhibited its rise into the upper crust. Another possible source therefore, for the juvenile isotopic signature in these granite sheets, is the previously constructed Narsajauq terrane crust. Such crust would have a Nd isotopic composition (ε_{Nd}) of approximately +2.5 to -1.7 at 1.8 Ga (Fig. 38) and a mixture of ≈ 25-35% of crust of this composition with 75-65% average 2.7 Ga Superior Province crust can explain the isotopic composition of the IVUJ granites. The Nd isotopic signature of the CNF samples indicate that crustal sources were very important in their petrogenesis and may have been the sole source. Partial melting models involving the Superior Province crust and the juvenile Narsajauq terrane crust indicate that partial melting of a mixture of these crusts (≈80% Narsajuaq terrane and 20% old Superior Province) can explain the Nd isotopic composition of these rocks (Fig. 38).

1.6.3.4 *Late suite*

The Late suite consists of syenogranite pegmatite dykes and small rare granitic plutons intruded at 1758 Ma and 1742 Ma, respectively (Table 1). The petrogenesis of one Late suite granitic pluton (the Lac Duquet monzogranite) was discussed in detail in a previous paper (Dunphy et al. in press), which documented a crustal origin for the granite. The granite contains significant inherited zircon grains ranging in age from 1.7 to 3.2 Ga which corresponds to the ages of the host rocks for the pluton. Initial ε_{Nd} ranging from -4.4 to -9.7 indicate a significant crustal component; it was determined that at least three sources, namely the Narsajuaq terrane (the Older and Younger suites), average 2.7 Ga and older 3.1 Ga Superior Province continental crust, were involved in the petrogenesis of the pluton. The geochemical signature of the granite is similar in composition to the surrounding continental crust and to other crustally derived granites. Anatexis of the continental crust surrounding the pluton due to structural thickening during the waning stages of the Ungava Orogeny is believed responsible for the generation of the Lac Duquet pluton and was the source for its inherited zircons (Dunphy et al. in press).

The syenogranite pegmatites of the Late suite have a slightly different composition than the Lac Duquet granites in that the former has higher overall abundances of the light REE's (Fig. 25), Sr and Ba, although U and Nb are lower. The Nd isotopic composition of the pegmatites is not as low as the Lac Duquet granite, with ε_{Nd} values of -1.6 to -4.2, which is consistent with partial melting of a mixture of average Narsajuaq terrane crust and Older Superior Province crust (Fig. 38). The presence of inherited monazite crystals in these pegmatites that correspond to the age of the surrounding Narsajuaq terrane plutons (Parrish 1990) supports the isotopic data suggesting that such rocks were involved in their petrogenesis. These inherited monazite grains may be responsible for the high light REE contents of the syenogranites as monazite is significantly enriched in these elements. It is proposed that the syenogranite pegmatites had a similar origin as the Lac Duquet granite and the Younger suite granites although a larger juvenile component is visible in the isotopic signature of the pegmatites.

1.6.4 Geotectonic model for the origin and evolution of the Narsajuaq terrane

The geological information presented here for the Narsajuaq terrane allows us to present a model for its origin and evolution within the larger, more regional, tectonic framework of the Ungava Orogen. Geotectonic models for the origin of the Cape Smith Belt and the Ungava Orogen have previously been presented (e.g., Francis et al. 1983; Picard et al. 1990; St-Onge and Lucas 1990b; St-Onge et al. 1992; Dunphy et al. 1995) and will only be discussed briefly here as the main goal in this present paper is to present a model that explains and incorporates the Narsajuaq terrane into the bigger picture of the Ungava Orogen.

1.6.4.1 *Arc magmatism and the Narsajuaq terrane*

The arc-related magmas of the Ungava Orogen span almost 80 Ma, from 1898 to 1821 Ma (Parrish 1989; St-Onge et al. 1992; Machado et al. 1993; R. Parrish, personal communication 1994), and are represented by both intrusive (Narsajuaq terrane) and extrusive (Parent Group) magmatic activity. The geological and geochemical data presented here demonstrate the similarity of the Narsajuaq terrane rocks to younger (Mesozoic) magmatic arc products, both in terms of geological field relationships and geochemical signature although slight differences have been noted which may be related to different source geochemistry.

Within the Narsajuaq terrane, the plutons of the Cape Smith suite represent the oldest subduction-related magmas (1898-1839 Ma; Table 1), which were emplaced into the lithologies of the Watts, Spartan and Parent Groups. Although these plutons are not primary mantle melts, they did not undergo significant modification (i.e., contamination by older crust) apart from fractional crystallization from their parental compositions. The isotopic signature of these plutons indicates a maximum of 8% crustal contamination, which is interpreted as related to sediment subduction. The slight enrichment in LILE's and light REE's in the most primitive samples from this suite relative to MORB indicate a source control for these elements.

The timing of the intrusion of the Older suite overlapped slightly with that of the Cape Smith suite, occurring from 1863-1844 Ma (Table 1). Some small differences in the geochemical signature have been noted between the Older suite and the Cape Smith suite in that slightly higher concentrations of the LILE's and the light REE's are found in the former, as well as a much larger range in isotopic composition. The Older suite magmas represent variably contaminated magmas derived from a range of parental melt compositions which are slightly less primitive than those of the Cape Smith suite. Some samples from the Older suite are as isotopically enriched as the Superior Province basement rocks, pointing to a significant crustal contribution. The presence of older (Archean) zircons within at least one Older suite sample (S-216) also points to a significant crustal component in the petrogenesis of these rocks. We suggest therefore that the crustal foundation for these rocks was an isotopically evolved crustal fragment, possibly part of the previously rifted Archean Superior Province. The least contaminated samples in the Older suite may have intruded regions that were peripheral to this crustal fragment or could have been injected along pre-contaminated conduits which did not come directly in contact with the older crustal contaminant. Fractionation and assimilation-fractional crystallization from parental magmas generated in this manner produced the more geochemically and isotopically evolved compositions of this suite.

A small hiatus in the magmatic activity occurred between 1844-1836 Ma and was followed by another 15 Ma of plutonic activity of the Younger suite (1836-1821 Ma; Table 1). A distinct series of rocks, monzodioritic in composition, are found primarily in the Younger suite and may have had an origin quite different than the other plutons of the Narsajuaq terrane. These K-rich rocks contain very high concentrations of the LILE's and light REE's and have pronounced fractionated REE patterns. Partial melting of a LILE- and light REE-enriched mantle source can account for the compositions of these monzodiorites, which are similar to the sanukitoid suite of the southwestern Superior Province (Stern et al. 1989). Significant negative Nb and Ti anomalies are present in the trace element pattern for these rocks, typical of subduction-related magmas.

The overall geochemical and isotopic signature of the Younger suite diorites is similar to those from the Older suite, although the most primitive samples from the Younger suite have lower trace element abundances (with the exception of Ba), suggesting that the sources for the two suites may have been slightly different; however, their petrogenesis and evolution may have been similar. As was the case for the Older suite, variable interaction with older crust occurred, producing a range in isotopic compositions for this suite. Plutons of intermediate composition are also present in the

Younger suite and were most likely derived from the more mafic and primitive diorites. The homogeneous nature of the Younger suite plutons contrasts with the more layered gneissic appearance of the Older suite, suggesting emplacement at higher crustal levels compared to the Older suite, possibly related to the partial exhumation of the Narsajuaq terrane crust during D_1 deformation (St-Onge and Lucas 1992).

Following another short period of magmatic quiescence, large sheets of granite intruded at 1800-1803 Ma (Table 1). These granites have a similar intrusive character as the rest of the Younger suite (large homogenous bodies, slightly but variably deformed) but have a geochemical signature indicative of derivation primarily (or exclusively for the CNF granites) from pre-exising older continental crust such as the Superior Province and/or the Narsajauq terrane itself. The slightly deformed nature of these plutons suggests that these plutons were syn-tectonic intrusions, and their age of emplacement coincides with the timing of the metamorphic retrogression event (1.81-1.79 Ga; Scott and St-Onge, in press). In order to generate such a large volume of felsic, evolved magma at this particular time in the geological evolution of the orogen, it is postulated that heating of the lower crust (possibly by tectonic thickening and/or by the presence of mafic magma which underplated the crust) accompanied by the influx of fluids related to the retrogression event (see earlier section on metamorphism) resulted in partial melting of both the Archean (2.7 and/or 3.1 Ga) and the newly formed Proterozoic crust. The extreme difference in the isotopic signature of these youngest granitic magmas may reflect variation and heterogeneity in the age and composition of the crustal basement for the magmatic arc complex.

Late suite magmatism occurred ≈ 40-60 Ma after the intrusion of the Younger suite granitic sheets, during the waning stages of the regional Ungava Orogeny. This magmatic activity is mainly manifest in the form of small granitic plutons and orogen-wide cross-cutting pegmatite dykes. The presence of significant inherited components (zircon and monazite) as well as the geochemical and isotopic signature of these granites suggest that crustal sources were very important in their petrogenesis. It is likely that anatexis of the surrounding crust resulted in the generation of these magmas.

1.6.4.2 *Geotectonic model*

The evolution and development of the Narsajuaq terrane represents one aspect in the broader picture of the Ungava Orogen as a whole. Other authors have documented the geotectonic evolution of most of the other tectonostratigraphic units within the Ungava Orogen, which involved the rifting of the Archean Superior Province basement, generation of an oceanic basin and its subsequent closure, and amalgamation of these units into the Cape Smith Thrust-Fold Belt (e.g., Francis et al. 1983; Picard et al. 1990; St-Onge and Lucas 1990; Lucas et al. 1992; St-Onge et al. 1992). The Narsajuaq terrane fits into this picture beginning with the period representing the closure of the oceanic basin.

St-Onge et al. (1992) originally proposed that the development of the Narsajuaq terrane magmatic arc involved northward subduction of the rocks of the previously formed oceanic basin (Povungnituk, Chukotat and Watts Groups) under an oceanic crust beginning at approximately 1.90 Ga (Fig. 42a). Subduction related magmatism was occurring at the same time as accretion of the Povungnituk and Chukotat Groups to the southern margin of the Superior Province, based on the age of a gabbroic sill in the Chukotat Group (1870 Ma) that contains Archean xenocrystic zircons that are interpreted as inherited from the Superior Province, implying that this unit was already thrust onto the adjacent Superior Province margin (Lucas and St-Onge 1992). The lack of plutonic rocks having an arc affinity in the underlying Superior Province basement and in the Povungnituk and Chukotat Groups suggests that these units (Povungnituk and Chukotat) were part of the lower subducting plate and may have already accreted to the continental margin, hence were not outboard of the trench system at this time (St-Onge et al. 1992). The results of this present study indicate that while the Cape Smith suite of plutons was emplaced in a predominatly mafic oceanic crust and involved relatively minor contamination via sediment subduction, the petrogenesis of the other magmatic suites involved a more significant older crustal component. We envisage a scenario in which

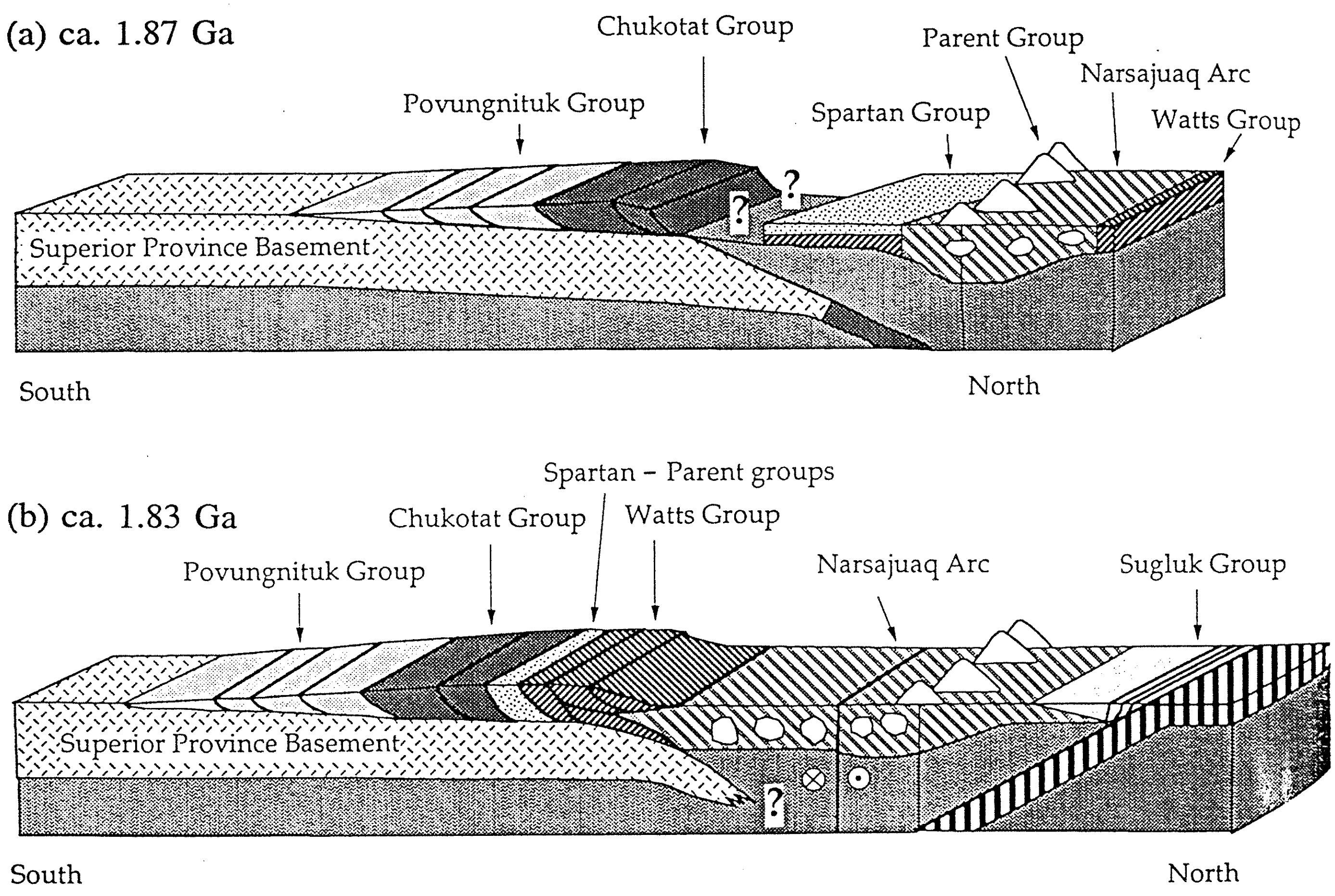


Fig. 1.42. Schematic block diagram from St-Onge et al. (1992) illustrating the possible geotectonic setting for the development of the Narsajuaq terrane. (a) at ca. 1.87 Ga; (b) at a. 1.83 Ga.

subduction was occurring (possibly at an oblique angle) under a crustal fragment that marked the ocean to continent transition across the previously rifted continental crust. In this model the oceanic crust of the Watts Group (and subsequently the emerging volcanic arc rocks of the Parent Group; see next section) was the foundation for the magmatic activity in one area (producing the Cape Smith suite) whereas the transition to the crustal fragment was the basement elsewhere, thereby accounting for the somewhat synchronous activity of the Cape Smith and Older suites as well as their different geochemical and isotopic signatures (Fig. 43a).

The short hiatus in the timing of the magmatic activity between the Older and Younger suites (from 1844 to 1836 Ma) may be related to a number of factors. Small changes in the angle of subduction of the down-going plate and its geometry have accounted for the temporal changes in arc magmatism in central Chile (Hildreth and Moorbath 1988). Attempted subduction of more buoyant fragments of crust such as continental fragments, oceanic islands or plateaus or even the relict spreading ridge may cause the system to become temporally jammed and may even incite a change in plate geometry (e.g., a flip in the direction of subduction occurred along the eastern Banda arc several million years ago following collision and attempted subduction of the Australian continental margin with the arc; Hamilton 1988). Although we cannot presently constrain which, if either, of these processes may have been important in the evolution of the Narsajuaq terrane, a break in the temporal sequence occurred. A different style of magmatism was the result, with plutonism now occurring at higher crustal levels. St-Onge et al. (1992) have proposed that subduction-polarity reversal occurred following the attempted subduction of the Superior Province continental lithosphere under the emerging arc (Fig. 42b). The ongoing collision between the Cape Smith Belt lithologies and the Superior Province may have started to affect the Narsajuaq terrane at this time, such that it may have been partially exhumed.

One of the oldest units emplaced after this event (1836-1834 Ma) was the distinct series of monzodiorite plutons which appear to have had a very distinct and different

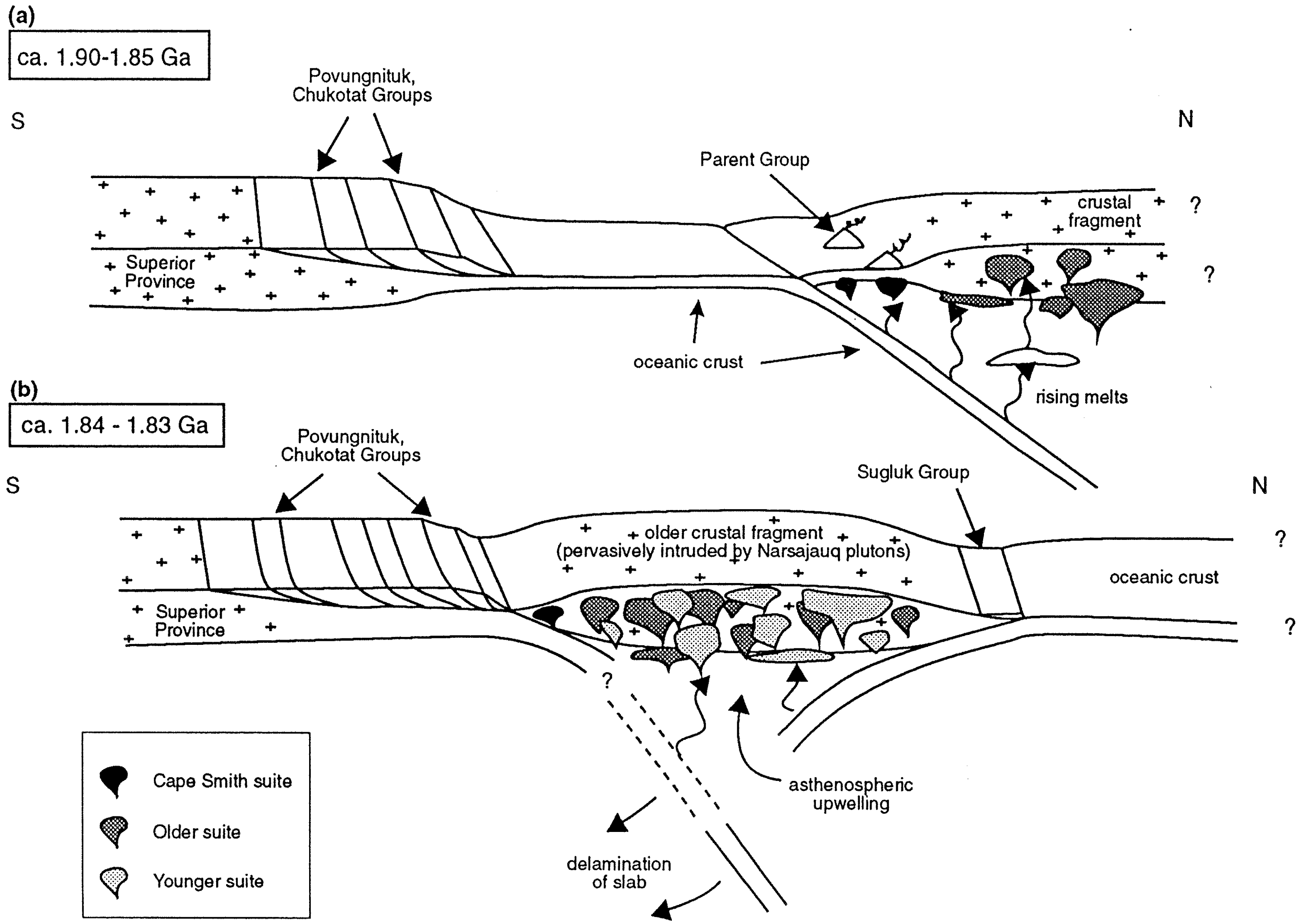


Fig. 1.43. Modified model for the origin of the Narsajuaq terrane involving a crustal fragment which may possibly have been the previously rifted Superior Province crust.

mantle source that was more LILE and light REE enriched compared to the source(s) for the other plutons. Heterogeneites in the mantle could account for the presence of such a distinct source, and an enriched mantle source has previously been postulated as the source of alkaline and other enriched magmas within the Cape Smith Belt (Picard et al. 1990; Gaonac'h et al. 1992, Scott et al. 1991; Dunphy et al. 1995). Another explaination for the presence of these magmas is related to the subduction-polarity reversal whereby delamination of the subducting slab and upwelling and adibatic melting of the subduction-enriched asthenospheric mantle occurred giving rise to the monzodiorites (Fig. 43b). At approximately the same time (1835-1821 Ma) diorite and tonalite plutons that lack the extreme LILE and light REE enrichment of the monzodiorites were intruded, indicating the presence of a separate and different mantle source. The geochemical and isotopic signature of these plutons is similar to the Older suite, and comparable sources and magmatic processes (fractionation, contamination and AFC) were likely involved. The Sugluk Group sediments are interpreted as having been deposited in a marginal basin located north of the emerging arc (bac-arc basin) where they were intruded by the Older and Younger suite plutons as well as tectonically emplaced along accretion-related faults (Fig. 43b).

Continuing deformation and oceanic basin closure occurred until final accretion of the magmatic arc terrane to the Superior Province basement, which is thought to have occurred after 1821 Ma, the age of the youngest arc-related pluton in the Narsajuaq terrane (St-Onge et al., 1992; Lucas et St-Onge, 1992; R. Parrish, personal communication 1994). Magmatic activity did not halt at this point, and large granitic sheets were subsequently intruded at ca. 1800 Ma. The generally low strain state indicates that deformation was waning at this time, although the CNF granitic sheet is cross-cut by an accretion-related fault (Fig. 3) indicating that accretion-related thrusting was not completely terminated at this point. The enriched geochemical and isotopic signature of these plutons, however, points to a predominantly crustal source; and they were most likely derived by anatexis of the surrounding continental crust which was related to crustal thickening and hydration during the accretion and retrogression events.

Final accretion and D_2 deformation is constrained to have terminated before 1758 Ma, the age of cross-cutting syenogranite pegmatite dykes that stitch together the entire orogen (Parrish 1989), although the structural orientation of these dykes indicate that they were emplaced during the regional-scale D_3 deformation event which folded all elements of the orogen about E-W axes (Lucas and St-Onge 1992). A subsequent and final deformation event (D_4) occurred which refolded the units about N-S axes producing the dome and basin interference pattern preserved in the units of the orogen. A small, undeformed granitic pluton which cross-cuts a D_4 fold constrains this deformation event to earlier than the pluton's emplacement, dated at 1742 Ma (Dunphy et al. 1991; in press).

1.6.5 Comparison to other magmatic arcs in the Trans-Hudson Orogen

The Narsajuaq terrane of the Ungava Orogen represents the plutonic roots of an Early Proterozoic magmatic arc complex that was accreted to the northern margin of the Archean Superior Province at ca. 1.82 Ga. Other segments of the Trans-Hudson Orogen also contain examples of Early Proterozoic plutonic arc complexes (Fig. 44), namely the Wathaman batholith of the western Trans-Hudson (Lewry et al. 1981; Fumerton et al. 1984; Meyer et al. 1992) and the De Pas batholith of the Labrador segment of the orogen (Martelain 1989; Wardle et al. 1990; van der Leeden 1990; Dunphy and Skulski 1995a). These batholiths have not been studied in great detail and only recently have some geochemical and geochronological data been accumulated. Their very occurrence, however, is important to note, and comparison of these batholiths with the Narsajuaq terrane may shed light on the processes of magma generation and continental growth in the Early Proterozoic. Current work by Dunphy as part of a separate project (Lithoprobe ECSOOT) is striving to better document the geology, geochronology and geochemistry of the De Pas batholith in an effort to increase our understanding of this area (Dunphy and Skulski 1995a,b). A more detailed and complete comparison of the arc magmatism in the Narsajuaq terrane and the De Pas batholith will be forthcoming following completion of the De Pas project.

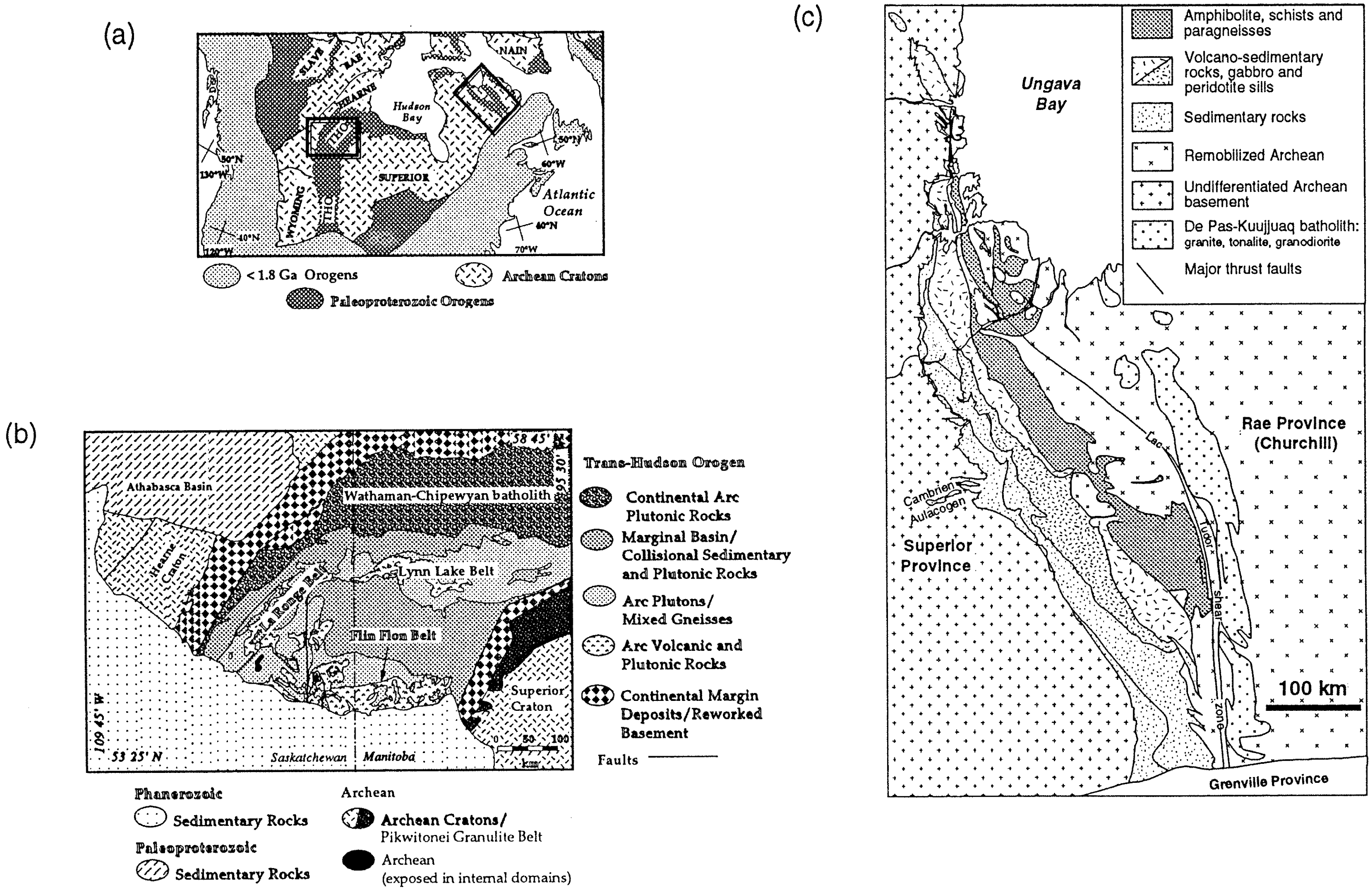


Fig. 1.44. Generalized geological maps for other segments of the Trans-Hudson Orogen: (b) western Manitoba/Saskatchewan segment; (c) Labrador segment. Both are located on the smaller regional scale map in (a). Maps modified from Skulski et al. (1993) and Lucas et al. (1994).

Other arc-related rocks of the Trans-Hudson Orogen are found within the Cape Smith Belt and in the western (Manitoba/Saskatchewan) segment of the Trans-Hudson Orogen (Flin Flon, La Ronge and Lynn Lake belts; Fig. 44b) which consist of volcanic, volcaniclastic and plutonic lithologies. We feel that investigation of the relationship between the Narsajuaq terrane and these other Early Proterozoic magmatic arcs is important as this may provide a more complete picture of magmatic arc processes and products during this time period in Earth's evolution. Comparisons have been hindered in the past due to the lack of easily available data, but projects such as the Geological Survey of Canada's Ungava Orogen study and the Lithoprobe - NATMAP project of the western Trans-Hudson Orogen have accumulated significant amounts of data which are presently being published. Future study of the various data sets should allow more detailed comparison; however, some preliminary comments may be made at this time.

1.6.5.1 *Plutonic arcs*

1.6.5.1.1 Wathaman batholith

The enormous Wathaman batholith (≈900 km long) is located in the Reindeer Zone of the western Trans-Hudson Orogen (Fig. 44b) and has been described as a relatively homogeneous granitoid body in terms of its field characteristics, lacking internal intrusive contacts which are typical of many other magmatic arc complexes (Fumerton et al. 1984). Limited geochemical data for the batholith suggest that it is not a single chemically coherent body and that it has a bulk composition similar to other Phanerozoic arc batholiths found along the western margin of North and South America (Meyer et al. 1992). In addition, U-Pb geochronology on various samples of the Wathaman batholith document a range in ages from approximately 1870 - 1845 Ma (Meyer et al. 1992) indicating multiple phases of magmatism. Meyer et al. (1992) have interpreted the geological and geochemical data (including isotopic data from Chauvel et al. 1987) for the batholith as indicating that it formed as a continental magmatic arc resulting from northward subduction of oceanic crust beneath a continental margin during the Early Proterozoic.

1.6.5.1.2 De Pas batholith

The De Pas batholith is located in the Labrador segment of the Trans-Hudson Orogen (Fig. 44c) and is a composite granitoid body extending over 600 km (Martelain 1989; van der Leeden et al. 1990; Wardle et al. 1990). Geological and geochronological studies have suggested that the De Pas batholith is a late- to post-tectonic intrusion based on its relative deformation state (compared to surrounding rocks; Wardle et al. 1990) and the available U-Pb ages for the batholith which range from 1840 to 1811 Ma (Bowring, cited in van der Leeden et al. 1990; Krogh 1986). Geochemical data on the De Pas batholith however, suggests that it has a magmatic arc affinity (Martelain 1989; van der Leeden et al. 1990). A preliminary model that can reconcile the different geological data suggests that the shape of the eastern Superior Province margin during the Early Proterozoic may have influenced the timing of collision and duration of arc magmatism along the length of the New Quebec Orogen (Wares and Skulski 1992). Collision between a promontory located along the NE margin of the Superior Province and the SE Rae block may have predated the demise of east-dipping subduction beneath the SE Rae Province that generated the De Pas batholith (Dunphy and Skulski 1995a).

1.6.5.1.3 Discussion

The overall compositional ranges of the plutons that form the Wathaman and De Pas batholiths are similar to those observed in the Narsajuaq terrane, although felsic intrusions are more common in the former whereas the Narsajuaq terrane has a bulk composition that is closer to tonalite. No plutonic units having a layered aspect similar to the Older diorite-tonalite suite of the Narsajuaq terrane have been reported for either the De Pas or Wathaman batholiths, which are dominated by megacrystic granite-granodiorite compositions (Fumerton et al. 1984; Martelain 1989; van der Leeden et al. 1990; Meyer et al. 1992). The current exposure level of the three terrains may also be different with the Wathaman batholith representing mesozonal levels at 7-17 km (Meyer et al. 1992) while the Narsajuaq terrane exposes depths of approximately 20-45 km. The exact crustal depths represented by the De Pas batholith is unknown; however, it appears to be exposed in oblique section from upper greenschist facies to granulite facies

(Martelain 1989; Clark et al. 1990). The differences in the geological features of the batholiths may be related to the different depths of exposure for the various complexes. The time span of magmatic activity for the batholiths was approximately 25-30 Ma, compared to almost 80 Ma of magmatism for the principal suites of the Narsajuaq terrane (1898-1821 Ma). The large difference in the time span of magmatism between the Narsajuaq terrane and the batholiths may partly be a function of better geochronological control in the Narsajuaq terrane (almost 30 dates have been produced) compared to the other studies, but it may also reflect a longer-lived magmatic arc.

Detailed comparison of the geochemical character of the Narsajuaq terrane plutons and the Wathaman batholith is precluded by the paucity of geochemical data for the latter, particularly for the more mafic compositions. The available trace element data for the Wathaman batholith overlap with most of the Narsajuaq terrane data, although in general a larger range in concentrations is noted for the latter. Nb concentrations show the largest differences between the two regions, with some samples from the Wathaman batholith having high values (≈31-45 ppm). Such elevated concentrations are rare in the Narsajuaq terrane samples. Meyer et al. (1992) suggest however, that the high Nb values in the Wathaman batholith may be a function of sample contamination during preparation for analysis. Published geochemical data is also sparse for the De Pas batholith, although it is currently the subject of study by the first author and data will be presented elsewhere. Preliminary studies on the De Pas batholith (Dunphy and Skulski 1995a) are consistent with previous models proposing a continental arc setting for its emplacement (Martelain 1989; van der Leeden et al. 1990). Samples from the De Pas batholith have slightly higher average trace element concentrations relative to the Narsajuaq terrane compositions which may reflect chemical differences in upper plate crustal compositions in the eastern Trans-Hudson orogenic system during the Early Proterozoic.

1.6.5.2 *Volcanic arcs*

1.6.5.2.1 Parent Group

Arc related volcanic rocks have been identified within the Ungava Orogen which consist of the volcaniclastic and volcanic lithologies of the Parent Group (Fig. 2; Picard et al. 1990; St-Onge and Lucas 1992; Barrette 1994). The volcanic rocks are dominated by andesitic basalt, although compositions from basalt to rhyolite are present, and occur both as massive and pillowed flows. The mafic to felsic volcanic and volcaniclastic rocks (crystal tuffs and lapilli tuffs) interfinger laterally with epiclastic sediments and occur as a fault-bounded package sandwiched between the Watts and Chukotat Groups in the western part of the Ungava Orogen (Picard et al. 1990; St-Onge and Lucas 1992; Barrette 1994). The lateral gradation of the Parent Group volcanics into sedimentary units was interpreted by St-Onge and Lucas (1992) as indicating an environment of subaerial deposition from discrete individual eruptive centres, accompanied by erosion and redeposition of the volcanic detritus.

Only a few averages of geochemical analyses have been published for the Parent Group (Picard et al. 1990) and have been interpreted as indicating a subduction-related origin for the rocks. One sample of a Parent Group rhyolite has been dated at 1860 ± 2 Ma, while a volcaniclastic rock contained several different zircon populations having ^{207}Pb/^{206}Pb ages of 1917, 1968 and 2423 Ma (Machado et al. 1993). Correlation of the Narsajuaq terrane plutons with the Parent Group volcanics/volcaniclastics was proposed by St-Onge et al. (1992) based on the similar arc-affinity and close temporal and spatial relationships of both units. The presence of some plutons from the Cape Smith suite of the Narsajuaq terrane in some parts of the Parent Group suggests that the correlations are justified, although a definitive relationship between the two units is difficult to establish, due to the lack of geochemical data for the Parent Group.

1.6.5.2.2 Flin Flon Belt

The relative paucity of data for the Parent Group contrasts with that for another region of the Trans-Hudson Orogen, that of the Flin Flon volcanic belt (Fig. 44b), which has recently been the focus of study by many researchers, resulting in a considerable body of information for this region. The Flin Flon Belt represents a collage of oceanic arcs, ocean floor basalts and oceanic plateau/oceanic island volcanics which have been juxtaposed in an accretionary complex at ca. 1.88-1.87 Ga (Lucas et al. 1995, and references therein). Although evidently not directly related to the Narsajuaq terrane plutonic arc, the Flin Flon arc volcanics may provide some insight into possible equivalent arc rocks in the Ungava Orogen. Detailed description and comparison of the volcanic arc rocks from the Flin Flon Belt and the Narsajuaq terrane arc is beyond the scope of this paper, however, some general comments can be made. The following discussion relies primarily on a series of recent papers that have synthesized previous information for the Flin Flon Belt and present new geological data (Lucas et al. 1994; Stern et al. 1995a, 1995b; Lucas et al. 1995, and references therein).

Arc volcanics of the Flin Flon Belt are dominated by isotopically juvenile volcanic assemblages of tholeiitic, calkaline and shoshonitic character which were erupted between 1904-1882 Ma, with the tholeiitic series extruded prior to the calc-alkaline and alkaline magmatism (Stern et al. 1995a). The geochemical composition of the tholeiitic volcanics (mainly basalt-basaltic andesite with lesser dacite and rhyolite) and some calc-alkaline volcanics is similar to modern island arc tholeiites. These rocks have been interpreted as derived from a highly refractory mantle source (more depleted than N-MORB; Stern et al. 1995a). Minor amounts of contamination involving recycled older crust through sediment subduction is postulated to account for the variation in Nd isotopic composition for these rocks (initial ε_{Nd} ranges from +4.8 to -0.4; Stern et al. 1995a). Other calc-alkaline rocks and the alkaline series have geochemical compositions that are distinct from the tholeiite series and a more fertile mantle source (N-MORB - OIB) is inferred for these rocks (Stern et al. 1995a). The geological field relationships and geochemical and geochronological data for these suites have led Stern et al. (1995a) to suggest that they were formed upon a thickened crust (the emerging arc), producing a more mature magmatic arc compared to the earlier juvenile suite. The Nd isotopic composition of these rocks does not indicate significant crustal contamination by older rocks except for one segment of the arc (Snow Lake) which is proposed to have been constructed on an Archean microcontinental fragment (Stern et al. 1995a). A very minor amount of Archean continental crust has been identified within the Flin Flon Belt (David and Syme 1995), occurring as fault bounded lozenges.

Arc-related plutonic rocks cut the Flin Flon volcanics and consist of younger (1866-1838 Ma) calc-alkaline diorite, tonalite and granodiorite plutons which intruded the previously amalgamated tectonostratigraphic assemblages (Lucas et al. 1995, and references therein). These plutons have juvenile Nd isotopic signatures suggesting little involvement of older crustal components (Lucas et al. 1995). Lucas et al. (1995) interpreted this later magmatic event as related to the development of a successor arc built upon the amalgamated collage of arc volcanics, ocean floor basalts and ocean plateau/island assemblages.

The most striking difference between the arc assemblages of the Flin Flon Belt and the Narsajuaq terrane is the overall juvenile nature of the former compared to the more evolved compositions of the latter. The aerial distribution of the Narsajuaq terrane rocks is dominated by intermediate compositions, while slightly more mafic compositions predominate in the Flin Flon Belt. The low- to medium-K compositions of the Flin Flon Belt contrast with the medium- to high-K plutons of the Narsajuaq terrane,, as do the proportions of tholeiitic versus calc-alkaline rocks, with the former being more abundant in the Flin Flon Belt but relatively rare in the Narsajuaq terrane. The Nd isotopic composition of the two regions is also significantly different, with a much larger range in initial ε_{Nd} values recorded in the rocks of the Narsajuaq terrane relative to the Flin Flon Belt arc assemblages. These differences are primarily related to the different geotectonic environments in which each segment was formed, with an oceanic arc setting for the Flin Flon Belt compared to a predominantly continental setting for the Narsajuaq terrane.

1.6.5 Crustal section

The excellent exposure of the rocks in the northern Ungava Peninsula has provided a superb opportunity to fully document the geological history and evolution of an Early Proterozoic orogenic belt (e.g., Francis et al. 1983; Picard et al. 1990; St-Onge and Lucas 1990; Dunphy et al. 1995). The present exposure of the Narsajuaq terrane is a consequence of its accretion to the Superior Province margin and subsequent deformation and exhumation which preserved the rocks of the Narsajuaq terrane in oblique section along the northern flank of a regional-scale antiform cored by the Archean Superior Province basement units (Lucas and St-Onge, 1995). The Narsajuaq terrane therefore presents a unique cross-section through the crust of an Early Proterozoic magmatic arc. Detailed discussion of the composition, structure and evolution of the Narsajuaq terrane section relative to other crustal sections of varying ages will be the subject of a future paper, however, preliminary comments may be made at this time.

Several examples of crustal sections have been documented at diverse localities around the world representing different time periods in Earth's evolution (e.g., the Archean Kapuskasing Zone (Percival and Card 1983) and Pikwitonei Belt (Fountain and Salisbury 1981), the Paleozoic Ivrea Zone (Berckhemer 1969; Pin 1990) and the Mesozoic Kohistan complex (Bard 1983)). Such sections generally contain various lithological components indicative of different crustal levels (i.e., sediments and volcanics representing upper crust and mafic/ultramafic bodies from the lower crust or possibly part of the upper mantle) although the most distinctive zonation within crustal sections is the changing metamorphic gradient with depth (Fountain and Salisbury 1981). Upper crustal rocks are generally at greenschist facies or even unmetamorphosed, although often these units are not preserved due to erosion. Increasing metamorphic grade to granulite facies occurs deeper in the crust, and the base of the crustal section is usually marked by a major fault zone or suture along which the section was emplaced (Fountain and Salisbury 1981). Geophysical tools, such as seismic reflection and refraction and gravity measurements, have played an important role in imagining the deeper levels of crustal sections, allowing extrapolation of the surface structures to deeper levels (e.g., Percival et al. 1992 and references therein).

Based on lithological, metamorphic and structural arguments, the present erosional surface of the Narsajuaq terrane can be considered as a mid-crustal section of Early Proterozoic age. Metamorphic P-T studies (see earlier section on metamorphism) have suggested that the Narsajuaq terrane is divided into two tectonic blocks that have repeated the crustal section (Monday 1994). Pressure determinations based on the metamorphic data indicate crustal depths ranging from approximately 22 to 45 km for the northern block and 25 to 38 km for the southern block (Monday 1994). These measurements indicate that the present erosion surface of the Narsajuaq terrane exposes mid-crustal levels of a magmatic arc complex, with ≈ 15-25 km of structural section. Juxtaposition of higher grade rocks and lower grade rocks occurs not only along the discontinuity separating the northern and southern blocks but also along the main terrane-bounding fault which places the Narsajuaq terrane rocks next to the lower greenschist to middle amphibolite facies rocks of the Watts Group (Lucas and St-Onge 1992).

The geological data, including down-plunge strike-perpendicular structural sections (Lucas and St-Onge 1995), has been used to construct a schematic cross-section of the Narsajuaq terrane crust (Fig. 45). The section consists of two stacked columns separated by the proposed fault along its mid-section (Monday 1994) representing the northern and southern blocks, with the lowermost contact of the southern block defined by the major terrane-bounding fault. In the Narsajuaq terrane the main "base unit" is the layered diorite-tonalite gneiss of the Older suite, which is volumetrically dominated by intermediate compositions. Lithological variations with depth in the section are evident, with metasedimentary rocks of the Sugluk Group only occurring near the top of the preserved sections while significant mafic (dioritic) intrusions are predominantly found near the base. Such mafic intrusions are almost exclusively located in the southern block and are remarkably absent from the northern block. Conversely, the northern

NARSAJUAQ TERRANE

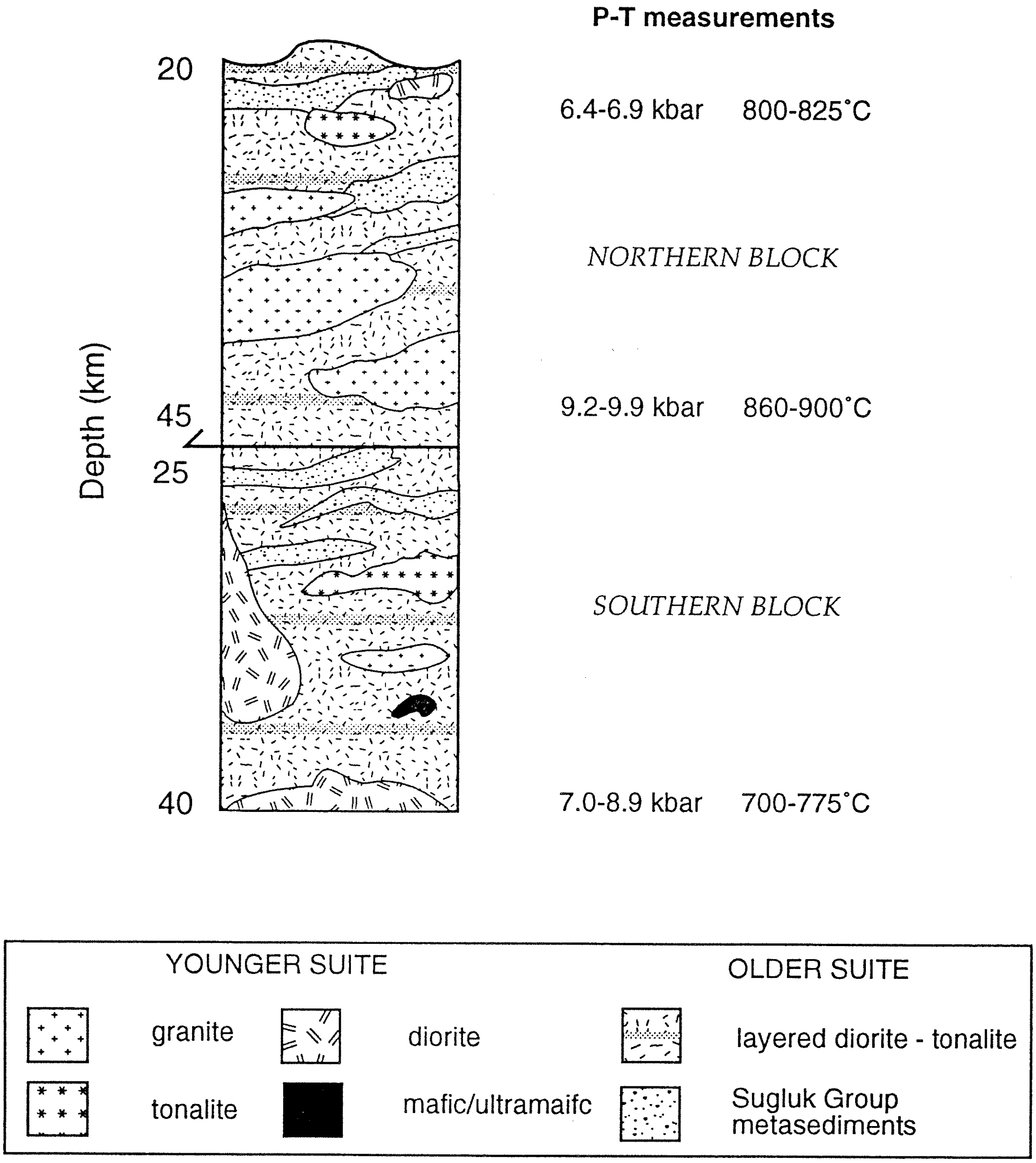


Fig. 1.45. Schematic crustal section through the Narsajuaq terrane. P-T ranges from Monday (1994).

block is dominated by huge granitic sheets which are more scarce to the south. The uppermost crust in most other crustal sections is dominated by metavolcanic and metasedimentary rocks at greenschist facies conditions. Volcanic rocks that may have once been directly associated with the plutonic units of the Narsajuaq terrane are found in the Parent Group (Fig. 2) which is presently preserved as imbricated slices south of the Narsajuaq terrane in the Cape Smith Belt. The Parent Group has not yet been studied in detail, and as such, a decisive link has yet to be established between the two units. Future geochemical and petrological investigations will help understand the relationship between the volcanics/volcaniclastics of the Parent Group and the plutonics of the Narsajuaq terrane. The Sugluk Group represents upper crust metasediments, although they are presently preserved at granulite facies, some 20-25 km deep in the crust, as a result of subduction and tectonic accretion to the arc terrane.

Crustal sections from the Ivrea Zone and Pikwitonei Belt (Fountain and Salisbury 1981) and the Kapuskasing Zone (Percival 1990) are also presented for comparative purposes (Fig. 46), and the reader is referred to Fountain and Salisbury (1981) for additional sections. Vertical scales were not given for all the sections, however, the relative position of the amphibolite- to granulite-facies transition is noted. The entire Narsajuaq terrane section lies within the granulite facies field. A considerable amount of compositional variation is displayed by the different crustal sections, independent of the metamorphic grade. The layered nature of the lower crust in the Narsajuaq terrane is very similar to that reported for the Kapsukasing Zone (Percival and Card 1983; Percival and McGrath 1986) whereas the other sections appear to consist of more homogenous packages of mafic or intermediate gneisses. It should be noted that Fountain and Salisbury (1981) separated amphibolite-facies quartzofeldspathic gneisses from the granulite-grade intermediate gneisses, however, we have represented them here as one single unit, as they are essentially of the same bulk composition. Large mafic/ultramafic bodies which have been noted at lower crustal levels in several sections are not present in the Narsajuaq terrane section, although the large diorite intrusions near the base of the southern block of the Narsajuaq terrane may be equivalent to mafic

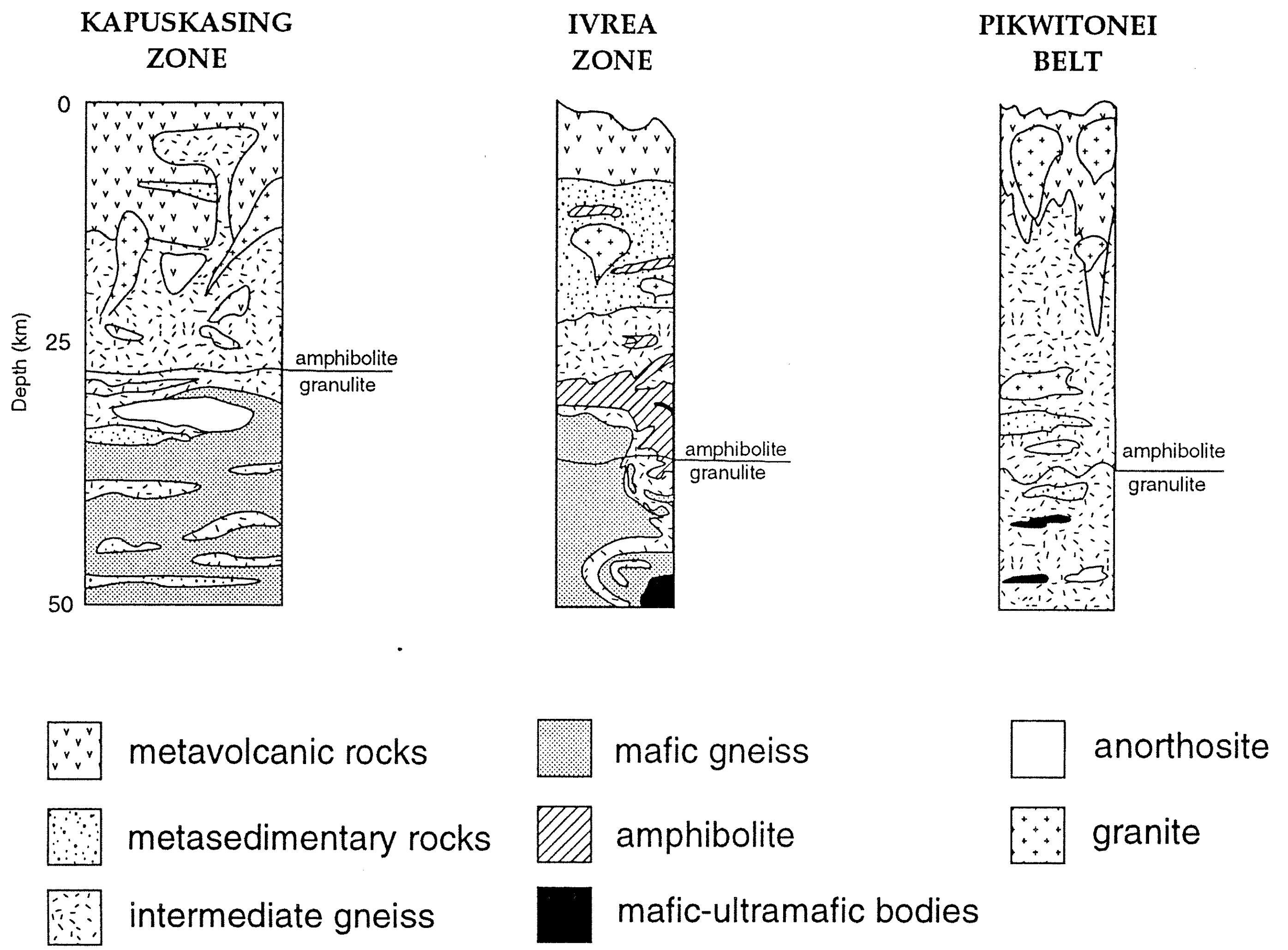


Fig. 1.46. Schematic crustal sections for various regions (Archean Kapuskasing Zone (Percival 1990), Ivrea Zone and Pikwitonei Belt (Fountain and Salisbury 1981)) for comparison with the Narsajuaq terrane section.

bodies from other sections. Large anorthosite bodies have been mapped in the Kapuskasing Zone, and smaller ones have been noted in the Pikwitonei Belt, but are absent from the Narsajuaq terrane. The uppermost crust consists of metavolcanic rocks in all sections except the Narsajuaq terrane although, as previously mentioned, the Parent Group may represent such crust. While metasedimentary rocks do occur at upper crustal levels in some sections, they are not restricted to such shallow depths and are present at depth in the Kapuskasing Zone and Pikwitonei Belt. The metasedimentary rocks of the Narsajuaq terrane occur at similar depths in the section, approximately 20-25 km. Granitic intrusions in the comparative sections (and all other sections in Fountain and Salisbury 1981) are found primarily at upper crustal levels associated with either metavolcanics or metasediments. The huge granitic sheets of the Narsajuaq terrane section occur at greater depths in the northern block but are rare in the southern block.

In addition to the lithological characterization of the Narsajuaq terrane crustal section, a geochemical composite composition for the terrane is desired for the purposes of comparison with other crustal compositions through time. Previous studies have documented the existence of geochemical differences between Archean and post-Archean rocks (e.g., Taylor and McLennan 1985; Martin 1986; Drummond and Defant 1990; Condie 1992) reflected in decreasing La_n/Yb_n ratios and the development of negative Eu anomalies, for example. Changes in the style of melting (slab versus wedge) have been invoked to account for some of the geochemical differences observed in arc volcanics of differing ages (e.g., Drummond and Defant 1990), although other factors such as the amount and rate of crust/mantle recycling and the secular cooling of the earth may also be important in accounting for the overall bulk compositional changes with time (Condie 1992). Comparison of the geochemical signature of the Early Proterozoic Narsajuaq terrane with Archean and Phanerozoic granitic suites is currently in a preliminary stage, but certain trends have been delimited such as decreasing ratios of La_n/Yb_n, Sr/Y, Th/U from the Archean through the Proterozoic to the Phanerozoic. The relative amount of fractionation of the REE's was noted to be similar in the intermediate composition Narsajuaq terrane samples and similar compositions of the Phanerozoic suite (see earlier section on comparision with magmatic arcs and Fig. 32c), suggesting a similar melting regime in which garnet did not play a significant role. If the melt generating conditions were the same for the Proterozoic Narsajuaq terrane samples and the Younger Phanerozoic samples, then some other factor(s) must be invoked to explain the notable geochemical differences between the suites, particuarily for Th and U. Differentiation, fractionation and other upper crustal processes from similar sources would produce magmas having comparable overall trace element abundances, although contamination by different crustal compositions could account for some differences. The most viable means to account for the geochemical differences is through variations in the source composition through time, which would result in significant geochemical differences in the derivative melts.

Detailed study of the geochemical composition of the Narsajuaq terrane may not answer all questions about the geochemical differences and the processes that generated them. However, the data will contribute to our understanding of crustal evolution, particularly during the Early Proterozoic as such information is currently sparse. While we have not answered all questions pertaining to the differnces in crustal composition through time, ongoing work is striving to meet these goals and will help constrain the various mechanisms responsible for such changes.

1.7. CONCLUSIONS

The Narsajuaq terrane represents the roots of an Early Proterozoic magmatic arc complex that was accreted to the margin of the Superior Province ca. 1.82 Ga. The terrane consists of four different plutonic suites spanning more than 150 Ma of magmatic activity, as well as a metasedimentary unit (Sugluk Group).

The oldest plutons in the Narsajuaq terrane are found in the Cape Smith suite which contains intrusions representing almost 60 Ma of plutonic activity, from 1898-1839 Ma. The Cape Smith suite plutons have a wide range in compostion (diorite-tonalite-granite) which primarily have a calc-alkaline major element signature, although a few samples are tholeiitic. Slight to moderate enrichment of the LILE's and light REE's are noted in these plutons, as well as small negative Nb and Ti anomalies. The isotopic signature of this suite is fairly restricted, with initial ε_{Nd} = +3.2 to +1.5 and initial $^{87}Sr/^{86}Sr$ ratios of 0.7020 - 0.7024. Fractionation of a slightly LILE- and light REE-enriched mantle source accompanied by minor amounts of subduction-related contamination by sediments produced the plutons of this suite, which were intruded into the predominantly mafic oceanic crust of the Watts, Parent and Spartan Groups.

The principal suite of the Narsajuaq terrane is the Older suite which consists of a well layered sequence of diorite-tonalite-granite having a higher and more variable concentration of the LILE's and light REE's relative to the Cape Smith suite. The isotopic composition of the Older suite plutons is also more variable than the Cape Smith suite, with initial ε_{Nd} values ranging from +4.0 to -10.7 and initial $^{87}Sr/^{86}Sr$ ratios of 0.7025 - 0.7048. The Older suite was intruded over a period of almost 20 Ma, from 1863-1844 Ma, and overlaps temporally with the Cape Smith suite. It is postulated that the Older suite was constructed on an Archean crustal fragment which interacted with the rising magmas to produce the range in geochemical and isotopic signatures measured in these rocks.

Intrusive into the Older suite is a series of discrete, more homogenous plutons (relative to the layered aspect of the Older suite), tnes of km in size, which make up the Younger suite. These plutons were intruded primarily over 15 Ma, from 1836-1821 Ma, with a series of Younger plutons emplaced as huge granitic sheets at 1803-1800 Ma. The Younger suite diorite, tonalite and granite plutons have geochemical and isotopic compositions that are similar to the Older suite, although a slightly larger range in initial ε_{Nd} values is recorded (+3.9 to -15.6). These rocks appear to have had a similar source, petrogenesis and evolution as the Older suite. A distinct series of plutons, monzodioritic in composition, are almost exclusively found in the Younger suite and contain very high concentrations of the LILE's and light REE's, as well as having very fractionated REE patterns. Partial melting of an enriched mantle source can account for the geochemical signature of these rocks. We suggest that melting occurred as a result of delamination of the subducting slab and upwelling of the subduction-modified asthenospheric mantle after a flip in the subduction zone geometry which occurred between 1844-1836 Ma. The youngest plutons of the Younger suite are the large CNF and IVUJ granitic sheets that contain variable but high concentrations of the LILE's and light REE's and have isotopic compositions indicative of significant crustal input (ε_{Nd} = -4.9 to -6.2 (IVUJ) and -17.5 to -18.5 (CNF)). These Younger suite granites were emplaced during the latter stages of accretion of the magmatic arc to the Superior Province margin and were generated by partial melting of the crust due to tectonic thickening, accompanied by fluid influx related to metamorphic retrogression of the crust at this time.

The Late suite consists of a series of undeformed and unmetamorphosed syenogranite pegmatite dykes that cross-cut all units of the orogen and small, rare granitic plutons emplaced primarily in the Narsajauq terrane, although one pluton was found to cross-cut all tectonostratigraphic units. The pegmatite dykes were emplaced at 1758 Ma, while the granitic plutons are slightly younger, with one pluton dated at 1742 Ma. The geochemical, isotopic and geochronological data for the plutons of this suite indicate that crustal anatexis was the principal mechanism by which they were generated during the final stages of deformation within the Ungava Orogen.

The geological evolution of the Narsajuaq terrane and the Ungava Orogen are similar to those observed in modern settings such as along the Banda-Sunda arc or the western margin of North and South America and indicate a continuity in magmatic and tectonic processes through time from at least the Early Proterozoic. The Narsajuaq terrane is a mid-crustal section through an Early Proterozoic magmatic arc that presents a unique opportunity to study crustal evolution through comparison with other crustal sections of Archean and Phanerozoic age. Geochemical differences between Archean and younger granitic suites, such as lower $(La/Yb)_n$ ratios in the latter, have been noted, while the Proterozoic Narsajuaq terrane contains overall lower Th and U concentrations compared to Younger (Phanerozoic) magmatic arc suites. While these differences have not been fully resolved, the geological and geochemical characterization of the Narsajuaq terrane presented here provide the basis for such future studies.

REFERENCES

Anderson, J.L., (*Editor*). 1990. The Nature and origin of Cordillerian magmatism. Geological Society of America Memoir **174** pp. 414.

Armstrong, R.L. 1968. A model for Sr and Pb isotope evolution in a dynamic Earth. Reviews of Geophysics, **6**:175-199.

Armstrong, R.L. 1991. The persistent myth of crustal growth. Australian Journal of Earth Sciences, **38**:613-630.

Armstrong, R.L., vs., and DePaolo, D.J. 1981. Comment and reply on Crustal growth and mantle evolution: Inferences from models of element transport and Nd and Sr isotopes. Geochimica Cosmochimica et Acta, **45**:1251-1254.

Baldwin, J.A., and Pearce, J.A. 1982. Discrimination of productive and nonproductive porphyritic intrusions in the Chilean Andes. Economic Geology, **77**:664-674.

Bard, J.P. 1983. Metamorphism of an obducted island arc: example of the Kohistan sequence (Pakistan) in the Himalayan collided range. Earth and Planetary Science Letters, **65**:133-144.

Barker, F., and Arth, J.G. 1984. Preliminary results, Central Gneiss Complex of the Coast Range batholith, southeastern Alaska: The roots of a high-K calc-alkaline arc? Physics of the Earth and Planetary Interiors, **35**:191-198.

Barker, F., Arth, J.G., and Stern, T.W. 1986. Evolution of the Coast batholith along the Skagway Traverse, Alaska and British Columbia. American Mineralogist, **71**:632-643.

Barrette, P.D. 1994. Lithostratigraphy and map-scale structure in the western Cape Smith Belt, northern Quebec: a tentative correlation between two tectonic domains. Canadian Journal of Earth Sciences, **31**:986-994.

Bateman, P.C. 1983. A summary of critical relations in the central part of the Sierra Nevada batholith, California, U.S.A. In Circum Pacific Plutonic Terranes. Edited by J.A. Roddick. Geological Society of America, Memoir **159**, pp. 241-254.

Bateman, P.C., and Chappell, B.W. 1979. Crystallization, fractionation, and solidification of the Tuolumne Intrusive Series, Yosemite National Park, California. Geological Society of America Bulletin, **90**:465-482.

Beach, A. 1980. Retrogressive metamorphic processes in shear zones with special reference to the Lewisian complex. Journal of Structural Geology, **2**:257-263.

Bégin, N.J. 1992. Contrasting mineral isograd sequences in metabasites of the Cape Smith Belt, northern Quebec, Canada: three new bathograds for mafic rocks. Journal of Metamorphic Geology, **10**:685-704.

Berckhemer, H. 1969. Direct evidence for the composition of the lower crust and the moho. Tectonophysics, **8**:97-105.

Brown, G.C. 1982. Calc-alkaline intrusive rocks: their diversity, evolution and relation to volcanic arcs. *In*: Andesites. *Edited by* R.S. Thorpe. John Wiley and Sons, New York, pp. 437-464.

Brown, G.C., Thorpe, R.S., and Webb, P.C. 1984. The geochemical characteristics of granitoids in contrasting arcs and comments on magma sources. Journal of the Geological Society of London, **141**:413-426.

Chauvel, C., Arndt, N.T., Kielinzcuk, S., and Thom, A. 1987. Formation of Canadian 1.9 Ga old continental crust. I: Nd isotopic data. Canadian Journal of Earth Sciences, **24**:396-406.

Clark, T., Marcoux, P., Bélanger, M., Wilson, C., and Avramtchev, L. 1990. Carte des gites mineraux du Quebec. Region de la Fosse du Labrador. Ministère de l'Energie et des Ressources du Québec, DV 84-01.

Condie, K.C. 1986. Origin and early growth rate of continents. Precambrian Research, **32**:261-278.

Condie, K.C. 1990. Growth and accretion of continental crust: Inferences based on Laurentia. Chemical Geology, **83**:183-194.

Condie, K.C. 1992. Evolutionary changes at the Archean-Proterozoic boundary. *In*: The Archean: Terrains, Processes and Metallogeny. *Edited by* J.E. Glover and S.E. Ho. Geology Key Centre & University Extension, University of Western

Australia, Perth, pp. 177-189.

Coney, P.J., Jones, D.L., and Monger, J.W. 1980. Cordilleran suspect terranes. Nature, **288**:329-333.

David, J., and Syme, E.C. 1995. U-Pb geochronology of late Neoarchean tonalites in the Flin Flon Belt, Trans-Hudson Orogen: Surprise at the surface. Canadian Journal of Earth Sciences, in press.

Davidson, J.P. 1987. Crustal contamination versus subduction zone enrichment: Examples from the Lesser Antilles and implications for mantle source compositions of island arc volcanic rocks. Geochimica Cosmochimica et Acta, **51**:2185-2198.

DePaolo, D.J. 1980. Sources of continental crust: Neodymium isotope evidence from the Sierra Nevada and Peninsular Ranges. Science, **209**:684-687.

DePaolo, D.J. 1981. Trace element and isotopic effects of combined wallrock assimilation and fractional crystallization. Earth and Planetary Science Letters, **53**:189-202.

Drummond, M.S., and Defant, M.J. 1990. A model for trondhjemite-tonalite-dacite genesis and crustal growth via slab melting: Archean to modern comparisons. Journal of Geophysical Research, **95**:21503-21521.

Dunphy, J. 1994. Evolution des roches plutoniques du domaine nord de la Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, MB 94-58.

Dunphy, J.M., and Skulski, T. 1995a. Early Proterozoic granitic magmatism in the Ungava and New Quebec Orogens: the Narsajuaq terrane and the De Pas batholith. *In*: Eastern Canadian Shield Onshore-Offshore Transect (ECSOOT) Meeting Report (Nov. 28-29, 1994). Wardle, R.J. and Hall, J. (*compilers*), LITHOPROBE Report 45, p. 37-50.

Dunphy, J.M., and Skulski, T. 1995b. The Early Proterozoic De Pas batholith, New Quebec Orogen, Canada: an Andean arc or a continental collision massif? Precambrian 95 meeting, program with abstracts, p. 245.

Dunphy, J.M., Ludden, J.N., and Francis, D. 1995. Geochemistry of mafic magmas from the Ungava orogen, Quebec, Canada and implications for mantle reservoir compositions at 2.0 Ga. Chemical Geology, **120**:361-380.

Dunphy, J.M., Ludden, J.N., and Parrish, R.R. in press. Stitching together the Ungava orogen, northern Quebec: geochemical, isotopic and geochronological constraints of late magmatic events. Canadian Journal of Earth Sciences, vol. 32.

Dunphy, J.M., Parrish, R.R., and Ludden, J.N. 1991. Stitching together the Ungava orogeny: the geochemistry and age relationships of the Lac Duquet pluton. Geological Association of Canada - Mineralogical Association of Canada, Program with Abstracts, **16**: A32.

Ellam, R.M., and Hawkesworth, C.J. 1988. Elemental and isotopic variations in subduction related basalts: evidence for a three component model. Contributions to Mineralogy and Petrology, **98**:72-80.

Elthon, D. 1989. Pressure origin of primary mid-ocean ridge basalts. In Magmatism in the Ocean Basins. Edited by A.D. Saunders and M.J. Norry. Geological Society, Special Publication 42, pp. 125-136.

Faure, G. 1986. Principles of Isotope Geology. Second edition. John Wiley and Sons, New York, pp. 589.

Fountain, D.M., and Salsibury, M.H. 1981. Exposed cross-sections through the continental crust: implications for crustal structure, petrology, and evolution. Earth and Planetary Science Letters, **56**:263-277.

Francis, D.M., Ludden, J.N., and Hynes, A.J. 1983. Magma evolution in a Proterozoic rifting environment. Journal of Petrology, **24**:556-582.

Fumerton, S.L., Stauffer, M.R., and Lewry, J.F. 1984. The Wathaman batholith: largest known Precambrian pluton. Canadian Journal of Earth Sciences, **21**:1082-1097.

Gaonac'h, H., Picard, C., Ludden, J.N., and Francis, D.M. 1989. Alkaline rocks from a Proterozoic volcanic island in the Cape Smith Thrust Belt, New Quebec. Geoscience Canada, **16**:137-139.

Gaonac'h, H., Ludden, J.N., Picard, C., and Francis, D.M. 1992. Highly alkaline lavas in a Proterozoic rift zone: Implications for Precambrian mantle metasomatic processes. Geology, **20**:247-250.

Gill, J. 1981. Orogenic andesites and plate tectonics. Springer-Verlag, New York, pp.

390.

Gray, C.M. 1977. The geochemistry of Central Australian granulites in relation to the chemical and isotopic effects of granulite facies metamorphism. Contributions to Mineralogy and Petrology, **65**:79-89.

Gromet, L.P., and Silver, L.T. 1987. REE variations across the Peninsular Ranges Batholith: implications for batholithic petrogenesis and crustal growth in magmatic arcs. Journal of Petrology, **28**:75-125.

Hamilton, W.B. 1988. Plate tectonics and island arcs. Geological Society of America Bulletin, **100**:1503-1527.

Harmon, R.S., and Rapela, C.W., (*Editors*). 1991. Andean Magmatism and its Tectonic Setting. Geological Society of America Special Paper 265 pp. 309.

Harrison, S.M., and Piercy, B.A. 1990. The evolution of the Antarctic Peninsular magmatic arc; evidence from northwestern Palmer Land. *In*: Plutonism from Antarctica to Alaska. *Edited by* S. Mahlburg Kay and C. Rapela. Geological Society of America, Special Paper 241, pp. 9-25.

Hart, S.R., and Brooks, C. 1977. The geochemistry and evolution of Early Precambrian mantle. Contributions to Mineralogy and Petrology, **61**:109-128.

Harvey, D. 1995. Géochime et traceur isotopique du Nd dans les intrusifs Archéens de la parite Nord-est de la Péninsule d'Ungava. M.Sc. thesis, Université de Montréal, Montreal, Quebec.

Hawkesworth, C.J., Gallagher, K., Hergt, J.M., and McDermott, F. 1994. Destructive plate margin magmatism: Geochemistry and melt generation. Lithos, **33**:169-188.

Hegner, E., and Bevier, M.L. 1991. Nd and Pb isotopic constraints on the origin of the Purtuniq ophiolite and Early Proterozoic Cape Smith Belt, northern Quebec, Canada. Chemical Geology, **91**:357-371.

Hildreth, W., and Moorbath, S. 1988. Crustal contributions to arc magmatism in the Andes of central Chile. Contributions to Mineralogy and Petrology, **98**:455-489.

Hoffman, P.F. 1985. Is the Cape Smith Belt (northern Quebec) a klippe? Canadian Journal of Earth Sciences, **22**:1361-1369.

Hoffman, P.F., 1988. United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annual Reviews of Earth and Planetary Science, **16**:543-603.

Hoffman, P.F. 1990. Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen. *In*: The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 15-39.

Huppert, H., and Sparks, R.S.J. 1988. The generation of granitic magmas by intrusion of basalt into continental crust. Journal of Petrology, **229**:599-624.

Hynes, A.J., and Francis, D.M. 1982. A transect of the Early Proterozoic Cape Smith foldbelt, New Quebec. Tectonophysics, **88**:23-59.

Irving, T.N., and Baragar, W.R.A. 1971. A guide to the chemical classification of the common volcanic rocks. Canadian Journal of Earth Sciences, **8**:523-548.

Johnston, A.D. 1986. Anhydrous P-T phase relations of near-primary high alumina basalt from the South Sandwich Islands. Contributions to Mineralogy and Petrology, **92**:368-382.

Kesler, S.E., Jones, L.M., and Walker, R.L. 1975. Intrusive rocks associated with porphyry copper mineralization in island arc areas. Economic Geology, **70**:515-526.

Krogh, T.E. 1986. Report to the Newfoundland Department of Mines and Energy on isotopic dating results from the 1985-1986 geological research agreement. Newfoundland Department of Mines and Energy, Open File Report, LAB 707, pp. 15.

Lamothe, D., Picard, C., and Moorhead, J. 1983. Région du lac Beauparlant, Bande de Cap Smith - Maricourt. Ministère de l'Energie et des Ressources du Québec, DP 84-39.

Le Bel, L., Cocherie, A., Baubron, J.-C., Fouillac, A.M., and Hawkesworth, C.J. 1985. A high-K, mantle derived plutonic suite from "Linga", near Arequipa (Peru). Journal of Petrology, **26**:124-148.

Lewry, J.F., and Collerson, K.D. 1990. The Trans-Hudson Orogen: extent, subdivision and problems. *In*: The Early Proterozoic Trans-Hudson Orogen of North

America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 1-14.

Lewry, J.F., Stauffer, M.R., and Fumerton, S.L. 1981. A Cordilleran-type batholitic belt in the Churchill Province in northern Saskatchewan. Precambrian Research, **14**:277-313.

Lopez-Escobar, L., Frey, F.A., and Oyarzun, J. 1979. Geochemical characteristics of central Chile (33-34 S) granitiods. Contributions to Mineralogy and Petrology, **70**:439-450.

Lucas, S.B. 1989. Structural evolution of the Cape Smith Thrust Belt and the role of out-of-sequence faulting in the thickening of mountain belts. Tectonics, **8**:655-676.

Lucas, S.B., and St-Onge, M.R. 1991. Evolution of Archean and Early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec. In Current Research, Part C. Geological Survey of Canada, Paper 91-1C, pp. 109-119.

Lucas, S.B., and St-Onge, M.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 2: Structural and metamorphic history. Canadian Journal of Earth Sciences, **29**:765-782.

Lucas, S.B., and St-Onge, M.R. 1995. Syn-tectonic magmatism and the development of compositional layering, Ungava orogen (northern Quebec, Canada). Journal of Structural Geology, **17**:475-491.

Lucas, S.B., Stern, R.A., and Syme, E.C. 1995. Flin Flon greenstone belt: Intraoceanic tectonics and the development of continental crust (1.92-1.84 Ga). Geological Society of America Bulletin, in press.

Lucas, S.B., St-Onge, M.R., Parrish, R.R., and Dunphy, J.M. 1992. Long-lived continent-ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. Geology, **20**:113-116.

Lucas, S.B., White, D., Hajnal, Z., Lewry, J., Green, A., Clowes, R., Zwanzig, H., Ashton, K., Schledewitz, D., Stauffer, M., Norman, A., Williams, P.F., and Spence, G. 1994. Three-dimensional collisional structure of the Trans-Hudson Orogen, Canada. Tectonophysics, **232**:161-178.

Lucassen, F., and Franz, G. 1992. Generation and metamorphism of new crust in magmatic arcs: a case study from northern Chile. Terra Nova, **4**:41-52.

Machado, N., David, J., Scott, D.J., Lamothe, D., Philippe, S., and Gariépy, C. 1993. U-Pb geochronology of the western Cape Smith Belt, Canada: New insights on the age of initial rifting and arc magmatism. Precambrian Research, **63**:211-223.

Mahlburg Kay, S., and Rapela, C.W., (*Editors*). 1990. Plutonism from Antarctica to Alaska. Geological Society of America Special Paper 241 pp. 263.

Mahlburg Kay, S., Kay, R.W., Citron, G.P., and Perfit, M.R. 1990. Calc-alkaline plutonism in the intra-oceanic Aleutian arc, Alaska. *In*: Plutonism from Antarctica to Alaska. *Edited by* S. Mahlburg Kay and C.W. Rapela. Geological Society of America, Special Paper 241, pp. 233-255.

Martelain, J. 1989. Le batholith de De Pas dans le segment oriental de l'orogene Trans-Hudsonien: Typologie et implications geodynamiques. PhD thesis, Université Claude Bernard, Lyon, France.

Martin, H. 1986. Effect of steeper Archean geothermal gradient on geochemistry of subduction-zone magmas. Geology, **14**:753-756.

Mason, D.R., and McDonald, J.A. 1978. Intrusive rock and porphyry copper occurrencess of the Papua New Guinea-Solomon Islands region: a reconnaissance study. Economic Geology, **73**:857-877.

McCulloch, M.T., and Gamble, J.A. 1991. Geochemical and geodynamical constraints on subduction zone magmatism. Earth and Planetary Science Letters, **102**:358-374.

McLennan, S.M. 1988. Recycling of the continental crust. Pure and Applied Geophysics, **128**:683-724.

Meyer, M.T., Bickford, M.E., and Lewry, J.F. 1992. The Wathaman batholith: An Early Proterozoic continental arc in the Trans-Hudson orogenic belt, Canada. Geological Society of America Bulletin, **104**:1073-1085.

Monday, P. 1994. Narsajuaq terrane, Ungava Orogen: Thermobarometry of granulite-facies metasediments and metaigneous rocks and tectonic implications. M.Sc. thesis, Queens University, Kingston, Ontario.

Myres, J.S., 1990. Precambrian tectonic evolution of part of Gondwana, southwestern Australia. Geology, **18**:537-540.

Norrish, K., and Hutton, J.T. 1969. An accurate X-ray spectrographic method for the analysis of a wide range of geological samples. Geochimica Cosmochimica et Acta, **33**:431-453.

Nye, C.J., and Reid, M.R. 1986. Geochemistry of primary and least fractionated lavas from Okmok volcano, central Aleutians: Implications for arc magmagenesis. Journal of Geophysical Research, **91**:10271-10287.

Park, A.F., 1991. Continental growth by accretion: A tectonostratigraphic terrane analysis of the evolution of the western and central Baltic Shield, 2.50 to 1.75 Ga. Geological Society of America Bulletin, **103**:522-537.

Parrish, R.R. 1989. U-Pb geochronology of the Cape Smith Belt and Sugluk block, northern Quebec. Geoscience Canada, **16**:126-130.

Parrish, R.R. 1990. U-Pb dating of monazite and its applications to geological problems. Canadian Journal of Earth Sciences, **27**:1431-1450.

Peacock, M.A. 1931. Classification of igneous rock series. Journal of Geology, **39**:54-67.

Pearce, J.A. 1983. Role of the sub-continental lithosphere in magma genesis at active continental margins. *In*: Continental basalts and mantle xenoliths. *Edited by* C.J. Hawkesworth and M.J. Norry. Shiva Publishing Ltd., Cambridge, pp. 230-249.

Pearce, J.A., Harris, N.B.W., and Tindle, A.G. 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. Journal of Petrology, **25**:956-983.

Percival, J.A. 1990. Archean tectonic setting of granulite terranes of the Superior Province, Canada: a view from the bottom. *In*: Granulites and Crustal Evolution. *Edited by* D. Vielzeuf and P. Vidal. Kluwer Academic Publishers, London, pp. 171-193.

Percival, J.A., and Card, K.D. 1983. Archean crust as revealed in the Kapuskasing uplift, Superior Province, Canada. Geology, **11**:323-326.

Percival, J.A., and McGrath, P.H. 1986. Deep crustal structure and tectonic history of the northern Kapuskasing uplift of Ontario: an integrated petrological-geophysical study. Tectonics, **5**:553-572.

Percival, J.A., Fountain, D.M., and Salisbury, M.H. 1992. Exposed crustal cross sections as windows on the lower crust. *In*: Continental Lower Crust. *Edited by* D.M. Fountain, R.J. Arculus and R.W. Kay. Elsevier Science, Amsterdam, pp. 317-362.

Perfit, M.R., Brueckner, H., Lawrence, J.R., and Kay, R.W. 1980. Trace element and isotopic variations in a zoned pluton and associated volcanic rocks, Unalaska Island, Alaska: a model for fractionation in the Aleutian calcalkaline suite. Contributions to Mineralogy and Petrology, **73**:69-87.

Picard, C., Lamothe, D., Piboule, M., and Oliver, R. 1990. Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt (New-Quebec). Precambrian Research, **47**:223-249.

Pin, C. 1990. Evolution of the lower crust in the Ivrea Zone: a model based on isotopic and geochemical data. *In*: Granulites and Crustal Evolution. *Edited by* D. Vielzeuf and P. Vidal. Kluwer Academic, London, pp. 87-110.

Pitcher, W.S. 1982. Granite type and tectonic environment. *In*: Mountain Building Processes. *Edited by* K.J. Hsu. Academic Press, London, pp. 19-40.

Pitchowiak, S., Buchelt, M., and Damm, K.-W. 1990. Magmatic activity and tectonic setting of the early stages of the Andean cycle in northern Chile. *In*: Plutonism from Antarctica to Alaska. *Edited by* S. Mahlburg Kay and C.W. Rapela. Geological Society of America, Special Paper 241, pp. 127-144.

Rapp, R.P., Watson, B.E., and Miller, C.F. 1991. Partial melting of amphibolite/eclogite and the origin of Archean trondhjemites and tonalites. Precambrian Research, **51**:1-25.

Roddick, J.A. 1983. Geophysical review and composition of the Coast Plutonic Complex, south of latitude 55° N. *In* Circum-Pacific Plutonic Terranes. *Edited by* J.A. Roddick. Geological Society of America, Memoir 159, pp. 195-211.

Roddick, J.A., (*Editor*). 1983. Circum Pacific Pluton Terranes. Geological Society of

America Memoir 159 pp. 316.

Rogers, G., and Hawkesworth, C.J. 1989. A geochemical traverse across the North Chilean Andes: evidence for crust generation from the mantle wedge. Earth and Planetary Science Letters, *91*:217-285.

Rudnick, R.L., and Presper, T. 1990. Geochemistry of intermediate- to high-pressure granulites. *In* Granulites and crustal evolution. *Edited by* D. Vielzeuf and T. Presper. Kluwer Academic Publishers, Boston, pp. 523-551.

Rudnick, R.L., McLennan, S.M., and Taylor, S.R. 1985. Large ion lithophile elements in rocks from high-pressure granulite facies terrains. Geochimica Cosmochimica et Acta, **49**:1645-1655.

Saunders, A.D., Tarney, J., and Weaver, S.D. 1980. Transverse geochemical variations across the Antarctic Peninsula: implications for the genesis of calc-alkaline magmas. Earth and Planetary Science Letters, **46**:344-360.

Schroeder, B., Thompson, G., Sulanowska, M., and Ludden, J.N. 1980. Analysis of geologic materials using an automated X-ray fluorescence system. X-ray Spectrometry, **9**:198-205.

Scott, D.J., Helmstaedt, H., and Bickle, M.J. 1992. Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. Geology, **20**:173-176.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1989. The 1998 Ma Purtuniq ophiolite: imbricated and metamorphosed oceanic crust in the Cape Smith Thrust Belt, northern Quebec. Geoscience Canada, **16**:144-147.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1991. Geology and chemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. *In* Ophiolite Genesis and Evolution of the Oceanic Lithosphere. *Edited by* T. Peters, A. Nicolas and R.G. Coleman. Kluwer Academic Publishers, Amsterdam, pp. 817-849.

Scott, D.J., and St-Onge, M.R. in press. Constraints on Pb closure temperature in titanite based on rocks from the Ungava Orogen, Canada: implications for the U-Pb geochronology and P-T-t path determinations. Geology.

Sekine, T., and Wyllie, P.J. 1982. Phase relationships in the system KAlSiO4 - Mg2SiO4 - SiO2 - H2O as a model for hybridization between hydrous siliceous melts and peridotite. Contributions to Mineralogy and Petrology, **79**:368-374.

Silver, L.T., and Chappell, B.W. 1988. The Peninsular Ranges Batholith: an insight into the evolution of the Cordilleran batholiths of southwestern North America. Transactions of the Royal Society of Edinburgh: Earth Sciences, **79**:105-121.

Skulski, T., Wares, R.P., and Smith, A.D. 1993. Early Proterozoic (1.88-1.87 Ga) tholeiitic magmatism in the New Quebec orogen. Canadian Journal of Earth Sciences, **30**:1505-1520.

Smith, A.D., and Ludden, J.N. 1989. Nd isotopic evolution of the Precambrian mantle. Earth and Planetary Science Letters, **93**:14-22.

St-Onge, M.R., and Lucas, S.B. 1990a. Evolution of the Cape Smith Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. *In:* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 313-351.

St-Onge, M.R., and Lucas, S.B. 1990b. Early Proterozoic collisional tectonics in the internal zone of the Ungava (Trans-Hudson) orogen, Lacs Nuvilik and Sugluk map areas, Quebec. *In*: Current Research, Part C. Geological Survey of Canada, Paper 90-1C, pp. 119-132.

St-Onge, M.R., and Lucas, S.B. 1992. New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. *In*: Current Research, Part C. Geological Survey of Canada, Paper 92-1C, pp. 31-41.

St-Onge, M.R., and Lucas, S.B. 1993. Geology of the eastern Cape Smith Belt: Parts of the Kangiqsujuaq, Cratère du Nouveau-Québec, and Lacs Nuvilik map areas, Quebec. *In*: Geological Survey of Canada, Memoir 438, pp. 110.

St-Onge, M.R., Lucas, S.B., and Parrish, R.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Canadian Journal of Earth Science,

29:746-764.

St-Onge, M.R., and Lucas, S.B. 1995. Large-scale fluid infiltration, metasomatism and re-equilibration of Archean basement granulites during Paleoproterozoic thrust belt construction, Ungava orogen, Canada. Journal of Metamorphic Geology, **13**:509-535.

St-Onge, M.R., Lucas, S.B., Scott, D.J., Bégin, N.J., Helmstaedt, H., and Carmicheal, D.M. 1988. Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec. *In*: Current Research, Part C. Geological Survey of Canada, Paper 88-1C, pp. 1-18.

Stern, R.A., Hanson, G.N., and Shirey, S.B. 1989. Petrogenesis of mantle-derived LILE-enriched Archean monzodiorites and trachyandesites (sanukitoids) in southwestern Superior Province. Canadian Journal of Earth Sciences, **26**:1688-1712.

Stern, R., Percival, J., and Mortensen, J. 1994. Geochemical evolution of the Minto block: a 2.7 Ga continental magmatic arc built on the Superior proto-craton. Precambrian Research, **65**:115-153.

Stern, R.A., Syme, E.C., Bailes, A.H., and Lucas, S.B. 1995. Paleoproterozoic (1.90-1.86 Ga) arc volcanism in the Flin Flon Belt, Trans-Hudson Orogen, Canada. Contributions to Mineralogy and Petrology, **119**:117-141.

Stern, R.A., Syme, E.C., and Lucas, S.B. 1995. Geochemistry of 1.9 Ga MORB- and OIB-like basalts from the Amisk collage, Flin Flon Belt, Canada: Evidence for an intra-oceanic origin. Geochimica Cosmochimica et Acta, **59**:in press.

Sun, S.S., and McDonough, W.F. 1989. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. *In*: Magmatism in the ocean basins. *Edited by* A.D. Saunders and M.J. Norry. Geological Society, Special Publication 42, pp. 313-345.

Tatsumi, Y., Hamilton, D.L., and Nesbitt, R.W. 1986. Chemical characteristics of fluid phase released from a subducted lithosphere and origin of arc magmas: Evidence from high-pressure experiments and natural rocks. Journal of Volcanology and Geothermal Research, *29*:293-309.

Taylor, F.C. 1982. Reconnaissance geology of a part of the Canadian Shield, northern Quebec and Northwest Territories. In Geological Survey of Canada, Memoir 399, pp. 32.

Taylor, F.C., and Loveridge, W.D. 1985. A Rb-Sr study of granitiod intrusive rocks of the Cape Smith Belt, northern Quebec. *In*: Current Research, Part A. Geological Survey of Canada, Paper 85-1A, pp. 65-67.

Taylor, S.R., and McLennan, S.M. 1985. The continental crust: its composition and evolution. Blackwell Scientific Publications, Boston, pp. 312.

Truscott, M.G., and Shaw, D.M. 1990. Average composition of lower and intermediate continental crust, Kapuskasing structural zone, Ontario. *In* Exposed cross-sections of the continental crust. *Edited by* M.H. Salisbury and D.M. Fountain. Kluwer Academic Publishers, Boston, pp. 421-436.

van der Leeden, J., Belanger, M., Danis, D., Girard, R., and Martelain, J. 1990. Lithotectonic domains in the high-grade terrain east of the Labrador Trough (Quebec). *In*: The Early Proterozoic Trans-Hudson orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Survey of Canada, Special Paper 37, pp. 371-386.

Van Kranendonk, M.J., St-Onge, M.R. and Henderson, J.R., 1993. Paleoproterozoic tectonic assembly of Northeast Laurentia through multiple indentations. Precambrian Research, **63**:325-347.

Veizer, J. 1983. Geological evolution of the Archean-Early Proterozoic Earth. *In*: Earth's earliest biosphere. Its origin and evolution. *Edited by* J.W. Schopf. Princeton University Press, Princeton, N.J., pp. 240-259.

Wardle, R.J., Ryan, B., Nunn, G.A.G., and Mengel, F.C. 1990. Labrador segment of the Trans-Hudson orogen: Crustal development through oblique convergence and collision. *In*: The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 353-369.

Wares, R.P., and Skulski, T. 1992. Evolution of the 2.2-1.7 Ga New Quebec orogen.

EOS (Transactions American Geophysical Union), **73**: 333.

Weaver, B.L., and Tarney, J. 1981. Lewisian gneiss geochemistry and Archean crustal development models. Earth and Planetary Science Letters, **55**:171-180.

Weaver, B.L., and Tarney, J. 1983. Elemental depletion in Archaen granulite-facies rocks. *In*: Migmatites, Melting and Metamorphism. *Edited by* M.P. Atherton and C.D. Gribble, pp. 250-263.

Wilson, M. 1989. Igneous Petrogenesis. Unwin Hyman, London, pp. 466.

Winchester, J.A., and Max, M.D. 1984. Element mobility associated with synmetamorphic shear zones near Scotchport, NW Mayo, Ireland. Journal of Metamorphic Geology, **2**:1-11.

Wyllie, P.J. 1984. Constraints imposed by experimental petrology on possible and impossible magma sources and products. Philosophical Transactions of the Royal Society of London, **A310**:439-456.

Zindler, A. 1982. Nd and Sr isotopic studies of komatiites and related rocks. *In* Komatiites. *Edited by* N.T. Arndt and E.G. Nisbet. George Allen and Unwin, London, pp. 399-420.

Zindler, A., Hart, S.R., Frey, F.A., and Jakobsson, S.P. 1979. Nd and Sr isotope ratios and rare earth element abundances in Reykjanes Peninsula basalts: evidence for mantle heterogeneity beneath Iceland. Earth and Planetary Science Letters, **45**:249-262.

CHAPTER 2

Geochemistry of mafic magmas from the Ungava Orogen, Québec, Canada and implications for mantle reservoir compositions at 2.0 Ga

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Chemical Geology, v. 120 (1995), p. 361-380

submitted 24 December 1993, accepted 15 June 1994

ABSTRACT

The Ungava orogen of northern Quebec is one of the best preserved Proterozoic mobile belts of the world, recording more than 200 Ma of plate divergence and convergence. Voluminous magmatism associated with rifting of the Superior Province basement at 2.04 Ga resulted in the development of a volcanic rift margin sequence and an ocean basin. Four distinct mafic magma suites were erupted: (1) continental basalts (Eskimo Formation, western Povungnituk Group) with moderate to high Zr/Nb and negative Nb anomalies which have interacted with the continental crust ($\varepsilon_{Nd(2.0\ Ga)}$ to -7.4); (2) mafic lavas from the Flaherty Formation, eastern and central Povungnituk Group and some Watts Group lavas associated with passive margin rifting, having slightly enriched isotopic signatures ($\varepsilon_{Nd(2.0\ Ga)}$ = +2.7 to +4.4) compared to the contemporaneous depleted mantle, and high $(Nb/Y)_n$ and low Zr/Nb ratios (≈ 4.4 and 8.9 respectively); (3) a highly alkaline OIB-like suite ($\varepsilon_{Nd(2.0\ Ga)}$ = +2.3 to +3.2, $(Nb/Y)_n > 12$) within the Povungnituk Group composed of nephelinites, basanites and phonolites; and (4) depleted Mg-rich basalts and komatiitic basalts ($\varepsilon_{Nd(2.0\ Ga)} \approx +4.5$ to +5.5) with trace element characteristics of N-MORB, but with higher Fe and lower Al than primitive MORB (Chukotat Group, Ottawa Islands and some Watts Group samples). The ocean basin into which these lavas were erupted was subsequently destroyed during subduction between ca. 1.90-1.83 Ga resulting in the development of a magmatic arc (Narsajuaq terrane and Parent Group).

The Ungava magmas provide a unique window into the mantle at 2.0 Ga. The chemical and isotopic similarity of these Proterozoic magmas to modern-day magmas provide strong evidence that the interplay between depleted mantle, OIB mantle and sub-continental mantle during the Proterozoic was comparable to that of the modern Earth.

RÉSUMÉ

L'Orogène de l'Ungava du nord du Québec, une des ceintures mobiles les mieux préservée au monde, enregistre > 200 Ma de divergence et convergence de plaques tectoniques. Le rifting du socle de la Province du Supérieur ca 2.04 Ga a résulté au

développement d'une séquence de marge de rift volcanique et d'un bassin océanique caractérisé par de volumineux épanchements magmatiques. Quatre suites distinctes de roches volcaniques furent mises en places: (1) basaltes continentaux (Formation Eskimo, Groupe de Povungnituk occidental et central) avec des ratios Zr/Nb modérés à élevés et des anomalies négatives de Nb, qui ont interagis avec la croûte continentale ($\varepsilon_{Nd\,(2.0\,Ga)}$ plus grand que -7.4); (2) laves mafiques de la Formation de Flaherty, du Groupe de Povungnituk oriental et certaines laves du Groupe de Watts, associées au rifting d'une marge passive, qui ont une signature isotopique légèrement enrichie ($\varepsilon_{Nd\,(2.0\,Ga)}$ = +2.7 à +4.4) comparé au manteau contemporain appauvri, des ratios $(Nb/Y)_n$ élevés et des ratios Zr/Nb bas (4.4 et 8.9 respectivement); (3) une suite de type OIB hautement alcaline ($\varepsilon_{Nd\,(2.0\,Ga)}$ = +2.3 à +3.2, $(Nb/Y)_n > 12$), à l'intérieur du Groupe de Povungnituk, composée de néphélinites, basanites et phonolites; et (4) basaltes magnésiens appauvris et basaltes komatiitiques ($\varepsilon_{Nd\,(2.0\,Ga)}$ = +4.5 à +5.5) avec des éléments traces caractéristiques de N-MORB, mais avec Fe plus élevé et Al plus bas que le MORB primitif (Groupe de Chukotat, Iles d'Ottawa et certains échantillons du Groupe de Watts). Le bassin océanique dans lequel ces laves furent mises en place a été subséquemment détruit par subduction entre 1.90 et 1.83 Ga, résultant au développement d'un arc magamatique (terrane de Narsajuaq et Groupe de Parent).

Les magmas de l'Ungava fournissent une fenêtre unique de l'intérieur du manteau à 2.0 Ga. Les similarités chimiques et isotopiques de ces magmas Protérozoiques avec les magmas contemporains fournissent une forte évidence que l'interaction entre manteau appauvri, manteau OIB et manteau sub-continental pendant le Protérozoique se comparait à la terre moderne.

2.1 INTRODUCTION

Geochemical studies of mafic volcanic rocks in modern tectonic settings have identified a number of distinct mantle reservoirs involving asthenosphere, deep mantle (>670 km), ancient subcontinental lithospheric mantle and subduction-related sources (i.e., Morgan, 1971; Basaltic Study Volcanism Project, 1981; Sun and McDonough, 1989; Menzies, 1990). Our knowledge of the chemistry of the modern mantle can provide a basis for comparison with older rocks and their sources, allowing us to elucidate the evolution of the mantle through time (White and Hofmann, 1982; Hofmann et al., 1986; Zindler and Hart, 1986; Campbell and Griffiths, 1992). Commonly however, studies on ancient rocks suffer from limited and sometimes ambiguous geochemical data sets.

Geoscientific investigations of the Proterozoic Ungava orogen date back 100 years, beginning with the observations by Bell (1885) in the areas of Hudson Strait and Hudson Bay. Today our knowledge of the area is extensive and a comprehensive geochemical data set exists for these rocks. The Ungava orogen now represents one of the best exposed and well studied ancient orogens for which tectonic models involving plate divergence and subsequent convergence have been constructed (e.g., Baragar and Scoates, 1981; Francis et al., 1983; Arndt et al., 1987; Picard et al., 1990; St-Onge and Lucas, 1990b, 1993; Hegner and Bevier, 1991; Scott et al., 1991; St-Onge et al., 1992).

In this comprehensive review we summarize the geochemical and isotopic characteristics of the mafic rocks of the Ungava orogen, which when combined with an understanding of the geological framework for the region, allow us to present an integrated model for the petrogenesis and evolution of the magmas of the orogen and to assess the nature of the mantle reservoirs at 2.0 Ga.

2.2 GEOLOGICAL FRAMEWORK

The Ungava orogen (Fig. 1; St-Onge and Lucas, 1990a) of northern Quebec is an Early Proterozoic rifted continental margin to arc-continent collisional zone which represents one of the best exposed sections of the Trans-Hudson Orogen (Hoffman, 1990). More than 200 Ma of plate divergence and convergence is recorded within the tectonostratigraphic units of the orogen (Lucas et al., 1992; St-Onge et al., 1992).

Rifting of the 2.7 - 2.9 Ga. Superior Province basement ca. 2.04 Ga (Machado

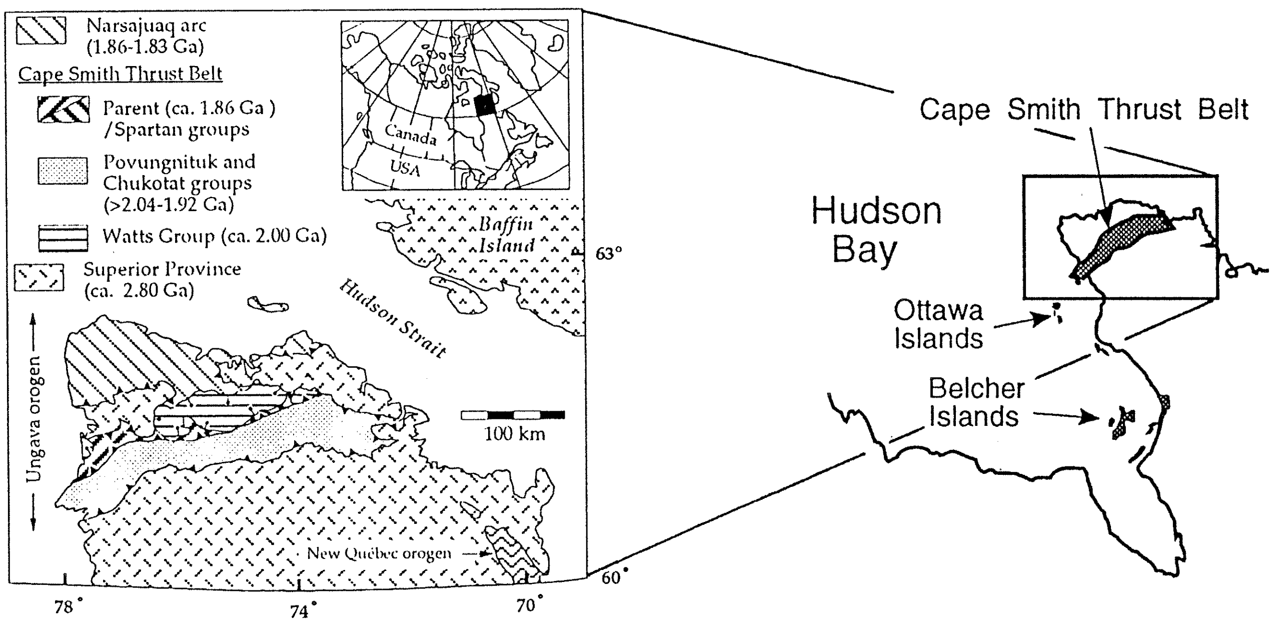


Figure 2.1: Maps illustrating the location and tectonic elements of the Ungava orogen and the associated Ottawa and Belcher Islands (after St-Onge and Lucas, 1992).

et al., 1993) resulted in the development of a volcanic rifted margin and oceanic basin. Voluminous magmatism comprising continental tholeiites, alkaline lavas, komatiitic basalt and associated Mg-tholeiites reflect the rifting event (Francis and Hynes, 1979; Francis et al., 1981, 1983; Hynes and Francis, 1982; Taylor, 1982; Arndt et al., 1987; Baragar and Scoates, 1987; Picard et al., 1990; St-Onge and Lucas, 1990b, 1993). In addition, the crustal portion of what has been interpreted as an ophiolite (St-Onge et al., 1988; Scott et al., 1989, 1991) and highly incompatible element-enriched alkaline lavas (Francis et al., 1983; Gaonac'h et al., 1989, 1992; Picard et al., 1990) record the Earth's earliest documented section of oceanic crust and the oldest alkaline ocean island suite respectively. Subduction resulted in the destruction of the oceanic basin and the development of a magmatic arc between approximately 1.90 - 1.83 Ga (St-Onge and Lucas, 1990a; Dunphy and Ludden, 1992; St-Onge et al., 1992). Arc - continent collision was initiated after 1.83 Ga with terminal collision occurring before 1.76 Ga (Parrish, 1989; St-Onge et al., 1992; Dunphy et al., 1995).

Deformation and metamorphic histories preserved within the lithologies of the Ungava orogen reflect multiple tectonic events. In general, four deformation events can be documented (Lucas and St-Onge, 1992): (1) a pre-accretion deformation (D_1), which is Archean in age in the Superior Province basement and Proterozoic in age for the other domains; (2) a Proterozoic accretion-related deformation (D_2), which is recorded in all elements of the orogen and manifest as a penetrative shear adjacent to the basal décollement, and thrusting and imbrication within the Proterozoic units; (3) regional-scale folding of all tectonic elements about east-west axes (D_3); and (4) subsequent cross-folding of all units about north-south axes (D_4).

2.3 A 2 Ga. CONTINENTAL RIFTED VOLCANIC MARGIN SEQUENCE

2.3.1 Geological features

The continental rift sequence (2.04 - 1.96 Ga; Parrish, 1989; Machado et al., 1993) of the Ungava orogen is exposed along the southern margin of the Cape Smith Belt and on the Belcher Islands of Hudson Bay (Fig. 1). Autochthonous and allochthonous sequences of clastic sedimentary units (semipelite, quartzite, ironstone, pelite) and overlying mafic pillowed and massive flows with minor interfingered clastic sediments characterize the Povungnituk Group within the Cape Smith Belt (Hynes and Francis, 1982; Francis et al., 1983; Picard et al., 1990; St-Onge and Lucas, 1990b, 1993). These authors have interpreted the volcanic and sedimentary rocks of the Povungnituk Group as lithologies related to rifting of the Superior Province continental crust. Also found within the upper Povungnituk Group is a sequence of basanites, nephelinites, phonolites and pyroclastic rocks representing a potential ocean island suite (Gaonac'h et al., 1989, 1992; Picard et al., 1990).

Within the predominantly sedimentary sequence of the Belcher Islands is a series of subaerial and subaqueous lava flows of the older Eskimo and younger Flaherty Formations which may be related to the Povungnituk Group (Jackson, 1960; Baragar and Lamontagne, 1980; Baragar and Scoates, 1981, 1987; Legault et al., in press). A recent Pb-Pb age for the Belcher Island volcanic Formations of 1960 +/- 100 Ma (Arndt and Todt, in press) supports this hypothesis as it agrees well with the published U-Pb ages for the Povungnituk Group.

Geological and geochemical studies of Early Proterozoic volcanic rocks exposed on other islands and along the coast of Hudson Bay (e.g., Sleeper Islands, Long Island, Manitounuk Islands and Richmond Gulf) have correlated these rocks with the formations of the Belcher Islands and the Povungnituk Group (Baragar and Lamontagne, 1980; Baragar and Scoates, 1981, 1987; Legault et al., in press).

2.3.2. Geochemical and isotopic characteristics

In the Cape Smith Belt, the Povungnituk Group consists of Fe- and Ti-rich continental tholeiites (maximum MgO 9.8% wt.) having variably enriched light rare earth elements (LREE's; Fig. 2a; Francis et al., 1983, Picard, 1989a, 1989b; Picard et al., 1990). A marked distinction between basalts of the western region and those of the central and eastern regions is seen in certain trace element ratios as illustrated in Figures

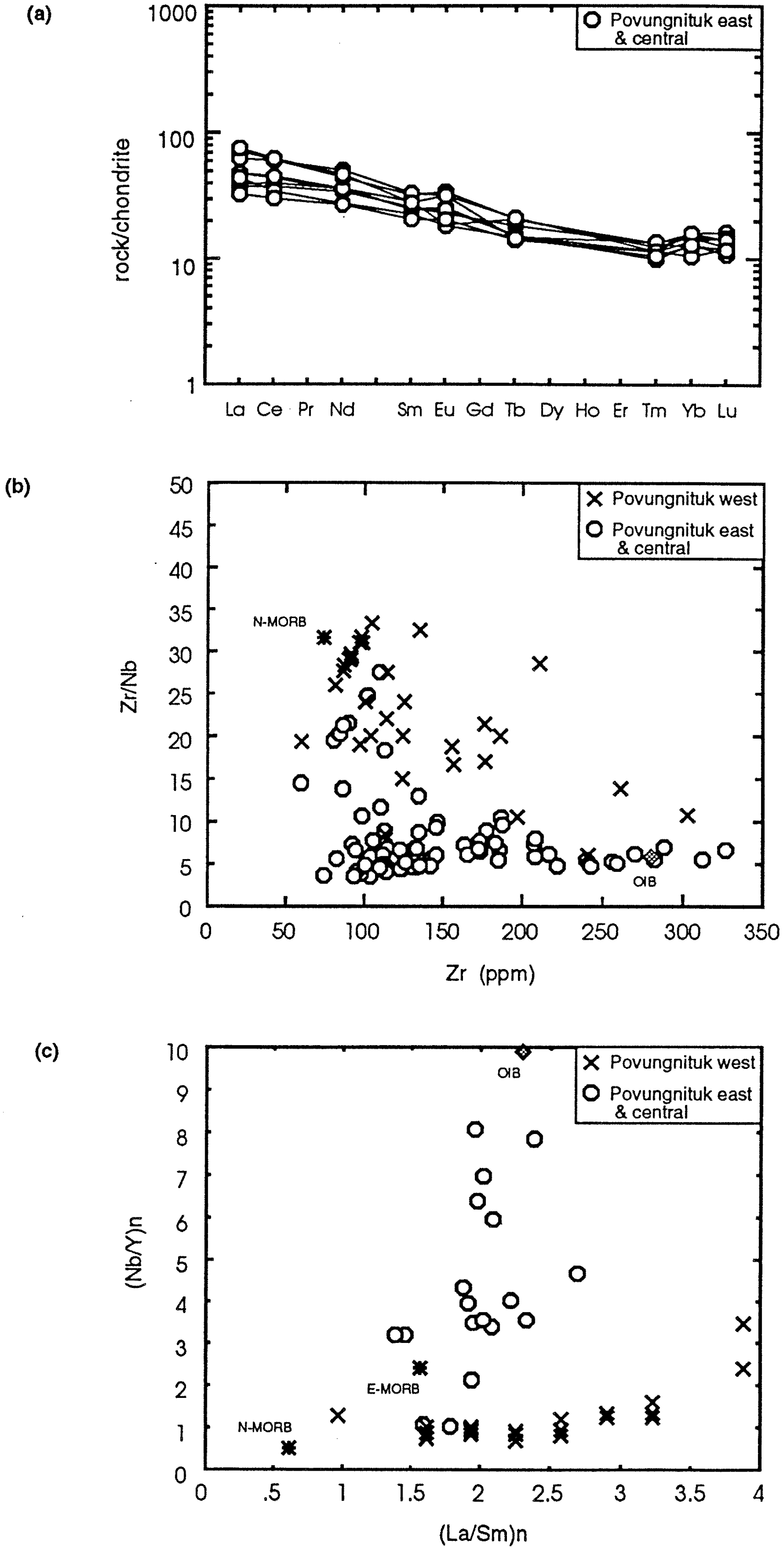


Figure 2.2. (a) Chondrite normalized REE diagram for representative samples from the eastern and central portion of the Povungnituk Group; (b) Zr/Nb vs Zr diagram for all Povungnituk Group samples; (c) $(Nb/Y)_n$ vs $(La/Sm)_n$ for available Povungnituk Group samples. Data compiled from Francis et al. (1983), and unpub. data; Picard (1989a, 1989b). Chondrite normalization factors and average compositions of N-MORB, E-MORB and OIB from Sun and McDonough (1989).

2b and 2c (Legault et al., in press), with the central and eastern basalts having low Zr/Nb ratios (mostly less than 11 although some samples have values up to 27.5) and high $(Nb/Y)_n$ ratios (Francis et al., 1983; Picard, 1989b). In addition, the central and eastern Povungnituk Group basalts do not have negative Nb anomalies. Melting of an enriched mantle source could explain the geochemical characteristics of these basalts. This contrasts with the lavas of the western Povungnituk Group which have negative Nb anomalies ($(Nb/La)_n = 0.2$-1.2), high Zr/Nb ratios (most samples are >11, up to 33.3) and much lower $(Nb/Y)_n$ ratios (average ratio of 1.7; Picard, 1989a) suggesting crustal contamination was involved in their petrogenesis. Representative analyses of these lavas are presented in Table 1 and a summary of their geochemical characteristics are found in Table 2.

The Povungnituk lavas have a range in initial ε_{Nd} (1.96 Ga) compositions (Fig. 3) from +1.9 to +4.1 (Zindler, 1982; Hegner and Bevier, 1991) relative to a value of approximately +5.5 to +6.5 for a contemporaneous model depleted mantle (Jahn et al., 1987, 1988). The most isotopically enriched Povungnituk lava is a rhyolite ($\varepsilon_{Nd(1.96\ Ga)} = +0.4$; Hegner and Bevier, 1991), potentially produced by partial melting and contamination during ponding of mafic magma at the crust-mantle interface during rifting (Francis et al., 1983; Hegner and Bevier, 1991). Assimilation of up to 15% of continental crust can explain the isotopic composition of the other Povungnituk samples (Hegner and Bevier, 1991), although these authors point out that these samples define a shallow trend toward enriched mantle compositions in the isochron diagram (Fig. 3) and suggest that they are the product of mixing of a depleted mantle component with magmas derived from enriched mantle sources. Only a few samples of Povungnituk Group basalts have been analyzed for their Sr-isotope composition, with the most reliable results coming from clean clinopyroxene separates which have low average initial $^{87}Sr/^{86}Sr$ of approximately 0.7022 (Zindler, 1982) which suggest crustal contamination was not prevalent in these samples. Pb isotopic analysis of mafic Povungnituk lavas also do not show evidence for contamination as they have a calculated model $^{238}U/^{204}Pb$ (μ^*) value of 7.99, comparable to values estimated for

Table 1.Major and trace element data for representative mafic samples from the Ungava orogen

Location	E Pov	E Pov	W Pov	W Pov	Pov alk	Pov alk	Eskimo	Eskimo	Flaherty	Flaherty
Sample number	76410	61a	207-2	210-1	1460	1470	Hb-4	Qk-1	Ft-8	Ft-21
SiO_2 (wt%)	53.21	47.30	45.50	47.90	42.70	43.11	47.32	52.26	47.04	46.66
TiO_2	1.67	2.17	3.36	1.44	4.85	5.46	2.18	1.33	2.41	1.49
Al_2O_3	12.10	17.20	12.70	12.90	9.51	10.90	13.69	14.67	12.79	13.51
Fe_2O_3	12.02	10.00	16.90	14.00	16.80	14.30	18.37	12.65	16.46	14.98
MgO	5.76	6.55	5.83	6.98	6.44	8.89	5.40	6.18	5.84	5.97
MnO	0.27	0.16	0.24	0.16	0.18	0.24	0.16	0.25	0.28	0.23
CaO	9.17	12.60	10.70	10.10	10.90	12.30	7.81	8.16	10.10	10.60
Na_2O	4.01	2.00	2.17	3.03	2.98	1.67	2.91	2.79	2.33	3.17
K_2O	0.14	1.06	0.29	0.09	1.77	0.91	0.82	1.13	0.23	0.42
P_2O_5	0.20	0.21	0.28	0.12	0.78	0.85	0.31	0.19	0.27	0.14
LOI	1.00	1.05	2.36	3.10	3.44	2.11	2.08	1.72	2.88	3.19
Total	99.55	100.30	100.33	99.82	100.35	100.74	101.05	101.33	100.63	100.36
Cr (ppm)	60	123	220	88	114	239	83	104	159	127
Ni	41	86	91	68	134	125	41	59	67	63
Sc	41.6	--	--	--	23.8	25.3	39.1	35.7	43.0	39.4
V	--	373	325	300	282	298	366	226	352	296
Rb	--	43	12	4	55	18	10	18	5	6
Sr	--	201	120	180	636	540	175	218	146	210
Ba	--	109	121	23	672	635	323	459	68	228
Ta	1.45	--	<5	<5	--	--	0.54	0.28	1.00	0.63
Nb	19	25	<3	9	120	150	11	8	20	12
Hf	3.04	--	--	--	--	--	4.60	2.96	4.02	2.55
Zr	183	170	87	150	580	810	190	117	165	93
Y	26	25	21	24	35	51	39	21	37	24
Th	--	--	5.00	7.00	--	--	2.01	2.84	1.76	1.30
U	--	--	--	--	--	--	0.15	0.29	0.48	0.33
La (ppm)	17.45	14.47	7.00	12.00	99.60	142.00	29.32	22.09	17.71	10.51
Ce	37.40	28.89	15.00	35.00	207.00	277.00	61.49	44.60	39.74	23.83
Pr	--	--	--	--	--	--	--	--	--	--
Nd	21.44	17.74	35.00	55.00	98.00	121.00	31.63	20.83	23.77	15.30
Sm	4.18	4.72	<2	<2	21.69	26.72	6.84	4.53	6.12	3.99
Eu	1.82	1.29	1.00	2.00	5.10	6.70	2.10	1.32	1.79	1.65
Gd	--	--	--	--	--	--	--	--	--	--
Tb	0.53	0.88	--	--	1.90	2.24	1.16	0.68	1.04	0.75
Dy	--	--	3.00	3.00	--	--	--	--	--	--
Ho	--	--	--	--	1.70	2.20	--	--	--	--
Er	--	--	--	--	--	--	--	--	--	--
Tm	0.26	--	--	--	--	--	0.60	0.32	0.34	0.51
Yb	2.13	3.18	--	--	2.70	3.30	4.26	2.32	4.03	2.87
Lu	0.29	0.42	--	--	0.36	0.40	0.66	0.41	0.59	0.42

Geochemical data compiled from Francis et al. (1983); Arndt et al. (1987); Picard (1989a, 1989b); Scott et al. (1991); Legault et al. (in press); Dunphy and Ludden (in prep.).

Table 1 (continued)

Location Sample number	Chukotat 65610	Chukotat 117a	Chukotat 127c	Ottawa G26	Ottawa G59	Watts S199	Watts E7	Watts E4	Nar L90b	Nar D28	Nar D41c
SiO_2 (wt%)	46.70	47.10	46.20	49.00	51.50	49.10	47.60	42.40	49.65	52.63	52.45
TiO_2	0.78	0.65	0.58	0.82	0.85	0.96	1.04	2.32	0.92	0.58	1.03
Al_2O_3	10.50	10.10	9.31	11.30	11.40	11.70	12.30	9.54	15.43	16.65	16.65
Fe_2O_3	11.58	12.30	13.10	12.67	12.34	14.28	13.24	17.19	10.11	8.27	10.13
MgO	12.90	14.40	17.30	13.10	11.10	6.27	9.03	11.64	7.61	6.57	4.52
MnO	0.16	0.18	0.21	0.00	0.00	0.19	0.18	0.22	0.17	0.15	0.16
CaO	12.60	10.60	10.30	11.30	9.80	10.12	10.12	11.84	9.17	9.18	8.55
Na_2O	0.39	0.66	0.54	2.08	2.60	2.11	2.13	0.41	3.10	3.75	4.25
K_2O	0.18	0.06	0.06	0.10	--	0.08	0.24	0.11	1.26	0.65	0.87
P_2O_5	0.05	0.03	0.01	0.00	0.00	0.07	0.07	0.30	0.24	0.15	0.28
LOI	2.93	2.66	3.55	1.40	2.37	3.43	2.41	3.33	0.54	0.97	0.28
Total	98.77	98.74	101.16	100.36	99.58	99.01	98.36	99.30	98.20	99.55	99.17
Cr (ppm)	1176	673	789	1358	783	147	294	644	231	211	116
Ni	532	528	544	382	206	82	142	447	146	59	45
Sc	34.5	34	34	41.0	40.0	40.6	35.5	32.0	27.4	23.8	27.2
V	--	227	210	283	291	367	353	389	--	--	--
Rb	--	3	3	2	2	1	3	1	15	9	7
Sr	--	110	28	83	108	172	158	258	836	907	660
Ba	--	31	21	--	--	26	113	23	542	445	374
Ta	--	.72	1.11	0.17	0.12	0.55	0.29	1.42	1.00	0.80	2.10
Nb	--	2	2	3	2	4	3	15	8	3	8
Hf	--	1.41	.93	1.27	1.20	0.64	0.72	2.52	3.70	1.30	2.40
Zr	32	45	40	44	42	70	73	144	134	47	90
Y	14	14	13	17	18	20	16	22	28	13	23
Th	--	.72	--	0.74	--	0.31	0.21	0.98	0.70	0.40	1.70
U	--	--	--	--	--	.11	.08	.29	0.20	0.20	2.40
La (ppm)	2.15	1.99	2.84	3.54	1.90	4.00	3.45	13.69	15.77	10.41	8.90
Ce	5.92	7.77	7.75	8.52	5.38	9.56	9.16	32.99	38.83	23.10	21.40
Pr	--	--	--	--	--	1.44	1.48	4.64	--	--	--
Nd	4.30	5.25	4.87	5.69	4.60	6.75	7.10	19.90	24.48	14.71	15.00
Sm	1.59	1.57	1.53	1.69	1.60	2.19	2.36	5.15	5.06	2.93	3.38
Eu	0.68	0.59	0.66	0.64	0.66	0.79	0.95	1.78	1.32	1.07	1.07
Gd	--	--	--	2.23	2.24	2.68	2.68	5.17	--	--	--
Tb	0.37	0.44	0.42	--	--	0.54	0.51	0.85	0.81	0.37	0.60
Dy	--	--	--	--	2.67	3.68	3.17	5.06	--	--	--
Ho	--	--	--	--	--	0.80	0.64	0.91	1.22	0.37	0.84
Er	--	--	--	1.60	1.69	2.42	1.87	2.20	--	--	--
Tm	--	--	--	--	--	0.37	0.25	0.28	--	--	--
Yb	1.50	1.29	1.45	1.48	1.59	2.33	1.53	1.65	2.83	1.34	2.30
Lu	0.21	0.21	0.24	0.22	0.23	0.33	0.20	0.22	0.40	0.19	0.32

Table 2. Summary of geochemical characteristics for the various mafic suites of the Ungava orogen.

Magma type	Volcanic unit	No. Samples [1]	Zr/Nb [2]	$(Nb/Y)_n$ [2]	$(Nb/La)_n$ [2]	$(La/Sm)_n$ [2]	Initial ε_{Nd}
Depleted oceanic magmas	Watts	7, 7	16.3-22.6 (20.2)	1.1-2.6 (1.5)	0.9-1.0 (0.9)	0.9-1.4 (1.1)	+3.9 to +6.1
	Chukotat	12, 12	13.3-49.0 (25.1)	0.4-1.3 (0.8)	0.3-1.1 (0.6)	0.7-2.0 (1.1)	+2.7 to +4.5
Enriched mantle magmas	Povungnituk alkaline	32, 13	2.1-7.8 (5.5)	12.2-34.5 (20.9)	0.5-1.9 (1.1)	2.4-11.4 (4.9)	+1.8 to +2.6
	Watts	3, 3	9.4-14.5 (11.2)	3.4-4.3 (4.0)	0.8-1.0 (0.9)	1.6-1.7 (1.7)	+3.0 to +3.4
	Flaherty	85, 10	6.3-9.1 (7.5)	3.1-4.4 (3.6)	1.0-1.3 (1.1)	1.5-2.1 (1.8)	+2.4 to +4.3
	Povungnituk (east)	75, 19	3.5-27.5 (8.2) most <11	1.1-11.9 (5.6)	0.5-2.4 (1.3) most >1	1.4-2.7 (2.0)	+1.9 to +4.1 [3]
Continental basalts	Eskimo	42, 8	9.2-20.4 (14.6)	1.7-2.9 (2.2)	0.3-0.5 (0.4)	2.4-3.3 (2.7)	-6.0 to -7.4
	Povungnituk (west)	34, 34	6.1-33.3 (23.1) most >11	0.7-7.6 (1.7)	0.2-1.2 (0.6)	1.0-11.3 (3.7)	+1.9 to +4.1 [3]
N-MORB [4]			31.8	0.5	0.9	0.6	
E-MORB [4]			8.8	2.4	1.3	1.6	
OIB [4]			5.8	10.6	1.2	2.4	
Upper crust [5]			7.6	7.3	0.8	4.3	

Geochemical data compiled from Francis et al. (1983); Picard (1989a, 1989b); Scott et al. (1991); Gaonac'h et al. (1992); Legault et al. (in press). Isotopic data from Zindler (1982); Chauvel et al. (1987); Smith and Ludden (1989); Hegner and Bevier (1991); Gaonac'h et al. (1992).

[1] Number of samples compiled. Second number (when present) refers to number of samples with REE analysis.

[2] Range of values for each unit, with average given in parentheses.

[3] Isotopic compositions for east and west Povungnituk are the same as sample locations are unknown.

[4] Averages from Sun and McDonough (1989).

[5] Average composition from Taylor and McLennan (1985).

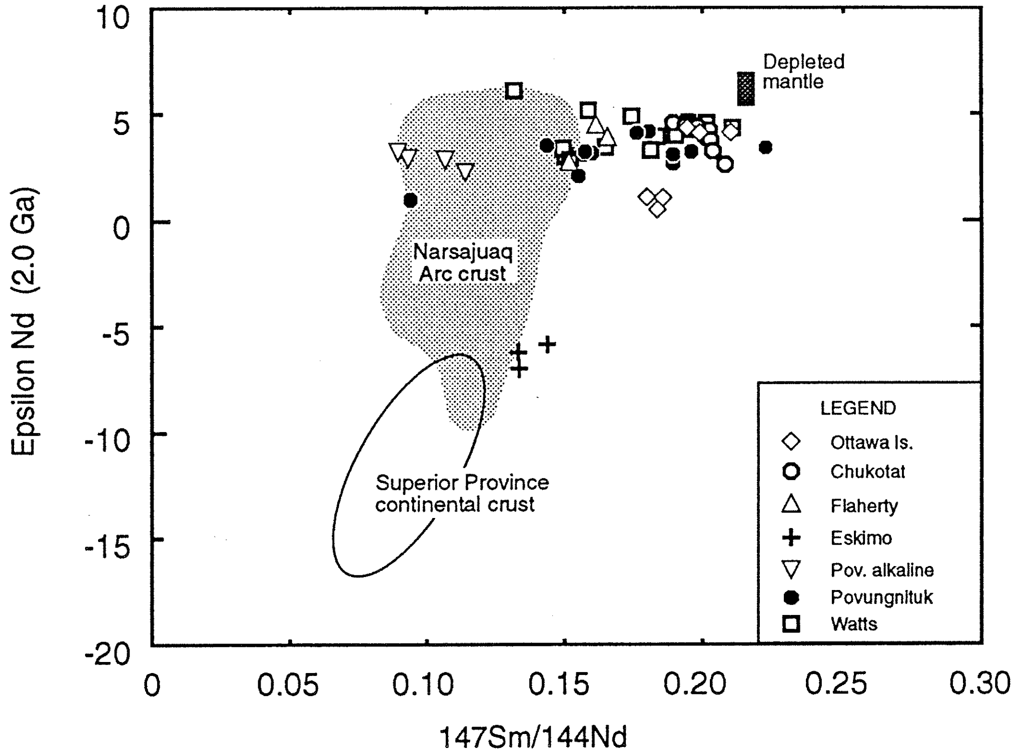


Figure 2.3. ε_{Nd} vs $^{147}Sm/^{144}Nd$ (recalculated to 2.0 Ga) for non-cumulate mafic samples from the Ungava orogen and associated Ottawa and Belcher Islands (Zindler, 1982; Chauvel et al., 1987; Smith and Ludden, 1989; Hegner and Bevier, 1991; Gaonac'h et al., 1992). The light shaded field delimits the range in Nd isotopic composition of the Narsajuaq arc rocks (Dunphy and Ludden, in preparation). Superior Province continental crust composition from Dunphy (unpublished data); depleted mantle compositions (dark rectangle) from Jahn et al. (1987, 1988). Note the trend of Povungnituk Group and Flaherty Formation samples toward enriched mantle compositions and the contamination trend of the Eskimo Formation and several Ottawa Islands samples toward Superior Province continental crust.

mantle rocks (Hegner and Bevier, 1991). The samples chosen for Pb isotopic analysis yield a poor isochron of 1.89 +/- 0.01 Ga (MSWD = 26; Hegner and Bevier, 1991), which is significantly younger than the U-Pb zircon ages for this unit of 2.04 - 1.96 Ga (Parrish, 1989; Machado et al., 1993). Hegner and Bevier (1991) suggest that the thermal event associated with collisional tectonics in the Cape Smith Belt is the reason for this discrepancy, resulting in perturbation of the Pb-isotopic system. The discrepancies between the various isotopic systems may be due to the fact that samples from different localities were chosen for isotopic and geochemical analyses, and cannot be directly correlated. In fact, for most of the samples analyzed from the Cape Smith Belt, isotope results and trace element data are not available on the same samples making realistic modelling of contamination processes impossible for the present.

The alkaline lava suite in the upper Povungnituk Group has highly fractionated REE's ($(La/Yb)_n$ = 14.2 - 45.8; Fig. 4) and trace element concentrations comparable to modern intra-plate alkaline magmas (Gaonac'h et al., 1989, 1992; Picard et al., 1990) such as those from the Massif Central (Chauvel and Jahn, 1984) or Fort Selkirk, (Francis and Ludden, 1990) which are interpreted as being derived from enriched mantle sources. The alkaline volcanics of the Povungnituk Group have higher TiO_2 (up to 6.8 wt%), Nb (75-400 ppm) and Zr (up to 1700 ppm) concentrations relative to modern suites, which may be characteristic of Early Proterozoic enriched mantle reservoirs (Gaonac'h et al., 1992). These lavas have an ε_{Nd} isotopic composition of +1.8 to +2.6 (at 1.96 Ga), and low Sm/Nd for both mafic and felsic lavas (Gaonac'h et al., 1989, 1992; Fig. 3), which is intermediate between those of MORB and bulk earth, occupying a position similar to most modern OIB reservoirs.

The Eskimo Formation of the Belcher Islands is characterized by moderate Zr/Nb ratios (9.2-20.4), pronounced negative Nb anomalies ($(Nb/La)_n$ = 0.3-0.5) and fractionated LREE's ($(La/Sm)_n$ = 2.4 - 3.3; Legault et al., in press; Fig. 5). These lavas are geochemically very similar to those of the western Povungnituk Group, and are also interpreted as crustally contaminated continental tholeiitic basalts (Arndt et al., 1987;

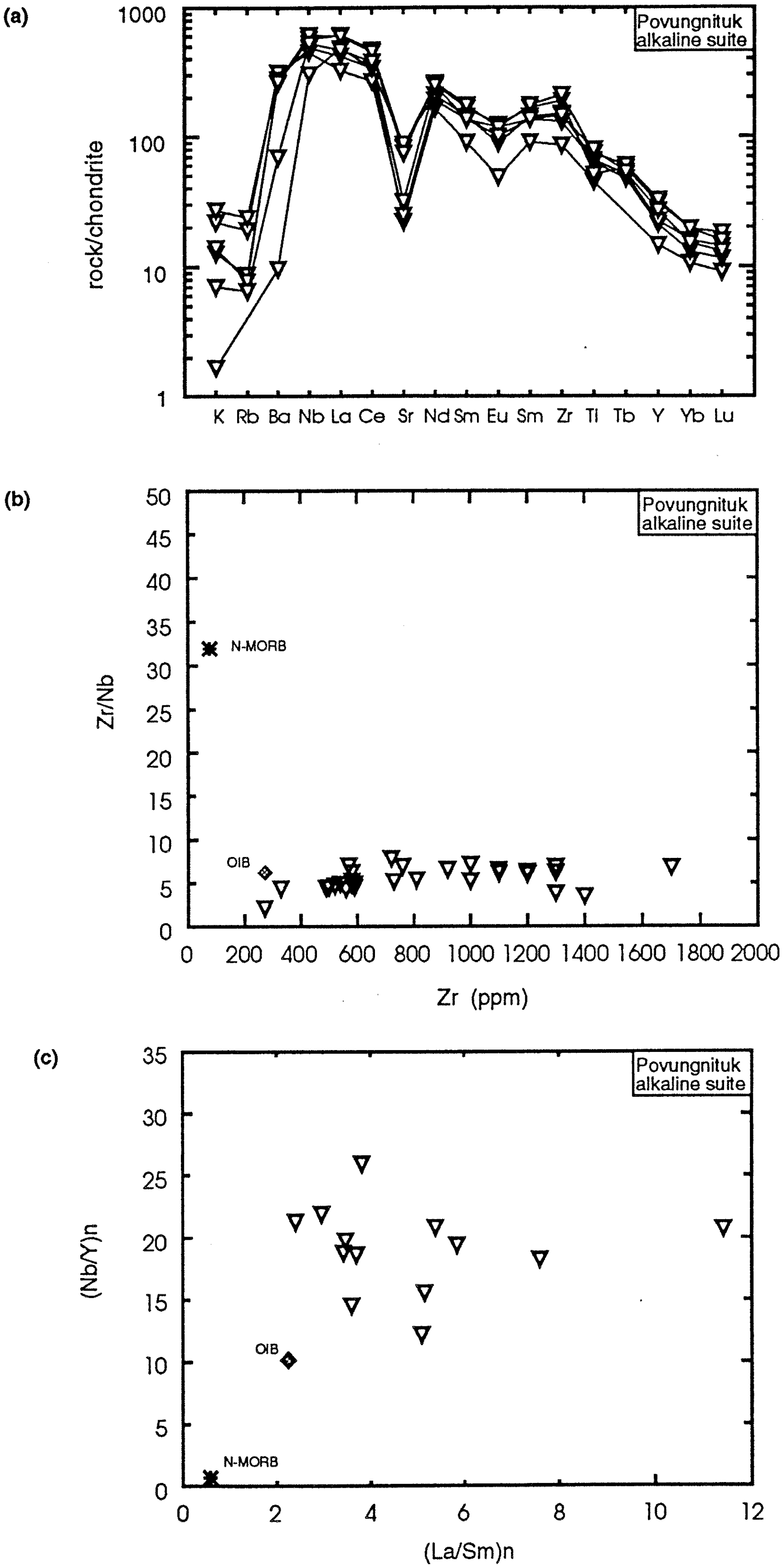


Figure 2.4. (a) Chondrite normalized trace element variation diagram and other geochemical variation diagrams (b) and (c) for the Povungnituk alkaline suite (Gaonac'h et al., 1992; Francis, unpub. data). Normalization factors and average compositions of N-MORB, E-MORB and OIB from Sun and McDonough (1989).

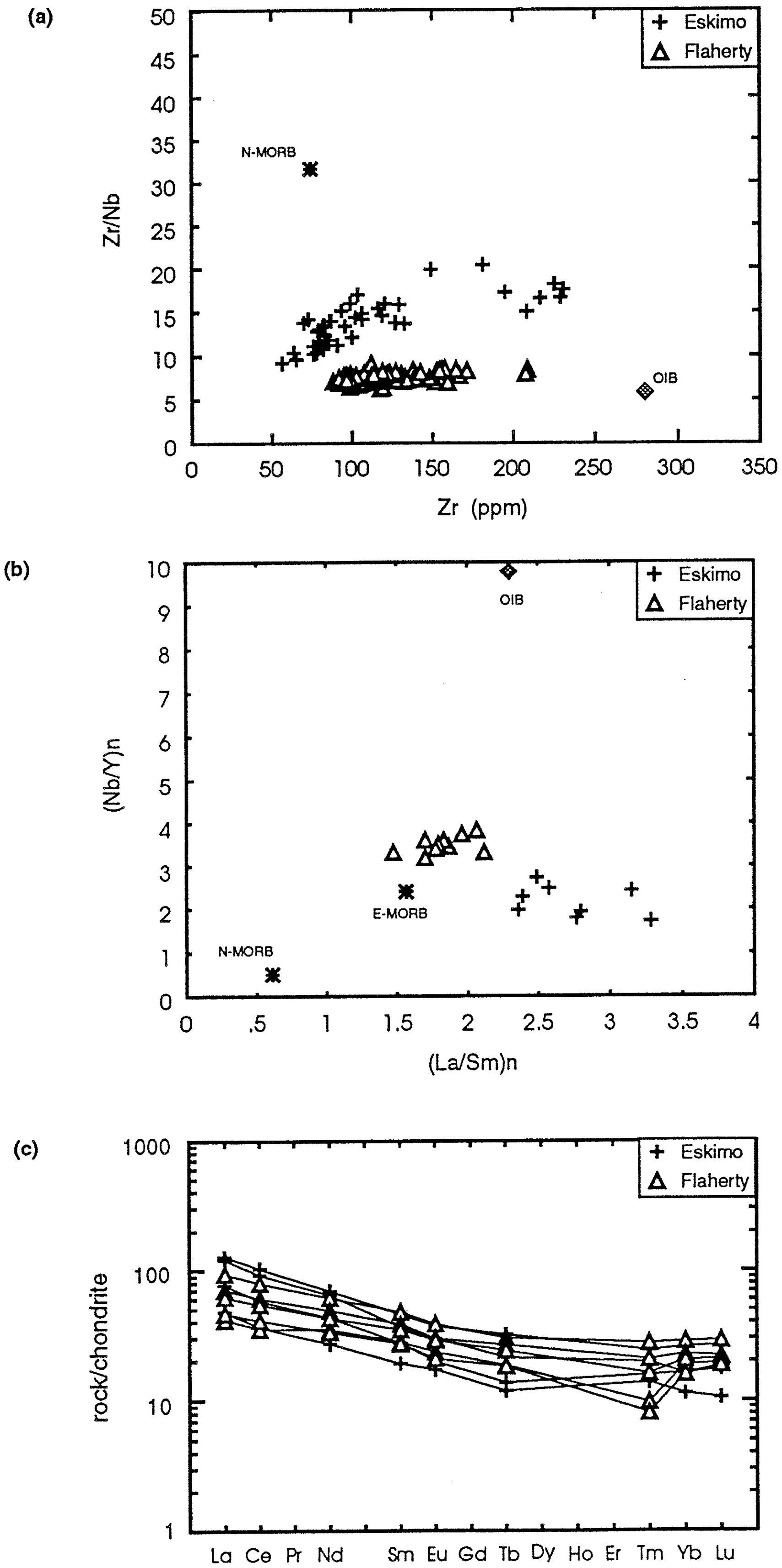


Figure 2.5. (a) and (b) Geochemical variation diagrams and REE diagram (c) for the Eskimo and Flaherty Formations. Data from Legault et al., (in press) and unpub. data. Note the similarity between the position of the Eskimo Formation and Flaherty Formation samples relative to the western and eastern Povungnituk Group samples respectively (Fig. 2). Normalization factors and average compositions of N-MORB, E-MORB and OIB from Sun and McDonough (1989).

Baragar and Scoates, 1987; Legault et al., in press). Petrogenetic modelling indicates that these basalts can be derived from a picritic parent via contamination and assimilation-crystal fractionation processes (Arndt et al., 1987; Legault et al., in press). The overlying Flaherty Formation lavas are similar to the central and eastern Povungnituk lavas in that they also contain negligible Nb anomalies, slightly enriched LREE's ($(La/Sm)_n = 1.5 - 2.1$) and low Zr/Nb ratios (Arndt et al., 1987; Legault et al., in press) and may have been derived from the same enriched mantle source which subsequently underwent open-system crystal fractionation (Legault et al., in press). The geochemical features of the Eskimo and Flaherty Formations are illustrated in Fig. 5 and representative analyses are presented in Table 1.

Contamination by continental crust is clearly responsible for the isotopic signature of the Eskimo Formation basalts (Fig. 3) which have $\varepsilon_{Nd(1.96\ Ga)}$ values of -6.0 to -7.4, which contrasts with the relatively uncontaminated Flaherty Formation lavas ($\varepsilon_{Nd(1.96\ Ga)}$ = +2.4 to +4.3; Chauvel et al., 1987). The calculated Th/U ratio of 6.7 for the magmas and 5.4 for the source of the Eskimo Formation basalts combined with a μ^* of 8.22 led Arndt and Todt (in press) to conclude more specifically that lower continental crust was the contaminant for these basalts. Consideration of ε_{Nd}, model μ^* values, as well as trace element geochemistry for the Flaherty Formation lavas, suggests that these rocks were derived from an enriched source, similar to that of modern plume-related basalts (Arndt and Todt, in press).

2.4. A 2.0 - 1.9 Ga OCEANIC RIFT SEQUENCE

2.4.1. Geological features

Structurally overlying and in thrust contact with the Povungnituk Group is the predominantly igneous Chukotat Group (ca. 1.92 Ga; Parrish, 1989) consisting of >5 km of submarine pillowed and massive basalt flows and mafic/ultramafic sills with minor associated sediments (Francis and Hynes, 1979; Hynes and Francis 1982; Francis et al., 1983; Picard et al., 1990; St-Onge and Lucas, 1990b, 1993). The Chukotat volcanics range in composition from Mg-rich komatiitic basalt with microspinifex textures (olivine-phyric basalt) to low-Mg plagioclase-phyric basalt and have been interpreted as recording the change from accumulation of continental rift-type volcanics to the formation of transitional to oceanic-type crust (Francis and Hynes, 1979; Francis et al., 1981; Hynes and Francis, 1982; Francis et al., 1983; Picard et al., 1990). Low-pressure crystal fractionation of observed phenocryst phases can explain the compositional variation of the Chukotat volcanics from olivine- to clinopyroxene- to plagioclase-phyric basalts (Francis et al., 1983).

Subaqueous komatiitic basalts occurring as massive and layered flows similar to some of the Chukotat basalts are also found in the Ottawa Islands of Hudson Bay (Baragar and Lamontagne, 1980; Baragar and Scoates, 1981; Arndt, 1982; Arndt et al., 1987). A Pb-Pb age of 1800 +/- 80 Ma was obtained on samples of the Ottawa Island volcanics (Arndt and Todt, in press), however, it is believed that this age does not represent the crystallization age of the rocks, but rather a later alteration or metamorphic event. Consequently, an age of 1.92 Ga (the age of the Chukotat Group; Parrish, 1989) has been assumed for the Ottawa Island samples.

Tectonically juxtaposed to the Chukotat Group in the Cape Smith Belt is the Watts Group which comprises a distinct assemblage of layered mafic and ultramafic rocks, massive and pillowed basalt flows, mafic sills and sheeted dykes and rare plagiogranite intrusions that have been interpreted as an ophiolite with a preserved thickness of 7.5-8 km (St-Onge et al., 1988; St-Onge and Lucas, 1990b, 1993; Scott et al., 1991, 1992). U-Pb analysis of zircons from gabbroic layers within the Watts Group yielded an age of 1998 +/-2 Ma (Parrish, 1989).

2.4.2 Geochemical and isotopic characteristics

The geochemical transition from continental rifting to oceanic rift magmatism can be seen in the lowermost sequences of the Chukotat Group volcanics which are slightly LREE enriched (Picard, 1989a; Picard et al., 1990) compared to the more enriched continental tholeiites of the Povungnituk Group and to the overlying depleted to slightly fractionated ($(La/Sm)_n$ = 0.7-2.0) Chukotat Group basalts (Fig. 6a; Francis et al., 1983; Picard, 1989a; Picard et al., 1990).

The mafic volcanics and komatiitic basalts of the Chukotat Group and most samples from the Ottawa Islands (MgO up to 19 wt%) as well as one population of Watts Group volcanics are similar to modern MORB in terms of trace element contents (e.g., relatively unfractionated REE's, low $(Nb/Y)_n$ ratios, high Zr/Nb). These features are presented in Fig. 6 and Tables 1 and 2 (Francis et al., 1983; Arndt et al., 1987; Picard, 1989a, 1989b; Picard et al., 1990; Scott et al., 1991). However, several important differences between modern MORB and these Early Proterozoic depleted oceanic magmas are evident, as the primitive magmas of the Early Proterozoic suite are distinctly more Fe- and Si-rich (Fig. 7) and Al-poor compared to modern primitive MORB magmas. In addition, the crystallization sequence determined for the Chukotat lavas of olivine followed by clinopyroxene, and then plagioclase is different from that of modern MORB in which olivine is followed by plagioclase and then clinopyroxene (Green et al., 1979; Basaltic Volcanism Study Project, 1981).

In terms of Nd isotopes, the most primitive rocks of the Ungava orogen are found in the MORB-like Watts Group rocks, which have high initial ε_{Nd} values of +3.9 to +6.1 at 2.0 Ga (Fig. 3; Hegner and Bevier, 1991), overlapping with the depleted mantle composition of Jahn et al. (1987, 1988). Pb isotopic analysis of samples from the Watts Group has produced a precise Pb-Pb isochron at 1.97 +/- 0.04 Ga (MSWD = 1.5; Hegner and Bevier, 1991), which is identical (within error) to the U-Pb zircon age. The Watts Group has a μ^* value of 8.00 and a calculated Th/U ratio of its source of 3.7, which are similar to mantle values (Hegner and Bevier, 1991). The Chukotat Group volcanics and some of the Ottawa Island basalts have $\varepsilon_{Nd(1.92\ Ga)}$ compositions ranging from +2.7 to +4.5 (Fig. 3) and a calculated μ^* = 8.09 (Zindler, 1982; Chauvel et al., 1987; Smith and Ludden, 1989; Hegner and Bevier, 1991; Arndt and Todt, in press). Other Ottawa Island komaitiites are isotopically more enriched (Fig. 3) with $\varepsilon_{Nd(1.92\ Ga)}$ values ranging from +0.3 to +0.9 (Chauvel et al., 1987) and μ^* = 8.33 (Arndt and Todt, in press). In

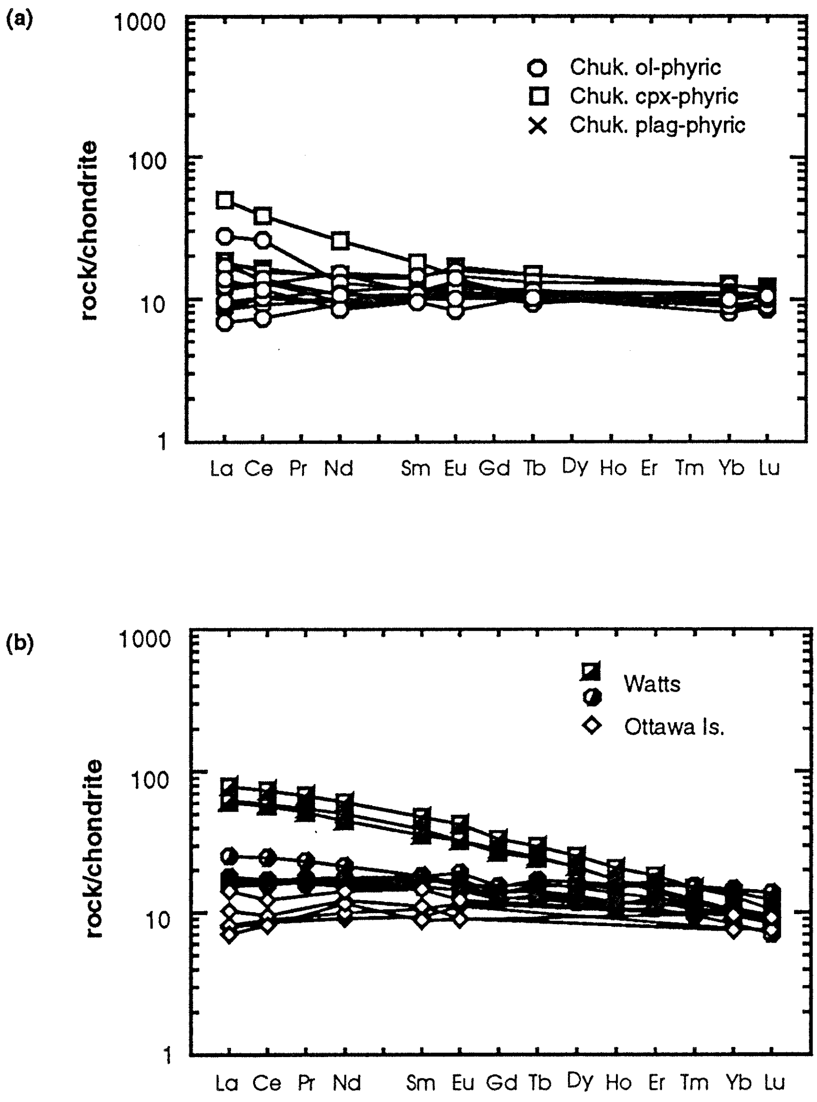
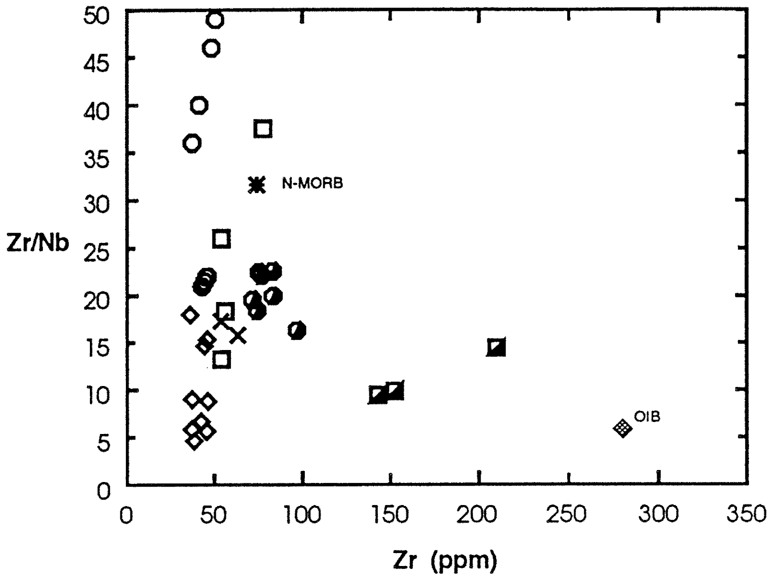
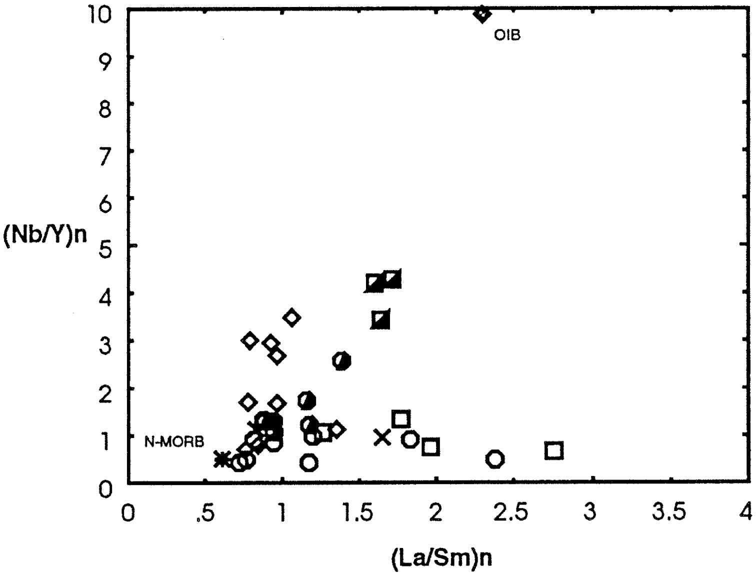


Figure 2.6: REE diagrams for representative samples from the (a) Chukotat Group; (b) Watts Group and Ottawa Island samples; (c) Zr/Nb vs Zr; and (d) $(Nb/Y)_n$ vs $(La/Sm)_n$ for same samples. Note the trend of some Watts Group samples toward enriched mantle compositions. Data compiled from Francis et al. (1983), and unpub. data; Arndt et al. (1987); Picard (1989b); and Scott et al. (1991).

(c)


Zr/Nb
Zr (ppm)
N-MORB
OIB
0
5
10
15
20
25
30
35
40
45
50
0
50
100
150
200
250
300
350

(d)


(Nb/Y)n
(La/Sm)n
N-MORB
OIB
0
1
2
3
4
5
6
7
8
9
10
0
.5
1
1.5
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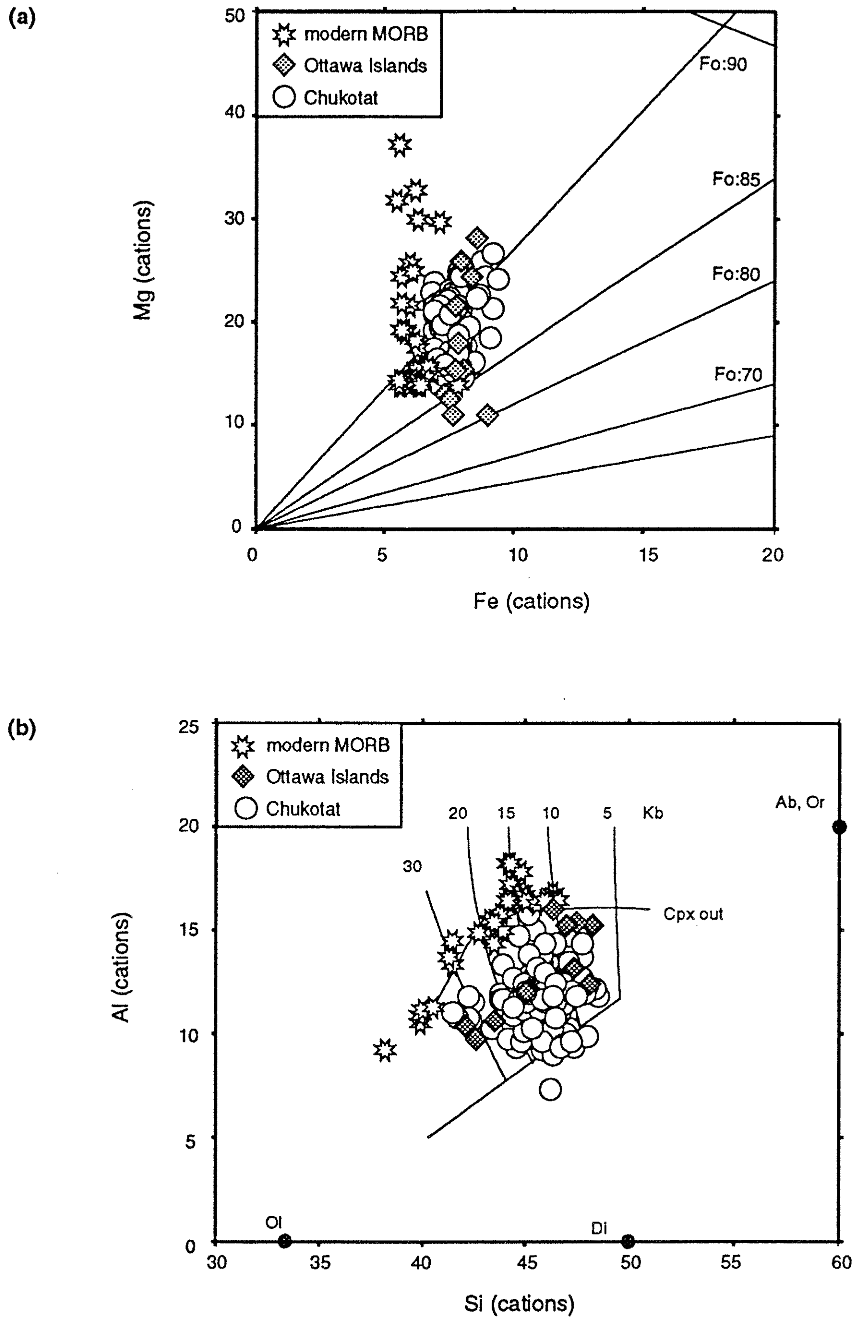

Figure 2.7: (a) Mg vs Fe and (b) Al vs Si (cation units) for samples from the Chukotat Group and the Ottawa Islands (data sources as in Fig. 6), with modern MORB for comparison (data from Blanchard et al., 1976; Jakobsson et al., 1978; Zindler et al., 1979; Bryan et al., 1981; Le Roex et al. 1981; Desonie and Duncan, 1990). MORB-like samples from the Ungava orogen are more Fe-rich and Al-poor compared to modern MORB.

addition, Arndt et al. (1987) report that these samples contain fractionated REE's (La_n/Yb_n = 1.5-4.0) and have negative Nb and Ta anomalies. These authors have all concluded that these Ottawa Island komaitiites have been contaminated by continental crust during their emplacement (Arndt et al., 1987; Chauvel et al., 1987; Arndt and Todt, in press).

On the basis of their geochemical and isotopic signature a second group of Watts volcanics can be delimited. These volcanics have fractionated REE profiles, higher $(Nb/Y)_n$, lower Zr/Nb (Fig. 6; Scott et al., 1991) and a lower initial ε_{Nd} of +3.0 to +3.4 (Fig. 3; Hegner and Bevier, 1991), relative to the previously mentioned MORB-like group. These two different groups are indistinguishable in outcrop and do not show consistent cross-cutting relationships (Scott et al., 1991). This second group of Watts samples have a geochemical signature that is very similar to that of the Flaherty Formation and the eastern Povungnituk Group (e.g. compare Watts Group samples in Fig. 6 with Figs. 2 and 5). The geochemical characteristics of these Watts Group samples may be attributed to either contamination of mantle-derived magmas or to derivation from an enriched mantle source. The lack of continental crust in the area and the absence of negative Nb anomalies argues against the role of contamination in the petrogenesis of these basalts. The close spatial relationship of the two groups suggests the synchronous existence and tapping of two distinct mantle reservoirs (depleted MORB-like and enriched OIB-like; Scott et al., 1991, 1992; Hegner and Bevier, 1991).

2.5 A 1.9 to 1.8 Ga ARC SEQUENCE

2.5.1 Geological features

A series of intrusive and supracrustal rocks found north of the Cape Smith Belt comprise the Narsajuaq arc which was accreted to the Superior Province basement ca. 1.83-1.82 Ga (St-Onge and Lucas, 1990a; Lucas et al., 1992; St-Onge et al., 1992). The plutons range in composition from pyroxenite to syenogranite, although the principal unit is a well layered sequence of tonalite and quartz diorite with cross-cutting monzogranite to syenogranite veins (St-Onge and Lucas, 1992). Discrete km-size bodies of quartz diorite, tonalite, and granite are intrusive into the layered tonalite-quartz diorite suite. The plutonic suite spans more than 70 Ma with the oldest pluton dated at 1.898 Ga (Parrish, 1989; St-Onge et al., 1992). Supracrustal rocks (Sugluk Group) consist predominantly of graphitic semi-pelites with lesser pelite, quartzite, calc-silicate and mafic volcanic(?) bands and make up approximately 20% of the exposed arc series (St-Onge and Lucas, 1992).

Sandwiched between the Watts Group and the Chukotat Group in the western portion of the Cape Smith Belt is the Parent Group, consisting of volcanoclastic tuffs and basaltic to rhyolitic lavas (basaltic andesite being the most common), intercalated with shales, siltstones and greywackes (Picard et al., 1990; Tremblay, 1991). A rhyolite of the Parent Group has recently been dated at 1.86 Ga by Machado et al. (1993).

2.5.2 Geochemical and isotopic characteristics

Geochemically the Narsajuaq magmas are predominantly calc-alkaline and regular trends of decreasing Fe_2O_3, MgO, CaO and TiO_2 with increasing SiO_2 are observed (Dunphy and Ludden, 1992). High concentrations of Ba and Th (up to 2800 and 14 ppm, respectively) and other LILE's, pronounced negative Nb and Ti anomalies, and variably fractionated REE's ($(La/Sm)_n$ = 1.6 - 7.4; $(La/Yb)_n$ = 2.2 - 73.3) are characteristic of these plutonic rocks (Table 1; Dunphy and Ludden, 1992). The most mafic plutons (approximately 45-48 wt% SiO_2, 6.9-9.0 wt% MgO) have Zr/Nb ratios ranging from 6.2 to 16.8 and variable $(Nb/Y)_n$ values of 1.8 to 5.1. The major and trace element chemistry of the Narsajuaq terrane rocks is similar to that previously reported for other volcanic arc granites (Fig. 8). In addition, the overall composition and volume of the granitoid rocks of the Narsajuaq plutonic complex is analogous to that exposed in the massive Mesozoic batholiths located along the western margin of North and South America from Alaska to Antarctica (Brown, 1982; Mahlburg Kay and Rapela, 1990). The Narsajuaq rocks have been interpreted as the plutonic core of a magmatic arc (Dunphy and Ludden, 1992; Lucas et al., 1992; St-Onge et al., 1992).

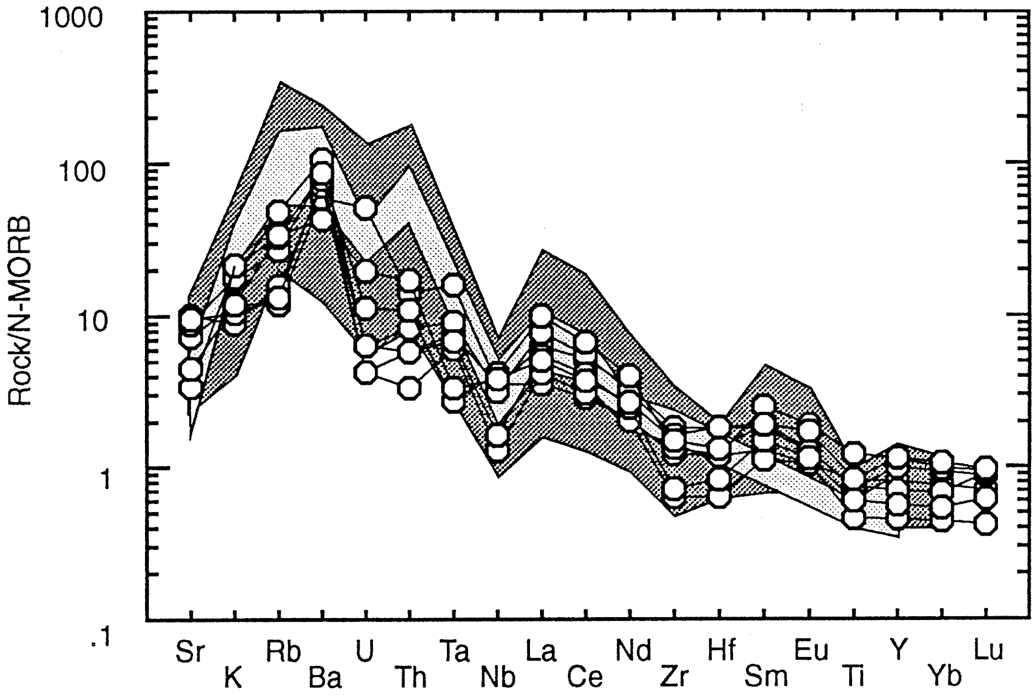

Figure 2.8: N-MORB normalized trace element variation diagram for representative mafic Narsajuaq arc samples (open circles; Dunphy and Ludden, in preparation) with fields representing data from the Peninsular Ranges Batholith (light shading; Silver and Chappell, 1988) and the Sunda-Banda arc (dark shading; Varne and Foden, 1986; Wheller et al., 1987; Stolz et al., 1990) for comparasion.

The Parent Group lavas are subalkaline and metaluminous and have both calc-alkaline and tholeiitic affinities (Picard et al., 1990). The volcanics have slightly fractionated REE profiles (La_n/Yb_n = 2.4-5.5) and have negative Nb anomalies (Picard, in preparation). The lithological and geochemical characteristics of the Parent Group volcanics/volcanoclastics led Picard et al. (1990) to suggest that these rocks were also emplaced in a magmatic arc environment, although unusually high concentrations of Nb (25-29 ppm) are reported for these rocks (Picard et al., 1990). The similar overall character and age of the Narsajuaq terrane plutonics and the Parent Group volcanics as well as their tectonic juxtaposition has led to the suggestion that the Parent Group may represent a volcanic edifice of the Narsajuaq arc (St-Onge et al., 1992; Dunphy and Ludden, in preparation).

Initial $^{87}Sr/^{86}Sr$ and $\varepsilon_{Nd(t)}$ values for the Narsajuaq magmas range from depleted (0.7020 and +4.0) to very enriched (0.7048 and -10.7), with the depleted values overlapping with those values reported for volcanics from within the adjacent Cape Smith Thrust Belt (Fig. 3; Dunphy and Ludden, 1994). The isotopic signature of the Narsajuaq plutons indicate an origin involving contamination of depleted mantle melts ($\varepsilon_{Nd} \approx +5$) by both crustal and subduction-related components (Dunphy and Ludden, in preparation).

2.6 DISCUSSION

The common association of voluminous eruption of continental flood basalts with rift margins has been previously documented and the inferred role of mantle plumes in the development of such magmas is hotly debated (e.g., Morgan, 1971; White and McKenzie, 1989; Campbell and Griffiths, 1990; Duncan and Richards, 1989). The relative contribution or predominance of a particular mantle source(s) (i.e. asthenosphere, lithosphere, lower mantle) in the geochemical signature of these magmas is currently a subject of contention (Gallagher and Hawkesworth, 1992; Griffiths and Campbell, 1990; Arndt and Christensen, 1992; Deniel et al., 1994). In the continental rifting stage magmas commonly display evidence for crustal contamination (e.g., Gariepy et al., 1983;

Carlson, 1984; Cox and Hawkesworth, 1984) involving upper and/or lower crust rocks. In addition, alkaline igneous rocks are also a feature of volcanic rifted margins (e.g., North Atlantic rifted margin, Larsen and Watt, 1985; Cameroon line, Fitton, 1987), representing small degree partial melts of the enriched mantle within the plume or enriched mantle from the subcontinental lithosphere.

The Ungava orogen displays all the geological components of a modern Wilson cycle, and represents one of the best studied Proterozoic orogens from both a geological and petrological perspective. Based on the extensive geochemical data base for the orogen we have evaluated the composition of various mantle reservoirs parental to the magmas of the orogen at 2.0 Ga (Tables 2 and 3).

2.6.1 Rifted continental margin magmas

Given the lithological associations, the massive volume of erupted magma (>6.5 km thick; St-Onge and Lucas, 1990b), as well as the temporal (Eskimo and Flaherty Formations) and spatial (east and west Povungnituk Group) association of transgressive sequences of relatively depleted magmas and clearly contaminated magmas, it has been proposed that these units represent a volcanic rifted continental margin (Hynes and Francis, 1982; Francis et al., 1983; Arndt et al., 1987; Picard et al., 1990; St-Onge and Lucas, 1990b, 1993). These sequences bear many of the geochemical and isotopic characteristics of modern continental rifted margins such as crustally contaminated continental basalts (i.e. the Eskimo Formation and western Povungnituk Group) and enriched mantle material (manifest in the Flaherty Formation and eastern Povungnituk Group lavas as well as the Povungnituk alkaline suite).

The rifting event that led to the formation of the lavas found in the Ungava orogen may have been contemporaneous with the arrival of a mantle plume (Hegner and Bevier, 1991; Scott et al., 1991). Melting within the mantle plume could have produced the enriched mantle signature of the magmas of the Flaherty Formation, the eastern Povungnituk Group and the low ε_{Nd} population (+3.0 to +3.4) of the Watts Group. The overall similarity in composition of rocks from these three units suggests that they may be petrogenetically related and may have been derived from the same mantle source. The average Zr/Nb and $(Nb/Y)_n$ ratios of these lavas (approximately 8.9 and 4.4, respectively) are comparable to E-MORB compositions (Table 2). As seen in Figure 3, most Povungnituk and all the Flaherty samples trend toward a component with an isotopic composition of approximately +2 and low Sm/Nd (0.1) defined by the Povungnituk alkaline suite. A realistic model for the petrogenesis of these lavas could involve this component as part of a mantle plume or as an enriched part of the subcontinental lithosphere. Contamination of similar magmas through interaction with continental crust appears to be responsible for the geochemical (Table 2) and isotopic signature (Fig. 3) of the Eskimo Formation and western Povungnituk Group samples (Arndt et al., 1987; Baragar and Scoates, 1987; Chauvel et al., 1987; Arndt and Todt, in press; Legault et al., in press). The incompatable element enriched alkaline rocks found in the upper Povungnituk Group could also represent melts derived from such an enriched source. The geochemical signature of these rocks is similar to that of modern highly alkaline basalts and to average OIB compositions (Fig. 4; Table 2). Small degrees of partial melting of the sub-continental lithospheric mantle could produce such chemical signatures. These Povungnituk alkaline rocks as well as other rocks having geochemical characteristics indicating a contribution of enriched mantle in their petrogenesis allow us to define an enriched mantle reservoir at 2.0 Ga having an ε_{Nd} composition of approximately +2.5 to +3.5 and a model μ^* value of approximately 8.3 (Table 3).

The entire spectrum of the rifted margin magmas in terms of trace elements and isotopic compositions, from slightly enriched basalts (Povungnituk Group and Flaherty Formation) to highly enriched lavas (Povungnituk alkaline suite) to crustally contaminated tholeiites (Eskimo Formation and some Ottawa Island samples) is entirely consistent with the range of magmas produced in the North Atlantic rifted margin during the lower Tertiary (e.g., Larsen et al., 1989).

Table 3. Isotopic reservoirs in the Ungava orogen recalculated to 2.0 Ga.

Source	$\varepsilon_{Nd(2.0\ Ga)}$ [1]	$^{147}Sm/^{144}Nd$ [1]	μ^* [2]	Th/U [2]
Depleted mantle	+4.5 to +5.5	>0.19	8.0-8.1	3.7
Enriched mantle	+2.5 to +3.5	>0.09 to 0.11	8.3	5.4
Arc source	+4.5 to +5.5	>0.09 to 0.14	≈ 8 [3]	≈ 3.8 [3]
Bulk crust	-7.5 to -16	0.07 to 0.12	7-10	4-15

[1] Nd compositions from this compilation (see table 2 for references); bulk crust Nd composition from Dunphy (unpub. data).

[2] μ^* and Th/U values from Arndt and Todt (in press) and Hegner and Bevier (1991).

[3] Approximation based on the composition of arc rocks from the Reindeer Lake zone of the Trans Hudson orogen (Arndt and Todt, in press).

[4] Represents the range of compositions from upper crust (Taylor and McLennan, 1985) to lower crustal granulites (Rudnick and Presper, 1990).

2.6.2 Depleted oceanic magmas

The oceanic basalts of the Watts Group, Chukotat Group and Ottawa Islands basalts were produced by melting of depleted mantle material. These rocks were generated upon complete rupture of the continental crust, resulting in the development of an oceanic basin with trace element and isotopic signatures characteristic of modern depleted basalts (Fig. 3 and Table 2). However, certain factors such as the high Fe and low Al of the komatiitic basalts, as well as the crystallization sequence of olivine followed by clinopyroxene, and then plagioclase indicates that these lavas are not identical to modern MORB. The Fe-rich and Al-poor nature of these lavas may reflect a higher potential temperature in the mantle during the Proterozoic and thus a deeper depth of initiation for melting and higher extents of melting (Klein and Langmuir, 1987). Alternatively, the mantle source for these magmas could have been more depleted and more Fe-rich than that of modern MORB. Samples from these sequences represent the best estimate of the Early Proterozoic depleted mantle reservoir for the Ungava region, which we here define as having an ε_{Nd} composition of $\approx +4.5$ to $+5.5$ and a model μ^* of approximately 8.0 (Table 3).

The Watts Group appears to represent the crustal portion of an ophiolite (Purtuniq ophiolite; St-Onge et al., 1988; Scott et al., 1989, 1991). The Purtuniq ophiolite differs from Phanerozoic ophiolites, however, in the following ways (Scott et al., 1991): (1) it has a thicker mafic volcanic sequence (preserved crustal thickness of 7.5-8 km); (2) it records the mutual presence of OIB-like and MORB-like dykes and volcanics; (3) there is a large age difference between the formation and subsequent emplacement of the unit (approximately 130 Ma); and (4) there is a lack of tectonic periodite representing the upper mantle. The thickness of the ophiolite may, however, be in accord with the proposed thicker oceanic crust during the early history of the Earth (Sleep and Windley, 1982). Although the presence of the both MORB-like and OIB-like dykes is consistent with observed relationships in modern oceanic crust (e.g., DSDP hole 504B, Emmermann, 1985; Tual et al., 1985) it has only rarely been observed in ophiolites (e.g. the East Taiwan ophiolite; Jahn, 1986). The late age of accretion for the Watts ophiolite has important tectonic consequences and may indicate an origin in a long-lived oceanic basin as opposed to modern ophiolites which appear to develop predominantly in back-arc basins (Dewey, 1976; Saunders et al., 1979).

2.6.3 Subduction related magmas

Destruction of the oceanic basin through subduction generated the arc sequences of the Narsajuaq terrane and the Parent Group. The least contaminated mafic rocks of the predominantly plutonic Narsajuaq terrane have an isotopic composition close to that estimated for the depleted mantle (Fig. 3 and Table 3). The large range in isotopic composition and the variable geochemical signature of the entire spectrum of arc rocks indicates that continental crust was important in their petrogenesis. A previous model for the Narsajuaq arc rocks based on preliminary geochemical and isotopic data suggested that these rocks were generated entirely in an oceanic setting (Lucas et al., 1992; St-Onge et al., 1992). An updated model incorporating new data involves northward-directed subduction under a thick crust consisting of transitional rifted to continental crust and can account for the observed geochemical and isotopic characteristics of the arc magmas (Dunphy and Ludden, in preparation). A more complete discussion of the petrogenesis of the plutonic rocks of the Narsajuaq terrane will be presented in a forthcoming paper (Dunphy and Ludden, in preparation).

2.7 SUMMARY

The magmatic suites of the Ungava orogen and associated islands preserve unique evidence for the existence of Early Proterozoic Wilson-cycle tectonics. Rifting of the Superior Province craton (synchronous with, or caused by, hotspot activity), resulted in continental breakup ca. 2.04 Ga and flood basalt magmatism leading to the development of oceanic crust which was subsequently destroyed during subduction and arc magmatism. As shown in the summary tables (Tables 2 and 3), the geochemical reservoirs involved in these magmatic events are geochemically and isotopically similar to modern equivalents, indicating a continuity in magmatic processes through time. The presence of these reservoirs at 2.0 Ga indicate that MORB, OIB and sub-lithospheric mantle reservoirs were isolated at this time and were sampled by melting in response to tectonic regimes similar to those of modern Wilson cycles (e.g., rifting, hotspot activity). In addition, the presence of an OIB source and/or an enriched subcontinental lithosphere suggest that mantle enrichment processes during the Early Proterozoic were essentially identical to those now in operation and is therefore strongly suggestive of an active crust-mantle recycling system associated with plate tectonics prior to 2.0 Ga.

ACKNOWLEDGEMENTS

This paper is the result of many years of study by a variety of workers in the Ungava orogen and associated areas, particularly by the Geological Survey of Canada and the Ministère de l'Energie et des Ressources du Quebec. Discussions with N. Arndt, A. Hynes, F. Legault, S.B. Lucas, D. Scott and M.R. St-Onge helped formulate some of the ideas presented. The generous use of unpublished data from F. Legault and H. Gaonac'h is greatly appreciated. JMD acknowledges an NSERC graduate fellowship and field-based support from the Geological Survey of Canada; JNL and DF acknowledge financial support from NSERC Canada and FCAR Quebec. D. Lambert and an anonymous reviewer are thanked for their constructive comments on the original manuscript.

REFERENCES

Arndt, N.T., 1982. Proterozoic spinifex-textured basalts of Gilmour Island, Hudson Bay. In: Current Research, Part A. Geol. Surv. Can. Pap., 82-1A, pp. 137-142.

Arndt, N.T. and Christensen, U., 1992. The role of lithospheric mantle in continental flood volcanism: thermal and geochemical constraints. J. Geophys. Res., 97:10967-10981.

Arndt, N.T. and Todt, W., in press. Formation of Canadian 1.9 Ga old Trans-Hudsonian continental crust: (II) Pb isotopic data. Chem. Geol.

Arndt, N.T., Brugmann, G.E., Lehnert, K., Chauvel, C. and Chappell, B.W., 1987. Geochemistry, petrogenesis and tectonic environment of Circum-Superior Belt basalts, Canada. In: T.C. Pharaoh, R.D. Beckinsale and D. Rickard (Editors), Geochemistry and Mineralization of Proterozoic Volcanic Suites. Geol. Soc., Spec. Pub. No. 33, pp. 133-145.

Baragar, W.R.A. and Lamontagne, C.G., 1980. The Circum-Ungava Belt in eastern Hudson Bay: the geology of Sleeper Islands and parts of the Ottawa and Belcher Islands. In: Current Research, Part A. Geol. Surv. Can. Pap., 80-1A, pp. 89-94.

Baragar, W.R.A. and Scoates, R.F.J., 1981. The Circum-Superior Belt: a Proterozoic plate margin? In: A. Kroner (Editor), Precambrian Plate Tectonics. Elsevier, New York, pp. 297-330.

Baragar, W.R.A. and Scoates, R.F.J., 1987. Volcanic geochemistry of the northern segments of the Circum-Superior Belt of the Canadian Shield. In: T.C. Pharaoh, R.D. Beckinsale and D. Rickard (Editors), Geochemistry and mineralization of Proterozoic Volcanic suites. Geol. Soc., Spec. Pub., No. 33, pp. 113-132.

Basaltic Volcanism Study Project, 1981. Basaltic Volcanism on the terrestrial planets. Pergamon Press, New York, pp. 1286.

Bell, R., 1885. Observations on the geology, mineralogy, zoology and botany of the Labrador Coast, Hudson Strait and Bay. Geol. Surv. Can., Report of Progress, 1882-84.

Blanchard, D.P., Rhodes, J.M., Dungan, M.A., Rogers, K.V., Donaldson, C.H., Brannon, J.C., Jakobs, J.W. and Givson, E.K., 1976. The chemistry and petrology of basalts from leg 37 of the Deep Sea Drilling Project. J. Geophys. Res., 81:4231-4246.

Brown, G.C., 1982. Calc-alkaline intrusive rocks: their diversity, evolution and relation to volcanic arcs. In: R.S. Thorpe (Editor), Andesites. John Wiley and Sons, New York, pp. 437-464.

Bryan, W.B., Thompson, G. and Ludden, J.N., 1981. Compositional variation in normal MORB from 22-25 N: Mid-Atlantic Ridge and Kane Fracture Zone. J. Geophys. Res., 86:11815-11836.

Campbell, I.H. and Griffiths, R.W., 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth Planet. Sci. Lett., 99:79-93.

Campbell, I.N. and Griffiths, R.W., 1992. The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. J. Geol., 92:497-523.

Carleson, R.W., 1984. Isotopic constraints on Columbia River flood basalt genesis and the nature of the subcontinental mantle. Geochim. Cosmochim. Acta., 48:2357-2372.

Chauvel, C. and Jahn, B.M., 1984. Nd-Sr isotope and REE geochemistry of alkali basalts from the Massive Central, France. Geochim. Cosmochim. Acta., 48:93-110.

Chauvel, C., Arndt, N.T., Kielinzcuk, S. and Thom, A., 1987. Formation of Canadian 1.9 Ga old continental crust. I: Nd isotopic data. Can. J. Earth Sci., 24:396-406.

Cox, K.G. and Hawkesworth, C.J., 1984. Relative contributions of crust and mantle to flood basalt magmatism, Mahabaleshwar area, Deccan Traps. Philos. Trans. R. Soc. London, A310:627-641.

Deniel, C., Vidal, P., Coulon, C., Vellutini, P.J. and Piguet, P., 1994. Temporal evolution of mantle sources during continental rifting: the volcanism of Djibouti (Afar). J. Geophys. Res., 99:2853-2869.

Desonie, D. and Duncan, R.A., 1990. The Cobb-Eickelberg Seamount Chain: Hotspot volcanism with mid-ocean ridge affinity. J. Geophys. Res., 95:12697-12711.

Dewey, J.F., 1976. Ophiolite obduction. Tectonics, 31:93-120.

Duncan R.A., and Richards, M.N., 1989. Hotspots, mantle plumes, flood basalts and true polar wander. Rev. Geophys., 29:31-50.

Dunphy, J.M. and Ludden, J.N., 1992. The Narsajuaq terrane, northern Quebec: a mid-crustal section of an Early Proterozoic magmatic arc. Eos (Trans. Am. Geophys. Union), 73:318 (abstract).

Dunphy, J.M. and Ludden, J.N., 1994. Geochemical and isotopic characteristics of an Early Proterozoic accreted magmatic arc (Narsajuaq terrane) northern Quebec, Canada. Eighth international conference on geochronologgy (ICOG-8), (abstract).

Dunphy, J.M., Ludden, J.N. and Parrish, R.R. 1995. Stitching together the Ungava orogen, northern Quebec: geochemical, isotopic and geochronological constraints from late magmatic events. Can. J. Earth Sci., submitted.

Emmermann, R., 1985. Basement geochemistry, hole 504B. In: R.N. Anderson, K. Honnorez and K. Becker (Editors), Initial Reports of the Deep Sea Drilling

Project. U.S. Government Printing Office, Washington, pp. 183-199.

Fitton, J.G., 1987. The Cameroon line, West Africa: a comparison between oceanic and continental alkaline volcanism. In: J.G. Fitton and B.G.J. Upton (Editors), Alkaline Igneous Rocks. Geol. Soc., Spec. Pub., No. 30, pp. 273-291.

Francis, D.M. and Hynes, A.J., 1979. Komatiite-derived tholeiites in the Proterozoic of New Quebec. Earth Planet. Sci. Lett., 44:473-481.

Francis, D. and Ludden, J., 1990. The mantle source for olivine nephelinite, basanite and alkaline olivine basalt at Fort Selkirk, Yukon, Canada. J. Petrol., 13:371-400.

Francis, D.M., Hynes, A.J. and Ludden, J.N., 1981. Crystal fractionation and partial melting in the petrogenesis of a Proterozoic high-MgO volcanic suite, Ungava, Quebec. Contrib. Mineral. Petrol., 78:27-36.

Francis, D.M., Ludden, J.N. and Hynes, A.J., 1983. Magma evolution in a Proterozoic rifting environment. J. Petrol., 24:556-582.

Gallagher, K. and Hawkesworth, C., 1992. Dehydration melting and the generation of continental flood basalts. Nature, 358:57-59.

Gaonac'h, H., Picard, C., Ludden, J.N. and Francis, D.M., 1989. Alkaline rocks from a Proterozoic volcanic island in the Cape Smith Thrust Belt, New Quebec. Geosci. Can., 16:137-139.

Gaonac'h, H., Ludden, J.N., Picard, C. and Francis, D.M., 1992. Highly alkaline lavas in a Proterozoic rift zone: Implications for Precambrian mantle metasomatic processes. Geology, 20:247-250.

Gariepy, C., Ludden, J. and Brooks, C., 1983. Isotopic and trace element constraints on the genesis of the Faeroe lava pile. Earth Planet. Sci. Lett., 63:257-272.

Green, D.H., Hibberson, W.O. and Jaques, A.L., 1979. Petrogenesis of Mid-ocean ridge basalts. In: M.W. McElhinny (Editor), The earth: its origin, structure and evolution. Academic Press, London, pp. 265-299.

Griffiths, R.W. and Campbell, I.H., 1990. Stirring and structure in mantle starting plumes. Earth Planet. Sci. Lett., 99:66-78.

Hegner, E. and Bevier, M.L., 1991. Nd and Pb isotopic constraints on the origin of the Purtuniq ophiolite and Early Proterozoic Cape Smith Belt, northern Quebec, Canada. Chem. Geol., 91:357-371.

Hoffman, P.F., 1990. Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen. In: J.F. Lewry and M.R. Stauffer (Editors), The Early Proterozoic Trans-Hudson Orogen of North America. Geol. Assoc. Can., Spec. Pap., No. 37, pp. 15-39.

Hofmann, A.W., Jochum, K.P., Seufert, M. and White, W.M., 1986. Nb and Pb in oceanic basalts: new constraints on mantle evolution. Earth Planet. Sci. Lett., 79:33-45.

Hynes, A.J. and Francis, D.M., 1982. A transect of the Early Proterozoic Cape Smith foldbelt, New Quebec. Tectonophysics, 88:23-59.

Jackson, G.D., 1960. Belcher Islands, Northwest Territories. In: Geol. Surv. Can. Pap., 60-20, pp.13.

Jahn, B.M., 1986. Mid-ocean ridge or marignal basin origin of the East Taiwan ophiolite: chemical and isotopic evidence. Contrib. Mineral. Petrol., 92:194-206.

Jahn, B.M., Auvray, B., Cornichet, J., Bai, Y.L., Shen, Q.H. and Liu, D.Y., 1987. 3.5 Ga old amphibolites from eastern Hebei Province, China: field occurrence, petrography, Sm-Nd isochron age and REE geochemistry. Precamb. Res., 34:311-346.

Jahn, B.M., Auvray, B., Shen, Q.H., Liu, D.Y., Zhang, Z.Q., Dong, Y.J., Ye, X.J., Zhang, Q.Z., Cornichet, J. and Mace, J., 1988. Archean crustal evolution in China: the Taish complex and evidence for juvenile crustal addition from long-term depleted mantle. Precamb. Res., 38:381-403.

Jakobsson, S.P., Jonsson, J. and Shido, F., 1978. Petrology of the western Reykjanes Peninsula, Iceland. J. Petrol., 19:669-705.

Klein, E.M. and Langmuir, C.H., 1987. Ocean ridge basalt chemistry, axial depth, and crustal thickness and temperature variation in the mantle. J. Geophys. Res., 92:8089-8115.

Larsen, L.M. and Watt, W.S., 1985. Episodic volcanism during break-up of the North Atlantic: evidence from the East Greenland plateau basalts. Earth Planet. Sci. Lett., 73:105-116.

Larsen, L.M., Watt, W.S. and Watt, M., 1989. Geology and petrology of the Lower Tertiary plateau basalts of the Scoresby Sound region, east Greenland. Bull. Geol. Soc. Greenland, 157:1-164.

Legault, F., Francis, D., Hynes, A.J. and Budkewitsch, P., 1994. Proterozoic continental volcanism in the Belcher Islands: implications for the evolution of the Circum Ungava Fold Belt. Can. J. Earth Sci. 31:153-1549.

Le Roex, A.P., Erlank, A.J. and Needham, H.D., 1981. Geochemical and mineralogical evidence for the occurrence of at least three distinct magma types in the 'Famous' region. Contrib. Mineral. Petrol., 77:24-37.

Lucas, S.B. and St-Onge, M.R., 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 2: Structural and metamorphic history. Can. J. Earth Sci., 29:765-782.

Lucas, S.B., St-Onge, M.R., Parrish, R.R. and Dunphy, J.M., 1992. Long-lived continent-ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. Geology, 20:113-116.

Machado, N., David, J., Scott, D.J., Lamothe, D., Philippe, S. and Gariépy, C., 1993. U-Pb geochronology of the western Cape Smith Belt, Canada: New insights on the age of initial rifting and arc magmatism. Precamb. Res., 63:211-223.

Mahlburg Kay, S. and Rapela, C.W. (Editors), 1990. Plutonism from Antarctica to Alaska. Geol. Soc. Am. Spec. Pap. No. 241, pp. 263.

Menzies, M.A., 1990. Petrology and geochemistry of the continental mantle: an historical perspective. In: M.A. Menzies (Editor), Continental Mantle, Oxford University Press, New York, N.Y., pp. 31-54.

Morgan, W.J., 1971. Convection plumes in the lower mantle. Nature, 230:42-43.

Parrish, R.R., 1989. U-Pb geochronology of the Cape Smith Belt and Sugluk block, northern Quebec. Geosci. Can., 16:126-130.

Picard, C., 1989a. Lithochimie des roches volcaniques protérozoïques de la partie occidentale de la Fosse de l'Ungava (région au sud du lac Lanyan). Ministère de l'Energie et des Ressources du Québec, ET 87-14.

Picard, C., 1989b. Pétrologie et volcanologie des roches volcaniques protérozoïques de la partie centrale de la Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, ET 87-07.

Picard, C., Lamothe, D., Piboule, M. and Oliver, R., 1990. Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt (New-Quebec). Precamb. Res., 47:223-249.

Rudnick, R.L. and Presper, T., 1990. Geochemistry of intermediate- to high-pressure granulites. In: D. Vielzeuf and T. Presper (Editors), Granulites and crustal evolution. Kluwer Academic Publishers, Boston, pp. 523-551.

Saunders, A.D., Tarney, J., Stern, C. and Dalziel, I.W.D., 1979. Geochemistry of Mesozoic marginal basin floor igneous rocks from southern Chile. Bull. Geol. Soc. Am., 90:237-258.

Scott, D.J., St-Onge, M.R., Lucas, S.B. and Helmstaedt, H., 1989. The 1998 Ma Purtuniq ophiolite: imbricated and metamorphosed oceanic crust in the Cape Smith Thrust Belt, northern Quebec. Geosci. Can., 16:144-147.

Scott, D.J., St-Onge, M.R., Lucas, S.B. and Helmstaedt, H., 1991. Geology and chemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. In: T. Peters, A. Nicolas and R.G. Coleman (Editors), Ophiolite Genesis and Evolution of the Oceanic Lithosphere. Kluwer Academic Publishers, Amsterdam, pp. 817-849.

Scott, D.J., Helmstaedt, H. and Bickle, M.J., 1992. Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. Geology, 20:173-176.

Silver, L.T. and Chappell, B.W., 1988. The Peninsular Ranges Batholith: an insight into the evolution of the Cordilleran batholiths of southwestern North America. Trans. R. Soc. Edinburg: Earth Sci., 79:105-121.

Sleep, N.H. and Windley, B.F., 1982. Archean plate tectonics: constraints and inferences. J. Geol., 90:363-379.

Smith, A.D. and Ludden, J.N., 1989. Nd isotopic evolution of the Precambrian mantle. Earth Planet. Sci. Lett., 93:14-22.

Stolz, A.J., Varne, R., Davies, G.R., Wheller, G.E. and Foden, J.D., 1990. Magma source components in an arc-continent collision zone: the Flores-Lembata sector, Sunda arc, Indonesia. Contrib. Mineral. Petrol., 105:585-601.

St-Onge, M.R. and Lucas, S.B., 1990a. Early Proterozoic collisional tectonics in the internal zone of the Ungava (Trans-Hudson) orogen, Lacs Nuvilik and Sugluk map areas, Quebec. In: Current Research, Part C. Geol. Surv. Can. Pap., 90-1C, pp. 119-132.

St-Onge, M.R. and Lucas, S.B., 1990b. Evolution of the Cape Smith Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. In: J.F. Lewry and M.R. Stauffer (Editors), The Early Proterozoic Trans-Hudson Orogen of North America. Geol. Assoc. Can., Spec. Pap., No. 37, pp. 313-351.

St-Onge, M.R. and Lucas, S.B., 1992. New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. In: Current Research, Part C. Geol. Surv. Can. Pap., 92-1C, pp. 31-41.

St-Onge, M.R. and Lucas, S.B., 1993. Geology of the eastern Cape Smith Belt: Parts of the Kangiqssujuaq, Cratère du Nouveau-Québec, and Lacs Nuvilik map areas, Quebec. In: Geol. Surv. Can. Mem., 438, pp. 110.

St-Onge, M.R., Lucas, S.B., Scott, D.J., Bégin, N.J., Helmstaedt, H. and Carmicheal, D.M., 1988. Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec. In: Current Research, Part C. Geol. Surv. Can. Pap., 88-1C, pp. 1-18.

St-Onge, M.R., Lucas, S.B. and Parrish, R.R., 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Can. J. Earth Sci., 29:746-764.

Sun, S.S. and McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. In: A.D. Saunders and M.J. Norry (Editors), Magmatism in the ocean basins. Geol. Soc., Spec. Pub., No. 42, pp. 313-345.

Taylor, F.C., 1982. Reconnaissance geology of a part of the Canadian Shield, northern

Quebec and Northwest Territories. In: Geol. Surv. Can. Mem., 399, pp. 32.

Taylor, S.R. and McLennan, S.M., 1985. The continental crust: its composition and evolution. Blackwell Scientific Publications, Boston, pp. 312.

Tremblay, G., 1991. Géologie de la région du lac Lessard (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 88-09.

Tual, E., Jahn, B.M., Bougault, H. and Joron, J.L., 1985. Geochemistry of basalts from hole 504B, Leg 83, Costa Rica rift. In: R.N. Anderson, J. Honnorez and K. Becker (Editors), Initial Reports of the Deep Sea Drilling Project. U.S. Government Printing Office, Washington, pp. 201-214.

Varne, R. and Fodden, J.D., 1986. Geochemical and isotopic systematics of eastern Sunda arc volcanics: implications for mantle sources and mantle mixing processes. In: F.C. Wezel (Editor), The origin of arcs. Elsevier, New York, pp. 159-189.

Wheller, G.E., Varne, R., Foden, J.D. and Abbott, M.J., 1987. Geochemistry of Quaternary volcanism in the Sunda-Banda arc, Indonesia and three-component genesis of island-arc basaltic magmas. J. Volcanol. Geotherm. Res. 32:137-160.

White, W.M. and Hofmann, A.W., 1982. Sr and Nd isotope geochemistry of oceanic basalt and mantle evolution. Nature, 296:821-825.

White, R.S. and McKenzie, D.P., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. J. Geophys. Res., 94:7685-7729.

Zindler, A., 1982. Nd and Sr isotopic studies of komatiites and related rocks. In: N.T. Arndt and E.G. Nisbet (Editors), Komatiites. George Allen and Unwin, London, pp. 399-420.

Zindler, A. and Hart, S.R., 1986. Chemical geodynamics. Ann. Rev. Earth Planet. Sci., 14:493-571.

Zindler, A., Hart, S.R., Frey, F.A. and Jakobsson, S.P., 1979. Nd and Sr isotope ratios and rare earth element abundances in Reykjanes Peninsula basalts: evidence for mantle heterogeneity beneath Iceland. Earth Planet. Sci. Lett., 45:249-262.

CHAPTER 3

Stitching together the Ungava Orogen, northern Quebec: Geochronological (TIMS and ICP-MS) and geochemical constraints on late magmatic events

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Canadian Journal of Earth Sciences, in press (v. 32), June 1995
submitted 22 September 1994, accepted 13 June 1995

ABSTRACT

Late magmatic activity in the Ungava Orogen of northern Quebec is manifest as granitic dykes and small, rare plutons which cross-cut all tectonostratigraphic elements of the orogen. Conventional U-Pb geochronology (TIMS) on one particularly important pluton which cuts all these domains (the Lac Duquet monzogranite) indicates its age of emplacement at 1742.2 +2.1/-1.5 Ma. This undeformed and non-metamorphosed pluton post-dates the youngest structures in the orogen (D_4 folds), thereby constraining the timing of the latest deformation to >1742 Ma. Laser ablation inductively coupled plasma mass spectrometry (ICP-MS) on zircons from the same sample identified a large range in $^{207}Pb/^{206}Pb$ ages of inherited grains from 1.7 to 3.2 Ga, corresponding to the ages of the host rocks for the pluton. This high-K peraluminous monzogranite pluton contains moderate to high concentrations of the large ion lithophile elements (LILE) and fractionated and enriched light rare earth elements (REE), similar in composition to the surrounding continental crust and to other crustally derived granites. Initial $^{87}Sr/^{86}Sr$ values of 0.7040-0.7051 and ε_{Nd} ranging from -4.4 to -9.7 indicate incorporation of a significant amount of older material in the petrogenesis of the pluton. It is proposed that anatexis of the surrounding continental crust due to structural thickening during the waning stages of the Ungava Orogeny resulted in the generation of the Lac Duquet pluton and was the source for its inherited zircons.

RÉSUMÉ

Le magmatisme tardif dans l'Orogène de l'Ungava, situé dans le nord du Québec, se manifeste par des dykes de pegmatite et des rares petits plutons granitiques qui recoupent tous les assemblages tectonostratigraphiques de l'orogène. La datation conventionnelle de U-Pb (TIMS) sur un pluton particulièrement important qui coupe ces assemblages (la monzogranite du Lac Duquet) indique un âge de mise en place à 1742.2 +2.1/-1.5 Ma. Le pluton, qui n'est pas déformé ou métamorphosé, recoupe les structures les plus jeunes dans l'orogène (des plis D_4) et par conséquence restreint cette déformation à >1742 Ma. L'ablation au laser et des analyses subséquentes par spectrométrie de masse du type ionisation dans un plasma (à induction couplée - ICP-

MS) sur des zircons a identifié une gamme dans les âges $^{207}Pb/^{206}Pb$ de 1.7 à 3.2 Ga, qui correspond aux âges des roches encaissantes du pluton. Ce monzogranite peralumineux est riche en K, et contient des concentrations modérées à élevées en elements lithophiles à rayons inoiques larges et des terres rares légères enrichies et fractionnées. Sa composition est comparable à celle de la croûte continentale et aux granites derivés de la croûte continentale. Les valeurs initiales de $^{87}Sr/^{86}Sr$ varient entre 0.7040 à 0.7051 et celle de ε_{Nd} entre -4.4 à -9.7 indiquant qu'une croûte ancienne était un composant important dans la pétrogenèse du pluton. Nous suggérons que la fusion partielle de la croûte continentale encaissante du pluton pendant et après l'événement orogénique de l'Ungava a généré le pluton du Lac Duquet, et cette croûte était la source de ces zircons hérités.

3.1 INTRODUCTION

Late-stage magmatic activity can be defined as magmatism occurring at the termination of, or following, an orogenic cycle or a major magmatic cycle within a particular region. Late magmatic activity has been documented in tectonic zones of all ages and locations on earth (e.g., the Archean Abitibi greenstone belt (Rive et al. 1990); the Proterozoic Trans-Hudson Orogen (Machado 1990); the Mesozoic to Cenozoic Andes mountain belt (Pitcher et al. 1985); the Tertiary Himalayan Range (Le Fort 1981; Deniel et al. 1987); and others (see Rogers and Greenberg 1990 and references therein)). Precise dating of late-stage orogenic plutons can contribute to an understanding of the timing of regional tectonic events such as terrane accretion or post-collisional deformation. In addition, geochronological and geochemical studies (including isotopic studies) on such plutons can also provide important information about the nature of the underlying crust such as its age and bulk composition (e.g., igneous or sedimentary, mafic or felsic, isotopically primitive or enriched). Inherited components such as xenoliths and/or xenocrysts are frequently present in these plutons due to incomplete melting during petrogenesis or assimilation of wall rocks during ascent and emplacement. Older zircons with or without younger overgrowths are a common inherited component which, with accurate dating, can help identify the sources for the magmas (e.g., Harrison et al. 1987; Williams and Claesson 1987; Paterson et al. 1992; Roddick and Bevier 1995). Such inherited components can complicate the interpretation of the isotopic and geochemical signature of the resulting plutons, particularly where involvement of several sources is indicated. In addition, certain accessory phases possibly present as xenocrysts (such as monazite, zircon, sphene and apatite) may contain up to 60% of the REE (rare earth element) mass budget (Reid 1990), and can profoundly affect the trace element and isotopic signature of the pluton.

Intrusion of granitic dykes and plutons during the termination of the Ungava Orogeny (St-Onge and Lucas 1990a) provides an excellent opportunity to study the potential effect of inherited components in their petrogenesis, as well as to constrain the timing of regional tectonic events. One particular pluton, the Lac Duquet monzogranite, intruded across the contact of two major tectonostratigraphic domains, effectively "stitching" them together. Conventional U-Pb geochronology (using thermal ionization mass spectrometry - TIMS) on a sample from this pluton constrained its age of emplacement and also indicated the presence of inherited components (Dunphy et al. 1991). In order to further document the age(s) of the source(s) and to quantify the proportions of the various components involved in the petrogenesis of this granite, a more detailed study of the inherited zircon grains was undertaken using inductively coupled plasma mass spectrometry (ICP-MS) with a laser ablation sampling technique. This technique has only recently been demonstrated as a viable and feasible complement to the conventional TIMS and SHRIMP (Sensitive High Resolution Ion MicroProbe, Compston et al. 1984) methods for measuring ^{207}Pb/^{206}Pb ages (Feng et al. 1993; Fryer et al. 1993). Here we present both TIMS and ICP-MS data for the Lac Duquet monzogranite and evaluate the effectiveness of the laser ablation ICP-MS technique as a tool for identifying and dating inherited zircon in this pluton. We then use the proportions of the zircon populations defined by the ICP-MS data, in conjunction with Nd and Sr isotopic data for the proposed sources, to model the isotopic composition of the granite and verify the results against actual isotopic and geochemical measurements from samples of the pluton.

3.2 GEOLOGICAL SETTING

The Ungava Orogen is an Early Proterozoic arc-continent collisional belt (Lucas et al. 1992; St-Onge et al. 1992) which represents one segment of the much larger orogenic system, the Trans-Hudson Orogen, one of the largest and best exposed Early Proterozoic orogenic belts in the world (Lewry and Collerson 1990; Hoffman 1990). The Ungava Orogen consists of three distinct tectonostratigraphic units (Fig. 1; St-Onge et al. 1992): (1) (par)-autochthonous Superior Province basement plutonic and supracrustal rocks of Archean age (St-Onge and Lucas 1990a, 1992; Lucas and St-Onge 1991); (2) autochthonous and allochthonous sedimentary and volcanic rocks associated with a ca. 2.04 - 1.92 Ga rift to drift margin (Povungnituk and Chukotat Groups; Hynes and Francis 1982; Francis et al. 1983; Picard et al. 1990; St-Onge and Lucas 1990b); and (3) allochthonous "suspect" crustal components of a ca. 2.00 Ga ophiolite (Watts Group; St-Onge et al. 1988; Scott et al. 1989, 1991; St-Onge and Lucas 1990b) and a ca. 1.90 - 1.80 Ga magmatic arc (Narsajuaq terrane; St-Onge and Lucas 1990a, 1992; Dunphy and Ludden 1992; St-Onge et al. 1992) and associated rocks (Parent/Spartan Groups; Lamothe et al. 1983; St-Onge et al. 1988; Picard et al. 1990; St-Onge and Lucas 1990b, 1992; Barrette 1994).

Each of the three tectonostratigraphic domains underwent different episodes of deformation at different times. Four deformation events have been documented (Lucas and St-Onge 1992): (1) a pre-accretion deformation (D_1), which is Archean for the Superior Province basement and Proterozoic for the other two domains; (2) a Proterozoic accretion-related deformation (D_2), which is recorded in all elements of the orogen; (3) regional scale folding of all tectonic elements about east-west axes (D_3); and (4) subsequent cross-folding of all units about north-south axes (D_4). Accretion-related deformation (D_2) occurred after 1.83-1.84 Ga (Lucas and St-Onge 1992) yet predates 1.76 Ga, the age of the orogen-wide cross-cutting pegmatite dykes (Parrish 1989) which were emplaced during the waning stages of D_3 (Lucas and St-Onge 1992).

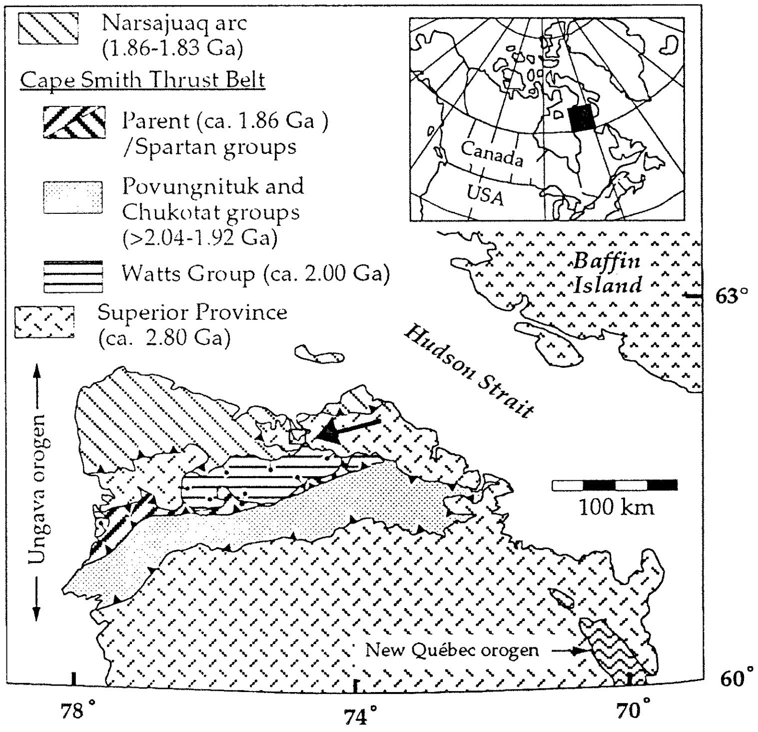

Fig. 3.1. Location map of the Ungava Orogen illustrating its principal tectonostratigraphic elements (after St-Onge and Lucas 1992). The location of figure 2 is outlined by the small box (indicated by the large black arrow).

Late magmatic activity within the Ungava Orogen is represented by the presence of pervasive granitic (*senso lato*) pegmatite dykes and sills that cross-cut all elements of the orogen, and small, rare granitic plutons that are concentrated in the Narsajuaq terrane. One intrusion that has particularly important structural significance is the Lac Duquet pluton (Fig. 2). This pluton cross-cuts the par-autochthonous Archean Superior Province basement (manifest mainly as veins and dykes that emanate from the pluton), a Cape Smith Belt thrust slice and Proterozoic Narsajuaq terrane arc rocks, which are all folded into a north-trending D_4 synform (St-Onge and Lucas 1990a; Lucas and St-Onge 1992). The structural position of the Lac Duquet pluton cutting this synform requires that the youngest deformation episode (D_4) predate its emplacement which is constrained at 1.74 Ga (see below). The Lac Duquet pluton is a medium-grained, equigranular biotite +/- muscovite monzogranite (based on Streckeisen's (1976) classification using modal mineral abundances) with accessory apatite, titanite, zircon and minor opaque phases. On its outcrop-scale of approximately 12 km^2, the Lac Duquet monzogranite is compositionally very homogeneous, exhibiting little variation in its mineralogy or textural relationships. The pluton is undeformed, with the exception of a weak foliation developed locally along its margins (possibly emplacement related). The pluton cuts the deformational fabrics in the host rocks. Near the structural top of the pluton abundant enclaves are found which have a composition and deformational character that is similar to the surrounding rocks of the Narsajuaq terrane. It is postulated that the enclaves represent country rock roof pendants within the pluton. Other smaller granitic plutons have been delineated within the Ungava orogen (not shown on Fig. 1, owing to scale) which possess characteristics (mineralogy, bulk composition, deformation state) similar to the Lac Duquet pluton.

3.3 GEOCHRONOLOGY

Conventional TIMS zircon geochronology on sample L-68-89 was used to determine the precise crystallization age of the Lac Duquet monzogranite and to determine if any inherited components were present. A complimentary study on the same sample using laser ablation ICP-MS was then undertaken to further document the

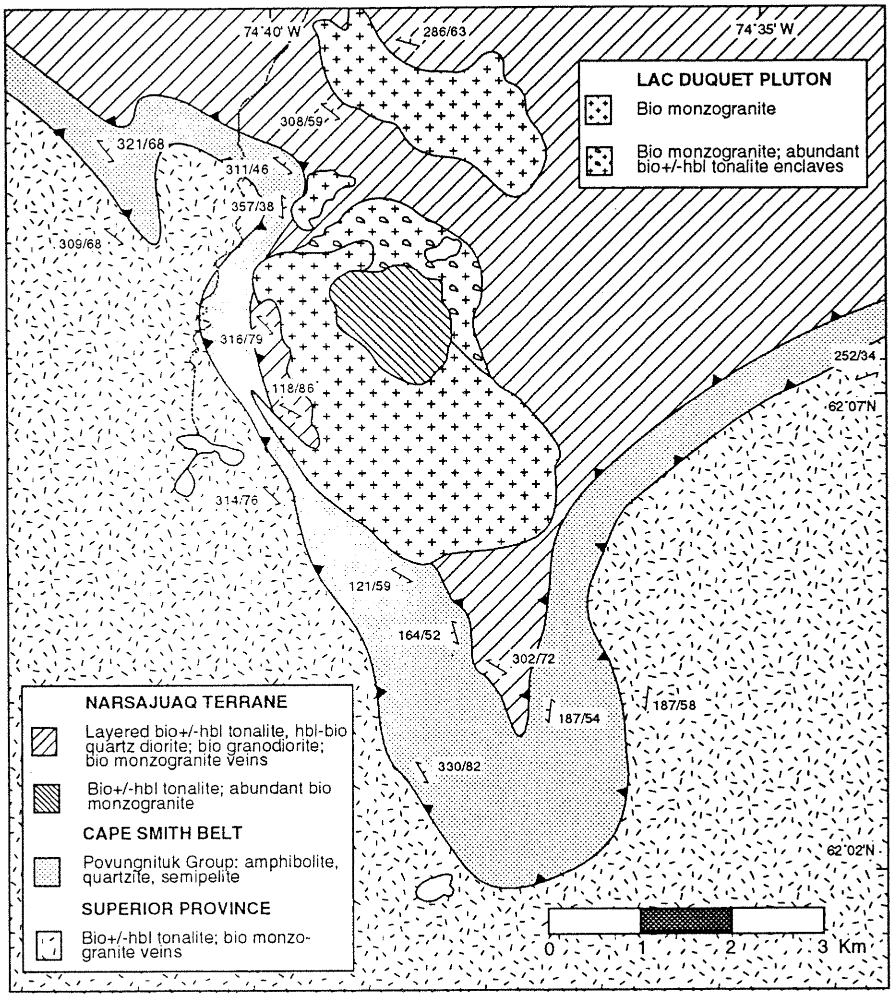

Fig. 3.2. Detailed (1:50 000 scale) map of the Lac Duquet pluton and surrounding tectonostratigraphic units. The pluton cross-cuts the Archean Superior Province basement (primarily veins and dykes that radiate from the pluton), a Cape Smith thrust slice (Povungnituk Group) and the plutonic units of the Narsajuaq terrane which have all been folded into a D_4 synform.

inheritance and contamination history of the pluton. Comparison of these results with the TIMS results documents the potential of this technique as a valuable tool to the geochronologist.

3.3.1 Analytical Techniques

Mineral separates for geochronology were extracted from approximately 30 kg of sample by standard techniques of crushing and grinding, followed by Wifley table concentration, heavy liquid and magnetic separation. Hand picking of grains, using criteria such as morphology, clarity and absence of cracks was used to select the highest quality grains for isotopic analysis. Analytical methods for the conventional analysis are summarized by Parrish et al. (1987) and include air abrasion (Krogh 1982), chemical dissolution and separation of U and Pb, followed by isotopic analysis on a Finnigan Mat 261 variable multicollector thermal ionization mass spectrometer at the Geological Survey of Canada, Ottawa. TIMS analytical results are presented in Table I and Figure 3.

The analytical method and operating parameters used for the laser ablation technique are presented in detail in Feng et al. (1993) and are summarized here. For laser ablation ICP-MS analysis, zircons were selected from the same concentrate using the same criteria as for the TIMS analysis. The zircons range in size from ≈ 70-100 µm, although some grains are up to 200 µm large. Zircons were mounted in epoxy and slightly polished to expose the grains which resulted in a reduction of their size (thickness). The zircon samples were ablated with a Nd:YAG laser which produced cylindrical pits of ≈ 40-60 µm in diameter for these samples. For each sample pit, 10-20 repeats were performed, depending on the size of the zircon sample. Given the drilling rate of the laser estimated by Feng et al. (1993) at 0.1-0.3 µm/s and a measurement of 100-200 s (depending on the number of repeats), a minimum grain thickness of 60 µm is necessary to ensure a complete analysis. For some grains on the polished mount this was not achieved since they were completely perforated before the end of the run. Data from such samples were discarded. In addition, any samples having < 3 ppm ^{207}Pb were rejected as such samples may yield inaccurate ages due to the decreased analytical precision (Feng et al., 1993). The vaporized material was transported to the ICP-MS by an argon carrier gas for analysis. A Fisons VG PQII+ICP mass spectrometer and a Fisons VG Laser-Probe at the Université de Montréal were used to measure the isotopic ratios of the samples. ICP-MS results are presented in Table II and Figure 4.

Correction for mass discrimination and instrument drift was made by using an external standard, a large zircon (CN-92-1) from a skarn in the Grenville Province which has an average $^{207}Pb/^{206}Pb$ ratio of 0.07784 ± 0.00006 (2σ) (TIMS analysis, Feng et al. 1993). The average of the means of $^{207}Pb/^{206}Pb$ for 12 sample pits in the standard as measured by ICP-MS in this study is 0.07845 ± 0.00052 (1σ) which is within 0.8% of the TIMS data. A large range of precisions (RSD) on the $^{207}Pb/^{206}Pb$ ratio for individual sample pits was measured in our samples (0.3-27%, average ≈ 6%) which is greater than that reported by Feng et al. (1993), and is most likely a result of Pb heterogeneity throughout the zircon grain. The RSD correlates with the $^{207}Pb/^{206}Pb$ age such that the average precision in terms of error on the age is ≈ 60 Ma at 1.7-1.8 Ga, ≈ 90 Ma at 2.0 Ga and ≈ 150 Ma for samples of Archean age (Fig. 4, inset). The $^{207}Pb/^{206}Pb$ ages and their uncertainties were calculated using the ISOPLOT program of Ludwig (1990).

Assessing the degree of concordance or discordance of the Pb ratios in laser ablation ICP-MS analysis is currently a challenge. Using the ICP-MS laser ablation technique Feng et al. (1993) determined the $^{207}Pb/^{206}Pb$ ages of samples that had known concordant TIMS ages. They demonstrated that for zircons having a grain size > 60 µm and concentration of $^{207}Pb > 3$ ppm, the ICP-MS age was within 1% of the TIMS age, whereas smaller zircons and those having lower ^{207}Pb concentrations produced inaccurate ages. A similar study by Fryer et al. (1993) also produced $^{207}Pb/^{206}Pb$ ICP-MS ages that agreed with the known TIMS age for the one sample analyzed. Feng et al. (1993) also analyzed zircons having a more complex behaviour (discordant, overgrowths). Although these grains produced $^{207}Pb/^{206}Pb$ ages that were compatible with the TIMS data, the interpretation of such data was more complex (Feng et al. 1993). Some samples with complex histories produced a wide range of ages, while another sample documented a systematic variation in $^{207}Pb/^{206}Pb$ ratios with depth through the grain (Fig. 2 from Feng et al. 1993), indicating the presence of a core-rim relationship. Evaluation of the constancy of the $^{207}Pb/^{206}Pb$ ratios during the ICP-MS analysis is one means of assessing the concordance of the analyzed grains, as concordant grains should not show large variations in this ratio, whereas grains with cores or those that are metamict would likely have significant variations in the ratios. The analytical procedure followed for the ICP-MS analysis in our present study did not permit us to evaluate the change in the $^{207}Pb/^{206}Pb$ ratios during the run and the reported values (Table II) are the mean ratio of the grain. Large variations in the individual ratios during a run, however, would be reflected in a large error on the mean ratio, which may reflect discordant behaviour. As we cannot establish with certainty whether or not the grains are concordant, the data must be considered as minimum ages, although zircons that have small errors are the most likely candidates for concordant grains.

3.3.2. Conventional TIMS Results

For the conventional TIMS analysis two size fractions (105-149 and >149 μm) from the least magnetic zircon separate from sample L-68-89 were hand picked into four different fractions and were used for isotopic analysis (Table I). Zircons from fraction B were round or oval in shape and were slightly cloudy due to small fractures and cracks present in the grains. Using the morphology of the zircon grains as a criteria, two euhedral and potentially magmatic zircon grains (B-2 and B-3) were analyzed individually. These grains have identical $^{207}Pb/^{206}Pb$ ages, with B-3 essentially concordant, and define a discordia line whose upper intercept is 1742.2 +2.1/-1.5 Ma (Fig. 3, Table I). This age is interpreted as the age of crystallization of this granite since the pluton post-dates the documented deformation and metamorphic events in the orogen. The ages do not correspond to any known potential source within the region, and as such are inconsistent with a xenocrystic origin. A third grain (B-4) produced a much older age of 2606 Ma with a large error (+33/-34 Ma). Fractions A and D (four grains each) consisted of elongate, generally clear grains, yellow-brown (A) to pink-grey (D) in

Table I. U-Pb TIMS analytical results for sample L-68-89 from the Lac Duquet pluton.

sample	no. of grains	wt (μg)	U (ppm)	Pb [a] (ppm)	$\frac{^{206}Pb}{^{204}Pb}$ [b]	Pb_c [c] (pg)	$\frac{^{208}Pb}{^{206}Pb}$ [a]	$\frac{^{206}Pb}{^{238}U}$ [d]	$\frac{^{206}Pb}{^{238}U}$ [d] (Ma)	$\frac{^{207}Pb}{^{235}U}$ [d]	$\frac{^{207}Pb}{^{235}U}$ [d] (Ma)	$\frac{^{207}Pb}{^{206}Pb}$ [d]	corr. coef.	$\frac{^{207}Pb}{^{206}Pb}$ [e] (Ma)
Lac Duquet monzogranite (L-68-89)														
A-1	4	49	825.2	274.5	4582	175	0.106	0.3138±0.08	1759.6±2.6	4.871±0.10	1797.3±1.7	0.1126±0.04	0.94	1841.3 ±1.3
B-2	1	37	388.7	119.7	3271	86	0.066	0.3055±0.09	1718.6±2.6	4.490±0.11	1729.2±1.7	0.1066±0.04	0.92	1742.0 +1.5/-1.6
B-3	1	15	165.2	51.86	3724	13	0.074	0.3092±0.09	1736.7±2.6	4.545±0.10	1739.2±1.7	0.1066±0.04	0.94	1742.2 ±1.3
B-4	1	10	90.25	51.48	133	233	0.193	0.4844±0.38	2546.4±15.9	11.69±1.24	2579.9±23.1	0.1750±1.01	0.70	2606.3 +33.2/-34.0
C-1	1	26	86.05	30.56	3447	14	0.145	0.3284±0.09	1830.4±2.8	5.076±0.10	1832.1±1.7	0.1121±0.04	0.93	1834.0 ±1.4
C-2	1	15	113.9	70.62	4188	13	0.214	0.5140±0.09	2673.8±3.8	13.22±0.10	2695.4±1.9	0.1865±0.03	0.95	2711.7 ±1.0
C-3	1	10	386.2	120.1	2653	29	0.074	0.3055±0.09	1718.6±2.6	4.585±0.10	1746.6±1.7	0.1088±0.04	0.91	1780.2 ±1.6
D-1	4	24	160.0	56.65	4202	20	0.136	0.3279±0.09	1827.9±2.8	5.387±0.10	1882.8±1.7	0.1192±0.04	0.94	1943.8 ±1.2

[a] Radiogenic Pb.

[b] Measured ratio, corrected for spike and Pb fractionation.

[c] Total common Pb, corrected for fractionation and spike.

[d] Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock); errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean in Ma.

[e] Corrected for blank and common Pb.

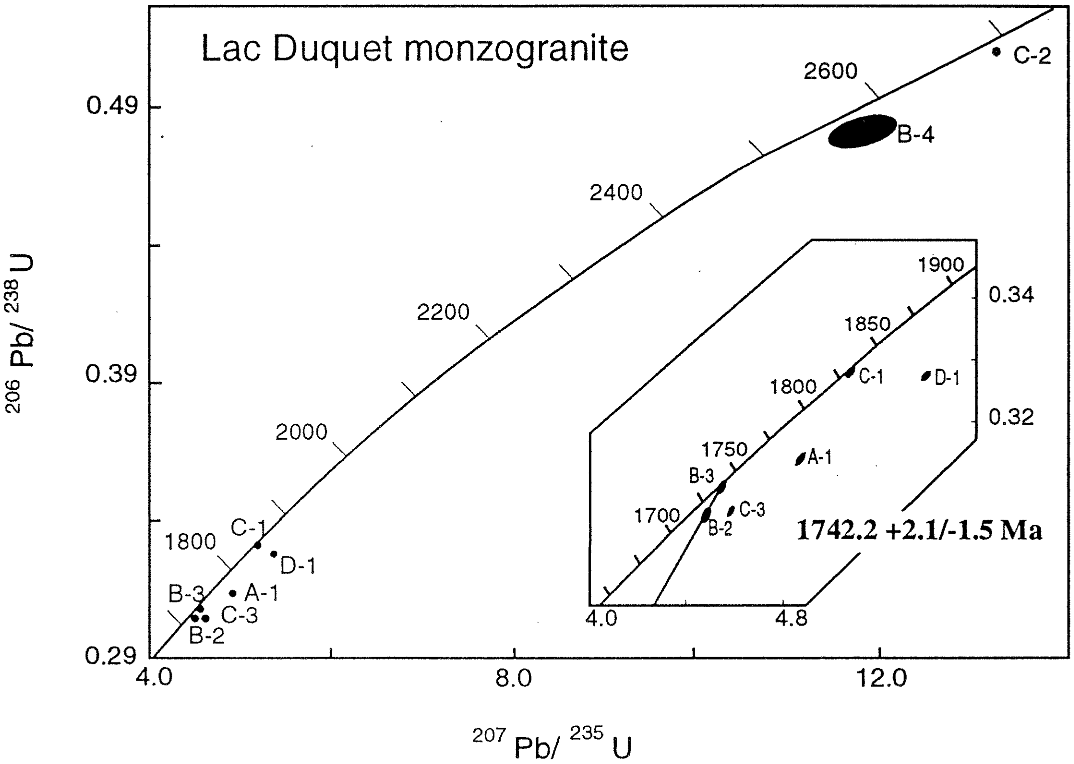

Fig. 3.3. U-Pb concordia diagram (conventional TIMS analyses) for sample L-68-89 from the Lac Duquet monzogranite. The inset shows the younger data with the interpreted igneous age of crystallization defined by grains B-2 and B-3 at 1742.2 Ma, whereas the various other zircon grains are interpreted as inherited. The size of the data ellipse reflects the 2 sigma error.

colour. Regular igneous zoning was visible in most grains, some of which also had small euhedral to subhedral cores. These fractions produced discordant analyses giving ^{207}Pb/^{206}Pb ages of 1841.3 ± 1.3 (A-1) and 1943.8 ± 1.2 (D-1) Ma. Two clear, flat, anhedral zircons, slightly pink in colour, were suspected as possible xenocrystic grains and have ^{207}Pb/^{206}Pb ages of 1834.0 ± 1.4 Ma (C-1) and 2711.7 ± 1.0 Ma (C-2). A third grain from the same fraction (C-3) has a younger discordant age of 1780.2 ± 1.6 Ma and is also interpreted as xenocrystic.

3.3.3 Laser Ablation ICP-MS Results

ICP-MS LaserProbe analysis of other zircons having various morphologies from this same sample (L-68-89) produced 110 useful ^{207}Pb/^{206}Pb ages on a total of 74 zircon grains (Table II). Most ages are single pit ages, although a number of grains (28) were large enough for multiple pit analysis. For these larger grains, only the average of the ^{207}Pb/^{206}Pb ratios and ages are reported. The ^{207}Pb/^{206}Pb ages span a range from 1.7 to 3.2 Ga with a clustering of ages from 1.7-2.1 Ga and a smaller maxima from 2.4-2.8 Ga (Fig. 4). Filtering of the data to include only samples with errors falling within 10% or 5% of the ^{207}Pb/^{206}Pb age does not change the overall distribution of the ages, although the oldest grains (>3.1 Ga) are eliminated (Fig. 4). Subsequent discussion, however, will focus on the full data set (n=74). Relatively few zircon grains have a ^{207}Pb/^{206}Pb age that corresponds with the TIMS U-Pb age of crystallization (approximately 20%, using the average error of ±60 Ma for ages of 1.7-1.8 Ga). However a significant number of grains (>80% of the total) have ages (± errors) that correspond to the age of the host rocks for the granite (1800-1898 Ma, Narsajuaq arc (St-Onge et al. 1992; Machado et al. 1993; R. Parrish, unpublished data); 1918-2038 Ma, Povungnituk and Chukotat Groups (Parrish 1989; Machado et al. 1993); 2740-2905 Ma, Superior Province (Parrish 1989, unpublished data)). In addition some very old zircons have been dated in the Superior Province basement of northern Quebec (3.4-3.6 Ga; R. Parrish, unpublished data; Stern et al. 1994). If such a crust was involved in the petrogenesis of this pluton it may have been the source for the very old zircon grains (3.1 Ga) in this sample.

Table II. Laser ablation ICP-MS data for sample L-68-89 from the Lac Duquet monzogranite.

grain # [a]	^{207}Pb (ppm)	^{207}Pb/^{206}Pb* ± 1σ	^{207}Pb/^{206}Pb age (Ma) ± 1 σ	grain # [a]	^{207}Pb (ppm)	^{207}Pb/^{206}Pb* ± 1σ	^{207}Pb/^{206}Pb age (Ma) ± 1 σ
M25-2d	5	0.1043 ± 16	1702 ± 28	J13-7a*	24	0.1196 ± 103	1950 ± 150
M25-2c	6	0.1062 ± 33	1735 ± 58	M25-3e*	9	0.1202 ± 92	1959 ± 140
M25-2e*	8	0.1077 ± 4	1761 ± 7	J14-5k	12	0.1204 ± 138	1962 ± 200
J14-9d	9	0.1077 ± 33	1761 ± 55	M25-1g*	12	0.1205 ± 102	1964 ± 150
J13-8a	22	0.1078 ± 33	1763 ± 56	J13-8d*	14	0.1210 ± 36	1971 ± 53
M25-1c*	17	0.1079 ± 27	1764 ± 46	J14-5a	11	0.1215 ± 81	1978 ± 120
M25-3j	9	0.1083 ± 74	1771 ± 120	M25-6k	18	0.1216 ± 99	1980 ± 150
J13-7e*	71	0.1086 ± 47	1776 ± 80	J14-9a*	25	0.1225 ± 9	1993 ± 13
M25-3b	6	0.1086 ± 34	1776 ± 56	M25-1m*	20	0.1230 ± 23	2000 ± 33
M25-1h*	15	0.1089 ± 13	1781 ± 22	M25-6i	4	0.1235 ± 27	2007 ± 39
J13-7f*	30	0.1090 ± 10	1783 ± 17	J14-5b	6	0.1237 ± 109	2010 ± 160
M25-1l	12	0.1092 ± 47	1786 ± 78	M25-3c*	11	0.1242 ± 80	2017 ± 110
M25-3f	4	0.1094 ± 38	1789 ± 63	J13-3d	15	0.1246 ± 53	2023 ± 75
M25-6b	21	0.1094 ± 60	1789 ± 100	J13-7g*	45	0.1259 ± 202	2041 ± 280
J13-7h*	17	0.1097 ± 43	1794 ± 71	J13-2a	4	0.1281 ± 47	2072 ± 65
M25-2a*	9	0.1102 ± 4	1803 ± 6	J13-8c*	17	0.1288 ± 100	2082 ± 140
M25-1a*	12	0.1108 ± 34	1813 ± 50	J14-4a	14	0.1291 ± 97	2086 ± 130
M25-3h	3	0.1112 ± 23	1819 ± 38	J14-6b	7	0.1341 ± 129	2152 ± 170
M25-2b	9	0.1112 ± 37	1819 ± 60	M25-6c*	25	0.1361 ± 122	2178 ± 160
J14-6f	19	0.1125 ± 32	1840 ± 52	J13-7c	19	0.1362 ± 118	2179 ± 150
J13-3f	21	0.1132 ± 58	1851 ± 93	M25-3k	7	0.1525 ± 87	2374 ± 97
J13-7b	30	0.1133 ± 29	1853 ± 47	J14-6d	21	0.1593 ± 230	2448 ± 240
M25-1i*	11	0.1141 ± 74	1866 ± 120	M25-1d	19	0.1605 ± 178	2461 ± 190
J14-9b*	17	0.1142 ± 25	1867 ± 39	M25-1o	25	0.1606 ± 133	2462 ± 140
J14-6e	7	0.1146 ± 49	1874 ± 77	J14-5d	23	0.1661 ± 225	2519 ± 230
M25-1b*	11	0.1148 ± 25	1877 ± 39	J13-3e	9	0.1694 ± 104	2552 ± 100
M25-1k	11	0.1153 ± 62	1885 ± 96	M25-3l	11	0.1734 ± 221	2591 ± 210
M25-3d	3	0.1154 ± 82	1886 ± 130	M25-1e*	20	0.1760 ± 17	2616 ± 16
M25-3n*	12	0.1158 ± 31	1892 ± 48	J14-9e*	30	0.1798 ± 17	2651 ± 16
J13-2b	3	0.1159 ± 15	1894 ± 23	J13-8e*	23	0.1802 ± 1	2655 ± 1
J13-7d	24	0.1166 ± 55	1905 ± 85	J13-8b	13	0.1831 ± 98	2681 ± 88
M25-1f	7	0.1170 ± 59	1911 ± 90	M25-6f	17	0.1840 ± 265	2689 ± 240
J14-9c*	11	0.1174 ± 16	1917 ± 25	J14-6a	34	0.1867 ± 358	2713 ± 320
M25-1j	16	0.1175 ± 19	1919 ± 30	J14-6g	30	0.1927 ± 66	2765 ± 56
J13-3b	11	0.1177 ± 66	1922 ± 100	M25-6d*	21	0.2092 ± 235	2899 ± 180
M25-2f*	17	0.1179 ± 30	1925 ± 46	M25-1n	26	0.2130 ± 99	2929 ± 75
M25-3g	11	0.1194 ± 24	1947 ± 35	M25-3a	13	0.2511 ± 686	3192 ± 430

[a] Grains marked with * after their name represent samples that were large enough for multiple hole analyses - only the average values for these grains are reported.

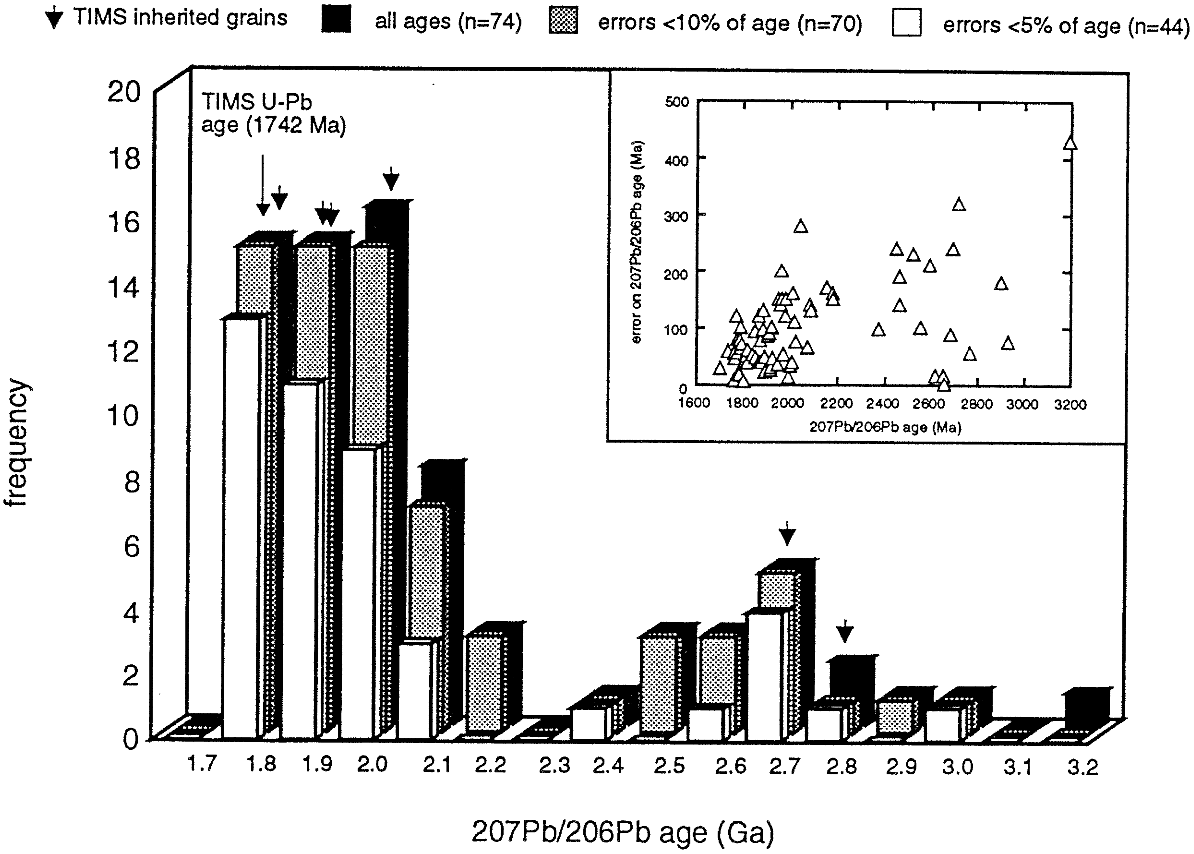

Figure 3.4. Histogram of the ^{207}Pb/^{206}Pb ages as determined by laser ablation ICP-MS analysis from the same sample (L-68-89) that was analyzed by TIMS. Three sets of data are shown (all ages, ages with errors <10% of the age and <5% of the age) to illustrate that the overall distribution of the ages does not change when the data is filtered to reduce the errors, although the oldest grains (>3.1 Ga) are eliminated from the filtered data sets. Arrows on top of the bars schematically refer to the ages of the zircon grains that were analyzed by TIMS. The large range in ages of zircon grains present in the granite reflects the multiple sources involved in its petrogenesis. The inset shows the range in errors on the ^{207}Pb/^{206}Pb ages illustrating the variation and slight increase in error with increasing age.

3.3.4 Comparison of TIMS and ICP-MS results

Conventional TIMS U-Pb geochronology constrains the age of crystallization of the Lac Duquet pluton at 1742 Ma. The youngest age obtained from the ICP-MS analysis (1702 ± 28 Ma) does not overlap with this age. However, by averaging this analysis with other analyses that fall within 3% of it (the approximate average precision for grains 1.7-1.8 Ga) gives an age of 1719 ± 43 Ma which is within error of the TIMS age. Although the ICP-MS age is not as precise as the TIMS age, it provides a reasonable estimate of the time of emplacement of the granite. Clearly the conventional TIMS analysis better constrains the age of this event. On the other hand, the laser ablation ICP-MS technique is a rapid and efficient tool in determining a minimum emplacement age for plutons which have a simple post-intrusion history, such as the Lac Duquet granite.

The greatest concentrations of ages as determined by the ICP-MS laser ablation method bracket the range of $^{207}Pb/^{206}Pb$ ages of the inherited grains as determined by the TIMS method (Fig. 4). Although the average errors on the $^{207}Pb/^{206}Pb$ ages are significantly larger for the ICP-MS technique relative to TIMS, the good correspondence between the two techniques for determining the distribution of inherited grains, coupled with the ease and speed of the laser ablation ICP-MS analysis (the quantity of data presented here for ICP-MS analysis was completed in only 12 hours), illustrates its potential as a geochronology tool, particularly for cases like the Lac Duquet pluton where a wide range in ages of inherited grains have been delimited.

3.4 GEOCHEMISTRY AND ISOTOPIC ANALYSES

In order to more fully characterize the signature of the late magmatic activity, representative samples were collected from the Lac Duquet pluton for geochemical and isotopic analysis. Given the small size of the Lac Duquet pluton and its compositional and textural homogeneity on the outcrop scale, it is likely that the samples were generated at the same time and are petrogenetically related.

3.4.1. Major and trace element geochemistry

The samples from the Lac Duquet pluton are mildly peraluminous (A/CNK = 1.07-1.14) and define a narrow range of SiO_2 concentrations from 72 to 76 wt% (Table III). They are rich in K_2O (3.7-4.9 wt%), have moderate concentrations of Al_2O_3 (14.1-15.9 wt%), Na_2O (3.4-4.5 wt%) and CaO (0.8-1.3 wt%) and low concentrations of Fe_2O_3 (0.6-1.1 wt%) and MgO (<0.24 wt%) (Table III). Moderate to high concentrations of the large ion lithophile elements (LILE's) are noted, with Ba contents of 725-1615 ppm, Rb = 72-111 ppm and Sr at 195-311 ppm (Fig. 5). Th and U concentrations are also enriched at 6.5-17.2 and 1.4-2.5 ppm, respectively. The overall concentration of the LILE's in the Lac Duquet pluton is similar to the average composition of the Superior Province continental crust exposed in the Ungava orogen (Harvey 1995) and crustally derived granites such as the Cretaceous plutons of the Old Woman Mountains area (OWP) of southeastern California (Miller and Wooden 1994; Fig. 5). The Lac Duquet monzogranite has fractionated light REE's with La_n/Sm_n ratios ranging from 4.4 to 8.8, but flat to only slightly fractionated heavy REE's without significant Eu anomalies (Fig. 5). The Lac Duquet pluton has lower REE concentrations (e.g., La = 10.7-23.9 ppm; Yb = 0.4-0.5 ppm) compared to the average composition of Superior Province crust and the OWP peraluminous granites (Fig. 5).

3.4.2 Isotope geochemistry

A series of samples from the Lac Duquet monzogranite were analyzed for their Sr and Nd isotopic compositions in an attempt to better characterize their sources and evolution. Although the field and geochemical data indicate little compositional variation in the pluton, several samples were analyzed, as it was deemed important to fully quantify the isotopic composition of the granite because it has frequently been noted that the isotopic composition of crustally-derived granites can be quite variable (e.g., Roddick and Compston 1977; Deniel et al. 1987). The calculated initial $^{87}Sr/^{86}Sr$ ratios (1742 Ma) for samples from the Lac Duquet pluton range from 0.7040 to 0.7051 (Table IV) and indicate that crustal contamination played a role in its petrogenesis. Nd isotopic analysis of samples from the Lac Duquet pluton support the contamination

Table III. Major and trace element analysis of whole rock samples from the Lac Duquet monzogranite.

sample:	D-19-89	L-67b-89	L-68-89	L-95-89	L-96-89	S-44-89	avg. Sup.	avg. OWP
SiO_2	74.26	74.01	74.60	72.47	75.03	75.99	62.92	72.41
TiO_2	0.10	0.14	0.09	0.07	0.08	0.07	0.48	0.12
Al_2O_3	14.77	14.37	14.06	15.86	14.16	14.21	15.45	14.14
Fe_2O_3	0.85	1.10	0.80	0.59	0.78	0.59	5.90	1.44
MnO	0.01	0.01	0.01	0.01	0.01	0.01	0.09	0.03
MgO	0.19	0.24	0.11	0.08	0.18	0.09	3.41	0.24
CaO	1.22	1.20	1.12	1.33	0.84	1.09	5.43	1.22
Na_2O	3.83	3.45	3.37	4.49	3.36	3.98	3.81	3.79
K_2O	4.52	4.61	4.82	4.66	4.94	3.67	2.28	4.39
P_2O_5	0.04	0.05	0.05	0.03	0.04	0.03	0.12	0.06
LOI	0.20	0.40	0.20	0.83	0.40	0.28	nd	nd
TOTAL	99.99	99.58	99.23	100.42	99.82	100.01	99.89	97.84
Rb	97.2	98.3	99.9	71.6	106.5	110.6	53.1	159
Sr	259	310	250	311	195	272	441	228
Ba	1145	1615	974	nd	1010	725	692	808
Y	3.2	2.7	2.9	1.5	3.5	3.3	12.7	14
Nb	6.5	8.3	7.3	4.0	7.4	6.7	6.2	15.9
Th	12.3	9.6	9.7	nd	17.2	6.5	6.5	11.6
U	1.5	2.1	1.6	nd	2.5	1.4	nd	1.1
Zr	106	133	103	128	84	53	149	81
Ni	5.4	6.9	4.7	nd	7.0	5.5	44.4	nd
Cr	6.5	4.9	3.3	nd	4.5	4.3	24.5	nd
La	23.90	11.20	21.20	nd	12.20	10.70	37.51	28.4
Ce	41.10	41.20	35.10	nd	29.70	18.30	68.21	52.8
Nd	13.90	10.20	12.90	nd	8.60	7.70	26.44	16.2
Sm	1.76	1.65	1.68	nd	1.57	0.88	4.89	4.69
Eu	0.55	0.57	0.55	nd	0.53	0.38	1.32	0.74
Tb	---	0.17	---	nd	0.15	0.15	0.40	0.72
Yb	0.50	0.40	0.50	nd	0.50	0.40	0.96	1.85
Lu	0.08	0.13	0.06	nd	0.06	0.05	0.17	0.22
A/CNK	1.10	1.12	1.10	1.07	1.14	1.14	0.83	1.07
La_n/Sm_n	8.8	4.4	8.1	nd	5.0	7.8	5.0	3.9
La_n/Yb_n	34.3	20.1	30.4	nd	17.5	19.2	28.03	11.0
locality (N)	6881250	6881750	6883000	6882250	6881500	6879500		
locality (E)	518500	517000	517400	519000	518750	527000		

Notes: Major elements in wt%, traces in ppm. LOI = loss on ignition; nd = not determined; --- below detection; A/CNK = molecular Al/(Ca+Na+K). Major and some trace element concentrations determined by X-ray fluorescence (XRF), REE's, Th, U by instrumental neutron activation analysis (INAA) at the Université de Montréal. Analytical techniques and precisions can be found in Boily and Ludden (1991).

[a] Avg. Sup. is the average composition of the Superior Province continental crust exposed in the Ungava Orogen (Harvey 1995); avg. OWP is the average of the Cretaceous peraluminous granite of the Old Woman Mountains area, southeastern California (Miller and Wooden 1994).

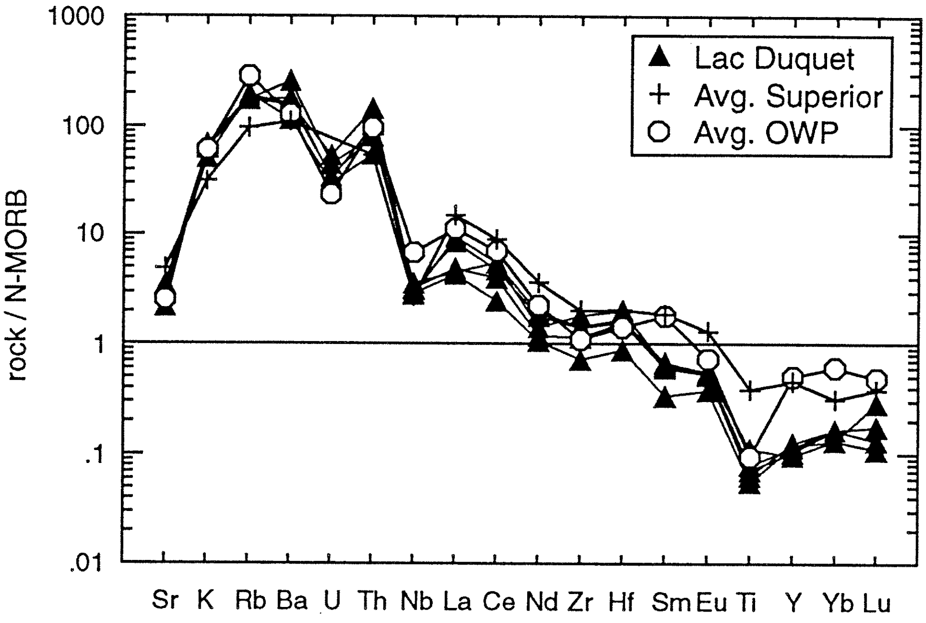

Fig. 3.5 N-MORB-normalized trace element variation diagram of samples from the Lac Duquet monzogranite (filled triangles) with average values of Superior Province continental crust (Harvey 1995) and the crustally-derived peraluminous Old Woman Mountains granites (Miller and Wooden 1994) for comparison. Normalization values are from Sun and McDonough (1989).

Table IV. Nd and Sr isotopic data for whole rock samples from the Lac Duquet pluton.

Sample	Age (Ma)	Nd (ppm)	Sm (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ [a]	$\varepsilon_{Nd\,(t)}$ [b]	Rb (ppm) [c]	Sr (ppm) [c]	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ [d]	$^{87}Sr/^{86}Sr_i$
Lac Duquet monzogranite											
D-19-89	1742	12.502	1.761	0.08513	0.511137	-4.4	97.2	259	1.09	0.73128	0.70402
L-68-89	1742	12.404	1.833	0.08931	0.511157	-4.9	99.9	250	1.16	0.73409	0.70506
L-96-89	1742	8.062	1.583	0.11864	0.511248	-9.7	106.5	195	1.59	0.74404	0.70432
S-44-89	1742	6.410	1.167	0.11002	0.511295	-6.9	110.6	272	1.18	0.73443	0.70489
L-67b-89	1742						98.3	310	0.92	0.72777	0.70475

Notes: Replicate reproducibility is approximately 0.006% for $^{143}Nd/^{144}Nd$, 0.4% for $^{147}Sm/^{144}Nd$, ± 0.5 initial ε_{Nd}, 0.004% for $^{87}Sr/^{86}Sr$. Analytical techniques for the preparation of samples are presented in Boily and Ludden (1991). Nd isotopic analyses were performed at the Geological Survey of Canada while Sr isotopic measurements were undertaken at the Université de Montréal and at the Université du Québec à Montréal.

[a] Normalized to $^{146}Nd/^{144}Nd = 0.7219$.

[b] Normalized to CHUR, present day values of $^{147}Sm/^{144}Nd = 0.1966$, $^{143}Nd/^{144}Nd = 0.512638$. Decay constant for $^{147}Sm = 6.54 \times 10^{-12}$.

[c] Concentrations determined by XRF.

[d] Normalized to $^{86}Sr/^{88}Sr = 0.1194$. Decay constant for $^{87}Rb = 1.42 \times 10^{-11}$.

hypothesis. Initial ε_{Nd} values for the Lac Duquet granite range from -4.4 to -9.7 (Fig. 6 and Table IV) indicating a significant Nd isotopic heterogeneity within the pluton. The range in Nd isotopic composition of the pluton is intermediate between the composition of the Ungava depleted mantle (Dunphy et al. 1995) and the Archean Superior Province continental crust (Harvey 1995) and overlaps with the range in Nd isotopic composition of the Narsajuaq terrane plutons (Dunphy et al. 1995), suggesting a mixed source for the granite (Fig. 6).

3.5 PETROGENESIS AND DISCUSSION

3.5.1 Magma sources

The overall geochemical composition of the Lac Duquet monzogranite is similar to that of average continental crust and crustally-derived granites. Close correspondence in the concentration of the LILE's are noted, although lower REE contents are measured in the Lac Duquet pluton relative to average crustal compositions (Fig. 5). Partial melting of such crust with retention of very small amounts of accessory phases such as monazite, apatite, zircon and titanite in the residue or subsequent fractionation of such phases would result in lower REE contents in the partial melts and would also produce REE patterns that are slightly fractionated and light REE enriched. The presence of such accessory phases in the Lac Duquet monzogranite suggests that they may have played such a role in its petrogenesis. Residual garnet may have also been important in decreasing the heavy REE abundances, but since this mineral strongly fractionates both the light and heavy REE, it would result in strongly fractionated REE patterns in the subsequent melts which is not observed in the Lac Duquet pluton (the heavy REE patterns are relatively flat (Tb_n/Yb_n = 0.8-1.9; Fig. 5).

Based on the documented field relationships, geochronology and geochemistry, several potential sources for the Lac Duquet monzogranite can be identified, namely the Archean Superior Province crust, the Narsajuaq terrane plutonic arc crust, and the lithologies present in the thrust slice of the Cape Smith Belt (primarily mafic metasediments of the Povungnituk Group), into which the pluton was intruded. The

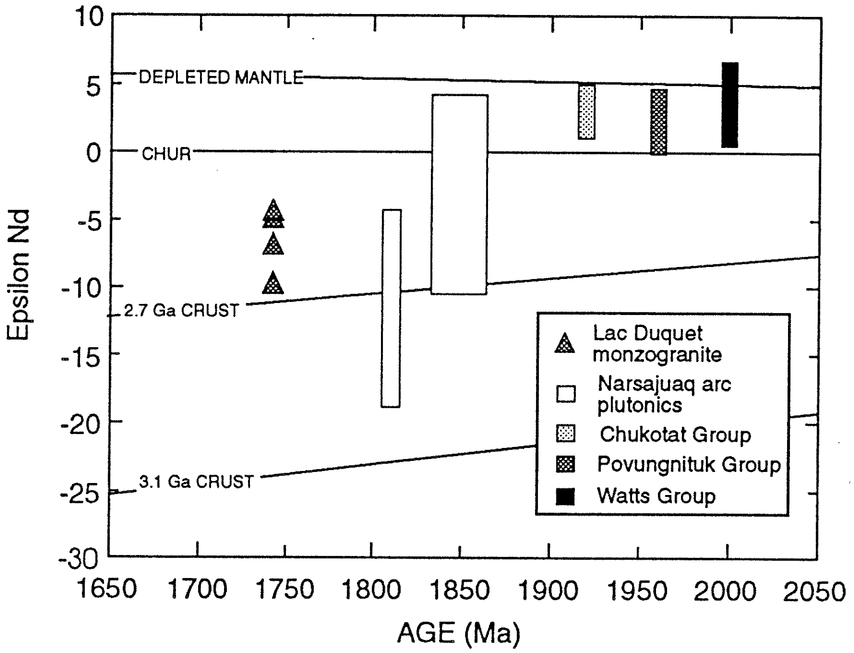

Fig. 3.6. Epsilon Nd (ε_{Nd}) evolution diagram. Samples from the Lac Duquet monzogranite span a range of ε_{Nd} values and may represent melts derived from partial melting of Archean Superior Province crust (either 2.7 Ga crust or 3.1 Ga crust) and Proterozoic Narsajuaq Terrane crust. Also shown for reference are ranges for Cape Smith units (Watts, Povungnituk and Chukotat Groups; Hegner and Bevier 1991) and Narsajuaq terrane lithologies (Dunphy and Ludden, in preparation). "2.7 Ga crust" and "3.1 Ga crust" represent the time-evolved Nd isotopic compositions for ca. 2.7 Ga and 3.1 Ga Superior Province crust from the Minto block (Stern et al. 1994). "Depleted Mantle" is the time-evolved Nd isotopic composition estimated for the Cape Smith depleted mantle (Dunphy et al. 1995).

results of isotope mixing calculations for the Lac Duquet monzogranite using the average composition of the Superior Province continental crust and the Narsajuaq terrane crust as end-members (Table V) are illustrated in figure 7. Two compositions are presented for the Superior Province: the average composition of the 2.7 Ga Minto block crust and the composition of a very old sample (> 3.1 Ga) from the same area (both from Stern et al. 1994). The Minto block is that part of the Superior Province located immediately south of the Ungava orogen (Fig. 1) and its composition is a valid approximation of the composition of the Superior Province end-member. The Nd isotopic composition (Hegner and Bevier 1991) and geochronological data (R. Parrish, unpublished data) for a Povungnituk metasedimentary sample indicate that the source for these sediments is old and isotopically evolved material such as the Superior Province continental crust. The potential isotopic contribution of the Povungnituk Group metasediments can therefore be adequately modelled by the composition of the Superior Province crust. It is clear from figure 7 that the entire spectrum of isotopic compositions of the Lac Duquet pluton cannot be explained by simple mixing of only two sources; all three end-member compositions are required to fully explain all the data. One of the samples (L-68-89) represents a mixture of ≈65% Narsajuaq terrane crust and 35% average Superior Province (Minto block) crust, whereas another sample (L-96-89) represents a magma containing approximately 35% of the older Superior Province crust (Fig. 7). Samples D-19-89 and S-44-89 lie between the three end-members, with D-19-89 apparently containing relatively little Superior Province crust while S-44-89 could represent a melt composed of approximately 65% Narsajuaq terrane crust, 20% average Superior Province crust and 15% old Superior Province crust. These results are compared to isotopic compositions calculated from the distribution of inherited zircons (see below).

3.5.2 Crustal anatexis

Thermal modelling of anatexis during crustal thickening by a variety of authors (England and Thompson 1984; Zen 1988; Patino Douce et al. 1990) has shown that partial melting of crustal rocks can occur during deformation events as a result of conductive heating following crustal thickening. An estimate of the P-T conditions in

Table V. End-member source compositions for mixing calculations (recalculated to 1.742 Ga).

Source	$^{87}Sr/^{86}Sr$	ε_{Nd}	Nd (ppm)	Sr (ppm)
Avg. Narsajuaq terrane crust [a]	0.7035	-2.3	28.6	1067
Avg. (2.7 Ga) Superior crust [b]	0.7139	-11.3	23.5	389
Old (3.1 Ga) Superior crust	0.7074	-24.0	30.0	420

[a] Narsajuaq terrane data from Dunphy and Ludden (in preparation).

[b] Superior Province crust (Minto block) compositions from Stern et al. (1994).

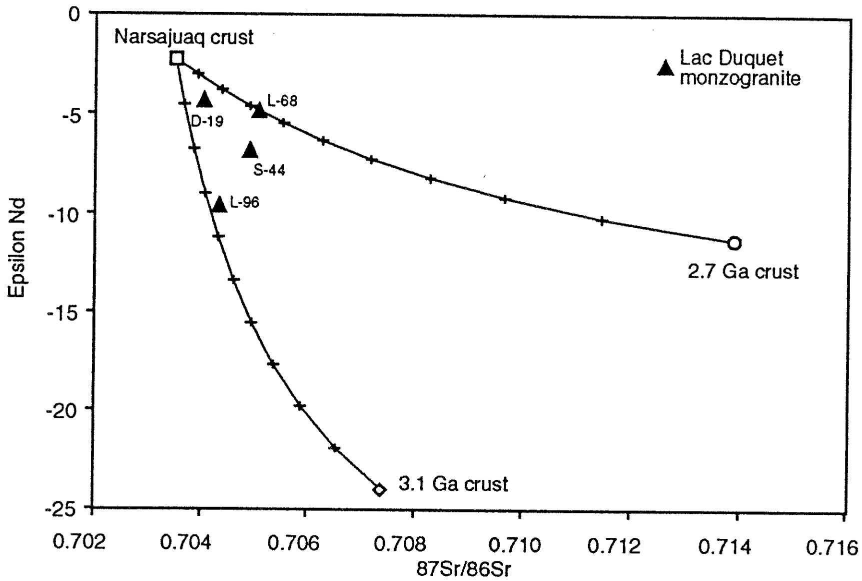

Fig. 3.7. Nd-Sr isotopic evolution diagram showing calculated mixing lines (at 1742 Ma) between the average composition of the Narsajuaq terrane crust (Dunphy and Ludden, in preparation) and two different crustal compositions (average 2.7 Ga and old 3.1 Ga crust) from the Superior Province (Minto block, Stern et al. 1994). End-member compositions are presented in table V. Ticks on the mixing lines represent 10% increments. Crustal anatexis involving various proportions of these three end-members can reproduce the range in Nd and Sr isotopic composition recorded in the samples from the Lac Duquet pluton.

the vicinity of the Lac Duquet pluton can be drawn from the results of a recent study by St-Onge and Lucas (1995) of the P-T conditions of the Superior Province basement exposed in the eastern part of the Orogen (Fig. 1) during the regional-scale retrogression event from granulite facies to amphibolite facies conditions. Their results indicate that the minimum P-T conditions during the retrogression event (ca. 1.81-1.79 Ga; Scott and St-Onge 1995) in the region of the Lac Duquet pluton were approximately 675-700 °C and 8.5-9.8 kbar. Wyllie (1977) has documented that water saturated granitic liquids can be produced by partial melting of continental gneisses at such P-T conditions. Fluids released from the dehydrating continental margin sequence (Povungnituk and Chukotat groups) during the retrogression event may have provided the excess H_2O necessary for crustal melting. It is therefore proposed that heating of the crust during the accretion and thickening event (D_2) as well as the subsequent deformation events (D_3 and D_4) eventually led to anatexis of the surrounding crust producing this granite (and other late granites) and providing a source for their inherited components.

3.5.3 Zircon inheritance

Zircon saturation temperatures for the samples of the Lac Duquet monzogranite calculated using the empirical model of Watson and Harrison (1983) range from ≈700-775 °C (average value of ≈745 °C) which is higher than the estimated melt temperature (minimum of 675-700 °C, see above) for these samples. This indicates that any zircon present in the source region of this granite would not have completely dissolved during the melting and emplacement of the pluton, thereby leading to zircon inheritance in the newly formed magma. These calculations agree with our geochronology studies which illustrate significant older zircon present in the pluton. The majority of the zircons in the sample from the Lac Duquet monzogranite are inherited, with approximately 75% having the same age as the Early Proterozoic lithologies present in the map area (Narsajuaq terrane and Cape Smith Belt units) and the remainder Archean in age, most likely derived from the Superior Province basement.

As a test of the accuracy of the geochronology data to reflect the relative distribution of the various sources involved in the petrogenesis of the Lac Duquet granite, the Nd isotopic composition of L-68-89 was calculated using the proportions of the various sources as indicated by ICP-MS analysis (≈75% Narsajuaq terrane and Cape Smith units, ≈24% average 2.7 Ga Superior Province crust, ≈1% old (>3.0 Ga) crust). Using these proportions and the Nd isotopic composition of the various sources (Table V), the calculated isotopic composition (ε_{Nd} at 1742 Ma) of this sample is -4.7 which is essentially identical (within analytical error) to the measured value (ε_{Nd} = -4.9) and agrees well with the data for this sample from figure 7 which shows virtually no contribution from the older crust and approximately 30% of average Superior Province crust. Sample L-96-89 falls along the mixing line between average Narsajuaq terrane crust and old Superior Province crust and should contain a greater proportion of older zircon (and a lesser amount of 2.7 Ga zircons) compared to L-68-89 in order to account for its isotopic composition. Therefore this sample would probably represent the best target for finding "old" zircons; a task that ICP-MS laser ablation geochronology could quickly and easily verify. Once the presence of such grains has been verified, conventional TIMS geochronology could be employed to precisely date such zircons.

3.6 CONCLUSIONS

The Lac Duquet pluton represents an important example of late magmatic activity in the Ungava Orogen, as this undeformed and post-metamorphic monzogranite cross-cuts the major tectonostratigraphic domains of the orogen. Conventional TIMS U-Pb geochronology on a sample from the pluton has constrained its age of emplacement at 1742 Ma and identified inherited zircon grains. Laser ablation ICP-MS analysis of zircons from the same sample documented a range in ages of 1.7 to 3.2 Ga for the inherited grains. More than 80% of the analyzed grains correspond to the ages of the surrounding host rocks for the granite which includes the Archean Superior Province crust, the Narsajuaq terrane and the Povungnituk Group of the Cape Smith Belt. The laser ablation ICP-MS method is a simple, rapid and relatively inexpensive technique that was capable of identifying the age populations of inherited zircon grains present in the Lac Duquet pluton. Combined with precise U-Pb dating of primary magmatic grains, the ICP-MS analyses were fully able to document the geochronological history of the intrusion. The combined geochronological, isotopic and geochemical data for the Lac Duquet pluton indicate that the surrounding crustal rocks played a significant role in its petrogenesis. Anatexis of the thrust-thickened continental crust during the waning stages of the regional Ungava Orogeny was responsible for the generation of the granitic melts that formed the pluton and accounts for the presence of the inherited zircons and contaminated isotopic and geochemical signatures.

ACKNOWLEDGEMENTS

Field work in northern Quebec was supported by the Geological Survey of Canada (GSC) under the leadership of M.R. St-Onge and S.B. Lucas and the Polar Continental Shelf Project. Analytical work at the Université de Montréal was made possible with the kind assistance of B. Dionne, R. Feng and G. Gauthier. Implementation and development of the $^{207}Pb/^{206}Pb$ dating technique at the Université de Montréal was initiated by R. Feng and G. Gauthier. U-Pb geochronology and Sm-Nd isotopic work was carried out at the Geochronology Laboratory of the GSC with the assistance of R. Parrish and R. Stern. Rb-Sr work at the Université du Québec à Montréal was performed with the help of R. Stevenson and S. Prevec. JMD acknowledges a Natural Sciences and Engineering Research Council (NSERC) doctoral fellowship. JNL acknowledges support from NSERC research grants including a major installation and infrastructure grant to purchase the ICP-MS laser facility. S.B. Lucas is thanked for reviewing earlier versions of this manuscript. Journal reviews by J. David, A. Hynes and particularly an anonymous reviewer greatly improved the manuscript.

REFERENCES

Barrette, P.D. 1994. Lithostratigraphy and map-scale structure in the western Cape Smith Belt, northern Quebec: a tentative correlation between two tectonic domains. Canadian Journal of Earth Science **31**: 986-994.

Boily, M., and Ludden, J.N. 1991. Trace-element and Nd isotopic variations in Early Proterozoic dyke swarms emplaced in the vicinity of the Kapuskasing structural zone: enriched mantle or assimilation and fractional crystallization (AFC) process? Canadian Journal of Earth Science **28**: 26-36.

Compston, W., Williams, I.S., and Meyer, C. 1984. U-Pb geochronology of zircons from lunar breccia 73217 using a sensitive high mass-resolution ion microprobe. Proceedings of the 14th Lunar Science Conference, Journal of Geophysical Research, **89**: B525-B534.

Deniel, C., Vidal, P., Fernandez, A., Le Fort, P., and Peucat, J-J. 1987. Isotopic study of the Manaslu granite (Himalaya, Nepal): inferences on the age and source of Himalaya leucogranites. Contributions to Mineralogy and Petrology **96**: 78-92.

Dunphy, J.M., and Ludden, J.N. 1992. The Narsajuaq terrane, northern Quebec: a mid-crustal section of an Early Proterozoic magmatic arc. Eos Transactions (American Geophysical Union) **73**: 318.

Dunphy, J.M., Parrish, R.R., and Ludden, J.N. 1991. Stitching together the Ungava orogeny: the geochemistry and age relationships of the Lac Duquet pluton. Geological Association of Canada - Mineralogical Association of Canada, Program with Abstracts, **16**: A32.

Dunphy, J.M., Ludden, J.N., and Francis, D. 1995. Geochemistry of mafic magmas from the Ungava orogen, Quebec, Canada and implications for mantle reservoir compositions at 2.0 Ga. Chemical Geology **120**: 361-380.

England, P.C., and Thompson, A.B. 1984. Pressure - temperature - time paths of regional metamorphism I. Heat transfer during the evolution of regions of thickened continental crust. Journal of Petrology, **25**: 894-928.

Feng, R., Machado, N., and Ludden, J.N. 1993. Lead-geochronology of zircon by LaserProbe - Inductively Coupled Plasma Mass Spectrometry (LP-ICP-MS). Geochimica Cosmochimica et Acta, **57**: 3479-3486

Fryer, B.J., Jackson, S.E., and Longerich, H.P. 1993. The application of laser ablation microprobe-inductively coupled plasma-mass spectrometry (LAM-ICP-MS) to in situ (U)-Pb geochronology. Chemical Geology **109**: 1-8.

Francis, D.M., Ludden, J.N., and Hynes, A.J. 1983. Magma evolution in a Proterozoic rifting environment. Journal of Petrology, **24**: 556-582.

Harrison, T.M., Aleinikoff, J.N., and Compston, W. 1987. Observations and controls on the occurrence of inherited zircon in Concord-type granitoids, New Hampshire. Geochimica et Cosmochimica Acta **51**: 2549-2558.

Harvey, D. 1995. Géochimie et traceur isotopique du Nd dans les intrusifs Archéens de la partie Nord-est de la Péninsule d'Ungava. M.Sc. thesis, Université de Montréal, Montréal, PQ.

Hegner, E., and Bevier, M.L. 1991. Nd and Pb isotopic constraints on the origin of the Purtuniq ophiolite and Early Proterozoic Cape Smith Belt, northern Quebec, Canada. Chemical Geology, **91**: 357-371.

Hoffman, P.F. 1990. Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen. *In* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 15-39.

Hynes, A.J., and Francis, D.M. 1982. A transect of the Early Proterozoic Cape Smith foldbelt, New Quebec. Tectonophysics, **88**: 23-59.

Krogh, T.E. 1982. Improved accuracy of U-Pb ages by the creation of more concordant systems using an air abrasion technique. Geochimica Cosmochimica et Acta, **46**: 637-649.

Lamothe, D., Picard, C., and Moorhead, J. 1983. Région du lac Beauparlant, Bande de Cap Smith - Maricourt. Ministère de l'Energie et des Ressources du Québec, DP 84-39.

Le Fort, P. 1981. Manaslu leucogranite: a collision signature of the Himalaya. A model for its genesis and emplacement. Journal of Geophysical Research **86**: 10545-10568.

Lewry, J.F., and Collerson, K.D. 1990. The Trans-Hudson Orogen: extent, subdivision and problems. *In* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 1-14.

Lucas, S.B., and St-Onge, M.R. 1991. Evolution of Archean and Early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec. *In* Current research, part C. Geological Survey of Canada, Paper 91-1C, pp. 109-119.

Lucas, S.B. and St-Onge, M.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 2: Structural and metamorphic history. Canadian Journal of Earth Sciences, **29**: 765-782.

Lucas, S.B., St-Onge, M.R., Parrish, R.R., and Dunphy, J.M. 1992. Long-lived continent-ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. Geology, **20**: 113-116.

Ludwig, K.R. 1990. ISOPLOT - A plotting and regression program for radiogenic-isotope data for IBM-PC compatible computers, version 2.53. United States Geological Survey, Open-file report 88-557.

Machado, N. 1990. Timing of collisional events in the Trans-Hudson Orogen: Evidence from U-Pb geochronology for the New Quebec Orogen, the Thompson Belt, and the Reindeer Zone (Manitoba and Saskatchewan). *In* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 433-441.

Machado, N., David, J., Scott, D.J., Lamothe, D., Philippe, S., and Gariépy, C. 1993. U-Pb geochronology of the western Cape Smith Belt, Canada: New insights on the age of initial rifting and arc magmatism. Precambrian Research, **63**: 211-223.

Miller, C.F., and Wooden, J.L. 1994. Anatexis, hybridization and the modification of ancient crust: Mesozoic plutonism in the Old Woman Mountains area, California. Lithos, **32**: 111-133.

Parrish, R.R. 1989. U-Pb geochronology of the Cape Smith Belt and Sugluk block, northern Quebec. Geoscience Canada, **16**: 126-130.

Parrish, R.R., Roddick, J.C., Loveridge, W.D., and Sullivan, R.W. 1987. Uranium-lead analytical techniques at the geochronology laboratory, Geological Survey of Canada. *In* Radiogenic Age and Isotope Studies: Report 1. Geological Survey of Canada, Paper 87-2, pp. 3-7.

Paterson, B.A., Rogers, G., and Stephens, W.E. 1992. Evidence for inherited Sm-Nd isotopes in granitoid zircons. Contributions to Mineralogy and Petrology, **111**: 378-390.

Patino Douce, A.E., Humphreys, E.D., and Johnston, A.D. 1990. Anatexis and metamorphism in tectonically thickened continental crust exemplified by the Sevier hinterland, western North America. Earth and Planetary Science Letters, **97**: 290-315.

Picard, C., Lamothe, D., Piboule, M., and Oliver, R. 1990. Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt (New-Quebec). Precambrian Research, **47**: 223-249.

Pitcher, W.S., Atherton, M.P., Cobbing, E.J., and Beckinsale, R.D. 1985. Magmatism at a Plate Edge -- The Peruvian Andes. Blackie and Son.

Reid, M.R. 1990. Ionprobe investigation of rare earth element distributions and partial melting of metasedimentary granulites. *In* Granulites and Crustal Evolution. *Edited by* D. Vielzeuf and P. Vidal. Kluwer Academic Publishers, Netherlands, pp. 507-522.

Rive, M., Pintson, H., and Ludden, J.N. 1990. Characteristics of Late Archean plutonic rocks from the Abitibi and Pontaic Subprovinces, Superior Province, Canada. *In* The Northwestern Quebec Polymetallic Belt. *Edited by* M. Rive, P. Verpaelst, Y. Gagnon, J.M. Lulin, G. Riverin and A. Simard. Canadian Institute of Mining and Metallurgy, Special volume 43, pp. 65-76.

Roddick, J.C., and Compston, W. 1977. Strontium isotopic equilibration: a solution to a paradox. Earth and Planetary Science Letters **34**: 238-246.

Roddick, J.C., and Bevier, M.L. 1995. U-Pb dating of granites with inherited zircon: Conventional and ion microprobe results from two Paleozoic plutons, Canadian Appalachians. Chemical Geology **119**: 307-329.

Rogers, J.J.W., and Greenberg, J.K. 1990. Late-orogenic, post-orogenic and anorogenic granites: distinction by major-element and trace-element chemistry and possible origins. Journal of Geology, **98**: 291-309.

Scott, D.J., and St-Onge, M.R. 1995. Constraints on Pb closure temperature in titanite based on rocks from the Ungava Orogen, Canada: implications for U-Pb geochronology and P-T-t path determinations. Geology, in press.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1989. The 1998 Ma Purtuniq ophiolite: imbricated and metamorphosed oceanic crust in the Cape Smith Thrust Belt, northern Quebec. Geoscience Canada, **16**: 144-147.

Scott, D.J., St-Onge, M.R., Lucas, S.B., and Helmstaedt, H. 1991. Geology and chemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. *In* Ophiolite Genesis and Evolution of the Oceanic Lithosphere. *Edited* by T. Peters, A. Nicolas and R.G. Coleman. Kluwer Academic Publishers, Amsterdam, pp. 817-849.

St-Onge, M.R., and Lucas, S.B. 1990a. Early Proterozoic collisional tectonics in the internal zone of the Ungava (Trans-Hudson) orogen, Lacs Nuvilik and Sugluk map areas, Quebec. *In* Current research, part C. Geological Survey of Canada, Paper 90-1C, pp. 119-132.

St-Onge, M.R., and Lucas, S.B. 1990b. Evolution of the Cape Smith Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. *In* The Early Proterozoic Trans-Hudson Orogen of North America. *Edited* by J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 313-351.

St-Onge, M.R., and Lucas, S.B. 1992. New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. *In* Current research, part C. Geological Survey of Canada, Paper 92-1C, pp. 31-41.

St-Onge, M.R., and Lucas, S.B. submitted ms. Large-scale fluid infiltration, metasomatism and re-equilibration of Archean basement granulites during Paleoproterozoic thrust belt construction, Ungava Orogen, Canada. Journal of Metamorphic Geology.

St-Onge, M.R., Lucas, S.B., Scott, D.J., Bégin, N.J., Helmstaedt, H., and Carmicheal, D.M. 1988. Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec. *In* Current research, part C. Geological Survey of Canada, Paper 88-1C, pp. 1-18.

St-Onge, M.R., Lucas, S.B., and Parrish, R.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Canadian Journal of Earth Science, **29**: 746-764.

Stern, R., Percival, J., and Mortensen, J. 1994. Geochemical evolution of the Minto block: a 2.7 Ga continental magmatic arc built on the Superior proto-craton. Precambrian Research, **65**: 115-153.

Streckeisen, A.L. 1976. To each plutonic rock its proper name. Earth Science Reviews, **12**: 1-33.

Sun, S.S., and McDonough, W.F. 1989. Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. *In* Magmatism in the ocean basins. *Edited by* A.D. Saunders and M.J. Norry. Geological Society, Special Publication 42, pp. 313-345.

Watson, E.B., and Harrison, T.M. 1983. Zircon saturation revisited: temperature and composition effects in a variety of crustal magma types. Earth and Planetary Science Letters, **64**: 295-304.

Williams, I., and Claesson, S. 1987. Isotopic evidence for the Precambrian provenance and Caledonian metamorphism of high grade paragneisses from the Seve Nappes, Scandinavian Caledonides. Contributions to Mineralogy and Petrology, **97**: 205-217.

Wyllie, P.J. 1977. Crustal anatexis: an experimental review. Tectonophysics **43**: 41-71.

Zen, E.-a. 1988. Thermal modelling of stepwise anatexis in a thrust-thickened sialic crust. Transactions of the Royal Society of Edinburgh: Earth Sciences, **79**: 223-235.

CHAPTER 4

Évolution des roches plutoniques du domaine nord de la Fosse de l'Ungava

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Rapport du Ministère, publié en novembre 1994

RÉSUMÉ

Une variété de plutons ont été délimités dans le domaine nord de la Fosse de l'Ungava. Les plutons sont plutôt restreints aux groupes de Watts, de Spartan et de Parent et sont caractérisés par une composition dioritique à granodioritique. Ils présentent des taux de déformation et de métamorphisme variables. Dans la plupart des échantillons, la minéralogie primaire composée d'hornblende et/ou de biotite a été remplacée par un assemblage d'épidote, de chlorite et d'actinote. La saussuritisation du plagioclase est répandue. Les plutons du Groupe de Watts sont datés entre 1880 et 1840 Ma tandis que les âges des plutons du Groupe de Parent s'échelonnent entre 1898 et 1845 Ma. Un seul pluton a fait l'objet de datation U-Pb au sein du Groupe de Spartan, pour lequel un âge minimum de 1850 Ma a été obtenu.

Les plutons sont surtout des intrusions calco-alcalines métalumineuses caractérisées par l'enrichissement variable de certains éléments hygromagmaphiles tels que le Ba, le Th et le U. La plupart des échantillons sont relativement appauvris en Nb et en Ti et présentent des anomalies négatives pour ces éléments sur les diagrammes étendus d'éléments trace. Généralement, les échantillons plus felsiques sont caractérisés par des concentrations faibles de Ti et plus élevées d'éléments hygromagmaphiles. Les échantillons de tonalite et de granodiorite ont typiquement des terres rares légères fractionnées (La_n/Yb_n jusqu'à 32,3) tandis que les plutons plus mafiques ont généralement des spectres de terres rares plus atténuées (La_n/Yb_n = 2,5 à 8,2). L'abondance des terres rares (particulièrement les terres rares lourdes) est généralement plus faible dans les échantillons de tonalite et de granodiorite, suggérant le rôle important qu'a joué le fractionnement du grenat dans leur pétrogenèse. Les plutons sont juvéniles et possèdent des signatures isotopiques similaires quel que soit leur âge ou leur composition.

La similarité des compositions et des signatures géochimiques des plutons de la Fosse de l'Ungava et celles des plutons du terrane de Narsajuaq suggère qu'ils sont issus de processus pétrogénétiquement similaires. De plus, les plutons présents à l'intérieur du Groupe de Parent sont géochimiquement comparables aux roches volcaniques hôtes. Les lithologies du terrane de Narsajuaq et du Groupe de Parent ont été interprétées précédemment comme produits d'un arc magmatique. La composition des plutons de la Fosse de l'Ungava étant similaire à celle des roches du terrane de Narsajuaq et du Groupe de Parent ainsi que la proche association géographique et temporelle de ces unités suggèrerait qu'elles sont toutes des expressions de la même activité magmatique. Ceci suggère aussi que les plutons trouvés dans les groupes de Watts, de Spartan et de Parent ont été générés dans un environnement de subduction. Ces données sont compatibles avec le modèle suggéré par St-Onge *et al.* (1992) et impliquent la subduction d'une croûte océanique vers le nord, sous les groupes de Watts et de Spartan. Cette subduction responsable de la production des plutons étudiés aurait aussi généré l'arc volcanique/plutonique de Parent.

ABSTRACT

Within the Cape Smith Belt a variety of individual discrete plutons have been delimited occurring primarily in the central to western portion of the belt. The plutons are only found in the Watts, Spartan and Parent Groups and range in composition from diorites to granodiorites and are preserved in varrying states of deformation and metamorphism. In most samples the primary mineralogy of hornblende and/or biotite has been replaced by variable amounts of epidote, chlorite and actinolite and saussuritiztion of plagioclase is pervasive. Plutons in the Watts Group range in age from ca. 1880-1840 Ma while the Parent Group plutons are 1898-1845 Ma old. Only one age of >1850 Ma constrains the timing of plutonism within the Spartan Group.

The plutons are predominantly calc-alkaline metaluminous intrusives characterized by variable enrichment of certain large ion lithophile elements such as Ba, Th and U. Most samples have a relative depletion of Nb and Ti (seen as negative anomalies on spiderdiagrams). In general lower abundances of Ti and higher concentrations of the LILE's are seen in the more felsic samples. Tonalite and granodiorite samples typically have fractionated REE's (La_N/Yb_N up to 32.3) whereas the

more mafic plutons generally have flatter REE patterns ($La_N/Yb_N = 2.5\text{-}8.2$). The overall abundance of the REE's (especially the heavy REE) is lower in tonalite and granodiorite samples suggesting that garnet fractionation was important in their petrogenesis. Isotopcially the plutons are juvenile and have similar isotopic signatures regardless of their age or composition.

The overall similarity of the composition and geochemical signature of the Cape Smith Belt plutons and the Narsajuaq terrane plutonics suggests that they had a similar petrogenesis. In addition the plutons present within the Parent Group are geochemically comparable to the volcanics that they intrude. Both the Narsajuaq terrane and Parent Group lithologies have previously been interpreted as arc-related magmatic products. The similarity in composition as well as the close spatial and temporal association of the Cape Smith Belt plutons to the Narsajuaq terrane and Parent Group rocks suggests that they are expressions of the same magmatic activity and that the plutons found within the Watts, Spartan and Parent Groups were generated in a subduction-related environment. A model consistent with that of St-Onge *et al.* (1992) is proposed involving northward subduction of oceanic crust under the Watts and Spartan Groups which produced the plutons present within these units and also generated the volcanic/plutonic Parent arc.

4.1 INTRODUCTION

4.1.1 Travaux réalisés

La Fosse de l'Ungava du Québec septentrional correspond à la zone externide de l'orogène de l'Ungava d'âge Protérozoïque inférieur (Fig. 1; St-Onge et Lucas, 1990a). Ce dernier segment orogénique est lui-même une partie intégrante de l'orogène Trans-Hudsonien, un des orogènes du Protérozoïque inférieur des plus longs et des mieux exposés au monde (Lewry et Collerson, 1990; Hoffman, 1990).

Dans la bande du Cap Smith, une variété de plutons occupent principalement la région centrale à occidentale du domaine nord de l'orogène (Fig. 2). Les plutons varient de composition péridotitique à granitique et sont préservés dans des états de déformation

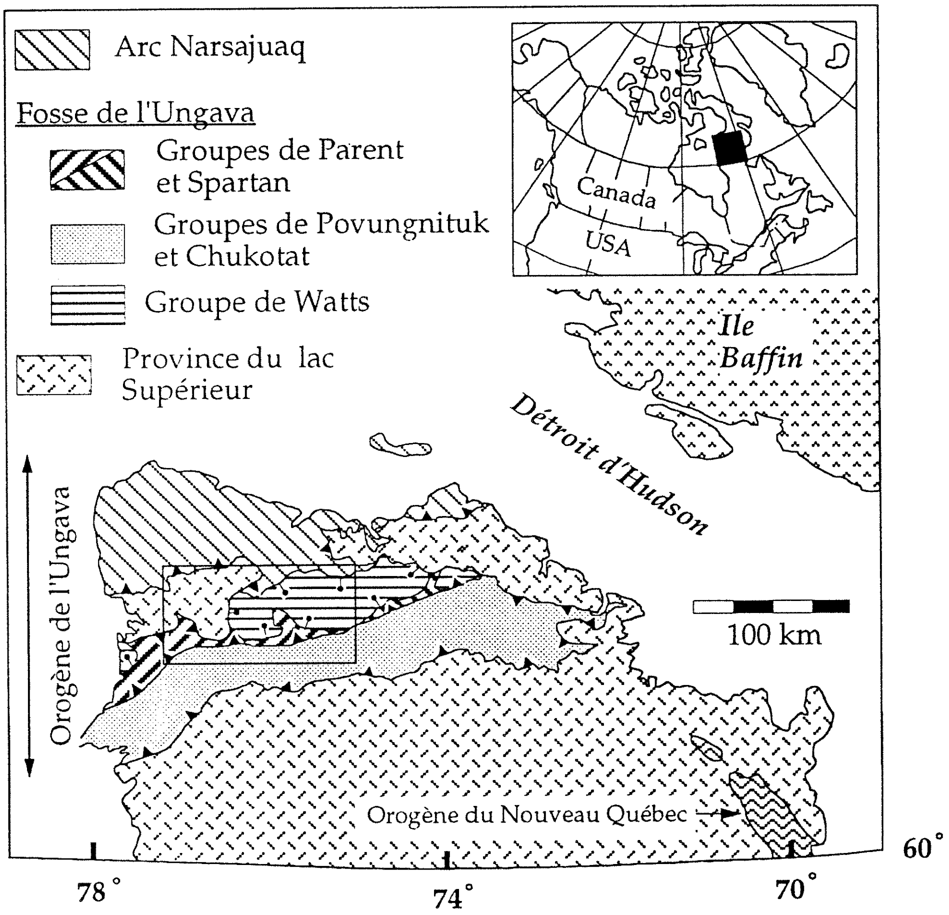
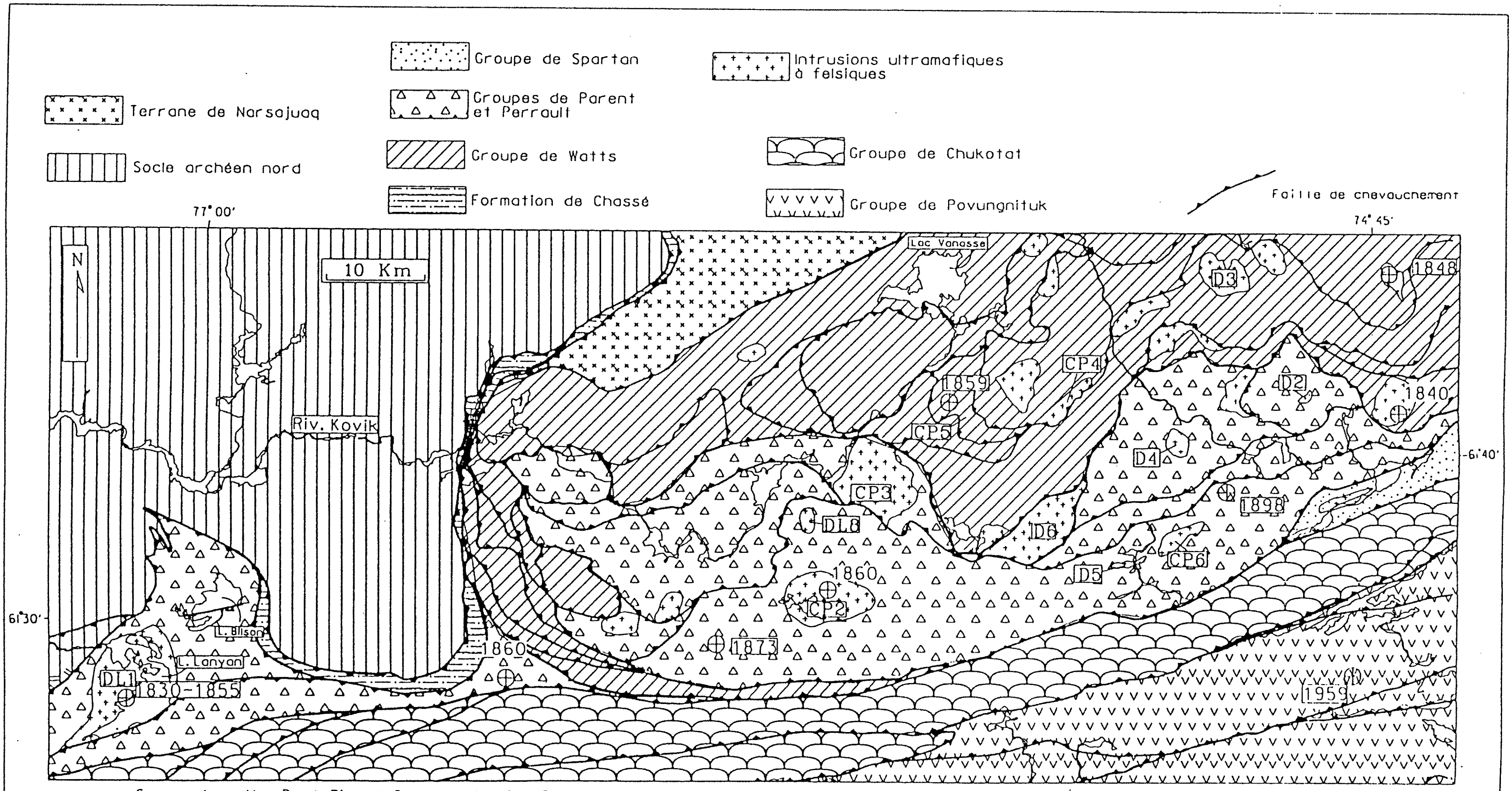


Figure 4.1. Carte localisant l'orogène de l'Ungava et ses éléments tectoniques (de St-Onge et Lucas, 1992). La région étudiée (voir Fig. 2) est indiquée.



Source: Lamothe, D. et Picard, C.., en rédaction, Synthèse géologique de la Fosse de l'Ungava, Ministère des Ressources Naturelles, MB 95-01.

Figure 4.2. Carte de localisation des plutons de la Fosse de l'Ungava (basée sur une compilation de Lamothe et Picard, en rédaction). Les lettres réfèrent aux noms des plutons et les chiffres représentent les âges obtenus par Parrish (1989) et Machado *et al.* (1993) par la méthode de géochronologie U-Pb.

et de métamorphisme variés. Ils sont généralement restreints aux groupes de Watts, de Spartan et de Parent alors que les groupes de Chukotat et de Povungnituk, structuralement sous-jacents, en sont dépourvus.

Ce rapport caractérise neuf plutons en utilisant des données de terrain, des lames minces (50) et des analyses géochimiques (43). Le but de ce travail est de présenter une synthèse de l'activité plutonique dans la Fosse de l'Ungava. La carte de localisation des plutons (Fig. 2) provient d'une compilation de Lamothe et Picard (en rédaction) elle-même basée sur les cartes et rapports de Lamothe *et al.* (1983), Hervet (1985), Tremblay (1987, 1989, 1991) et Barrette (1988). La carte et le sommaire des descriptions de terrain pour chaque pluton ainsi que les lames minces et les analyses géochimiques ont été mises à la disponibilité de l'auteure par le Ministère de l'Énergie et des Ressources du Québec (MERQ). Ces données constituent la base des interprétations présentées dans ce rapport. La caractérisation des plutons et leur relation les uns aux autres ainsi qu'à la séquence magmatique régionale (*e.g.* roches volcaniques du Groupe de Parent et plutons du terrane de Narsajuaq) sont fondamentales et vise une meilleure compréhension de l'évolution tectonique de l'orogène de l'Ungava.

4.1.2 Travaux antérieurs

La Péninsule d'Ungava, située dans le nord du Québec, a longtemps été le sujet d'investigations géoscientifiques, commençant avec les observations de Bell (1885) et de Low (1899, 1902) dans la région du détroit d'Hudson et de la baie d'Hudson. Plus tard, la cartographie de la bande du Cap Smith par une série de travailleurs (Bergeron, 1957, 1959; Beall, 1959, 1960; DeMontigny, 1959; Stam, 1961; Gélinas, 1962; Gold, 1962) aida à identifier le caractère structural de la région et fut suivie de cartographie à l'échelle de reconnaissance par Taylor (1974, 1982). D'autres études stratigraphiques et géochimiques furent entreprises par Baragar (1974, 1984), Schwarz et Fujiwara (1977), Francis et Hynes (1979), Hynes et Francis (1982), et Francis *et al.* (1983). Par la suite, la cartographie détaillée a été faite par le MERQ dans les portions occidentales et centrales de la bande (Lamothe *et al.*, 1983, 1986; Hervet, 1985; Moorhead, 1985, 1987, 1988, 1989; Tremblay, 1987, 1989, 1991; Barrette, 1988, 1989, 1990; Roy, 1989; Picard, 1989b, 1989a; Picard *et al.*, 1990) et la Commission Géologique du Canada (CGC) dans la région orientale (St-Onge *et al.*, 1986, 1987, 1988; St-Onge et Lucas, 1989; 1990b, 1993) et au nord de la Fosse de l'Ungava (St-Onge et Lucas, 1990a, 1992; Lucas et St-Onge, 1991; St-Onge *et al.*, 1992).

4.1.3. Remerciements

Le MERQ a mis à la disponibilité de l'auteure les cartes, échantillons, analyses géochimiques et informations complémentaires nécessaires à la rédaction ce rapport. D. Lamothe est remercié pour sa coopération, de son appui et plus particulièrement pour avoir répondu à maintes questions. C. Picard a gracieusement fait part des données et des informations pertinentes à son travail sur le Groupe de Parent en voie d'exécution. Les discussions avec J.N. Ludden, M.R. St-Onge et S.B. Lucas ont aidé à développer le modèle présenté ici. Le soutien logistique sur le terrain par la CGC sous la direction de M.R. St-Onge et de S.B. Lucas est grandement apprécié. La thèse de doctorat de l'auteure (sous la direction de J.N. Ludden, Université de Montréal) est basée sur ces travaux de terrain et sur les analyses à l'Université de Montréal, celles-ci obtenues avec l'aide de B. Dionne et G. Gauthier. Ces études analytiques portant sur la pétrogenèse du terrane de Narsajuaq, ainsi que les travaux de terrain, ont permis les comparaisons du plutonisme de la Fosse de l'Ungava et du terrane de Narsajuaq présentées dans ce rapport. L'auteure reconnait une bourse d'études post-graduées du CRSNG. La traduction du manuscrit a été faite par A. Sherman, qui a aussi contribué à la clarification de certaines des idées présentées ici. N. Bouchard est remerciée pour son aide lors de la mise en page et pour les retouches finales apportées au document.

4.2 CADRE GÉOLOGIQUE

L'orogène de l'Ungava a été divisé en trois unités tectonostratigraphiques (St-Onge *et al.*, 1992; Fig. 1): (1) roches plutoniques et supracrustales (par)-autochtones d'âge archéen du socle de la Province du Supérieur; (2) roches sédimentaires et volcaniques autochtones et allochtones associées à une marge de rift et de dérive d'env.

2,04 - 1,92 Ga (groupes de Povungnituk et de Chukotat); et (3) composantes crustales suspectes allochtones provenant d'une ophiolite de d'env. 2,00 Ga (Groupe de Watts) et d'un arc magmatique d'env. 1,90 - 1,81 Ga (arc de Narsajuaq et Groupe de Parent) et des roches associées à cet arc (Groupe de Spartan).

Les descriptions détaillées de la stratigraphie des unités de la bande du Cap Smith (groupes de Povungnituk, de Chukotat, de Spartan, de Parent et de Watts) sont présentées ailleurs (St-Onge et Lucas, 1990b, 1993; Picard *et al.*, 1990) et ne sont que résumées ici. Situé à la base de la Fosse de l'Ungava, le Groupe de Povungnituk consiste en une série de séquences autochtone et allochtones d'unités clastiques métasédimentaires. Celles-ci sont recouvertes par des coulées mafiques massives et en coussin ainsi que par des sédiments clastiques interdigités (Lamothe, *et al.*, 1983; St-Onge *et al.*, 1988; Picard, 1989a, 1989b). On y retrouve aussi une séquence alcaline composée de laves (basanites, néphélinites, phonolites) et de roches pyroclastiques (Gaonac'h *et al.*, 1989, 1992; Picard, 1989b; Picard *et al.*, 1990). Le Groupe de Povungnituk est recouvert par le Groupe de Chukotat, une unité principalement ignée qui comprend un minimum de 5 km de coulées de basaltes sous-marines en coussin et massives, et de filons-couches mafiques/ultramafiques avec une quantité mineure de sédiments associés (St-Onge *et al.*, 1987; St-Onge et Lucas, 1990b; Picard, 1989a, 1989b; Picard *et al.*, 1990). Les groupes de Povungnituk et de Chukotat sont en contact de chevauchement. Juxtaposée tectoniquement à ce dernier, on retrouve une séquence distinctive de roches mafiques et ultramafiques stratifiées, de coulées de basalte massives et en coussin, de filons-couches mafiques, de dykes en feuillets et de rares intrusions de plagiogranite (le Groupe de Watts; Lamothe *et al.*, 1983), interprétée comme une ophiolite d'épaisseur préservée de 7,5 à 8 km (St-Onge *et al.*, 1988; Scott *et al.*, 1991, 1992; St-Onge et Lucas, 1990b). Le Groupe de Parent est intercalé entre les groupes de Watts et de Chukotat dans laportion occidentale de la Fosse de l'Ungava. Cette unité consiste de shales, siltstones et grauwackes interlités avec des laves basaltiques à rhyolitiques (le basalte andésitique étant la composition la plus commune), et des tufs volcaniclastiques (Tremblay, 1989, 1991; Picard *et al.*, 1990; Picard, en préparation; St-Onge et Lucas, 1992). Plus à l'est, le Groupe de Spartan, une séquence sédimentaire clastique de pélites, de semipélites et de grès interlités, est observée à un niveau stratigraphique équivalent au Groupe de Parent (Lamothe *et al.*, 1983; St-Onge *et al.*, 1988).

4.3 PÉTROGRAPHIE ET GÉOCHRONOLOGIE DES PLUTONS DE LA BANDE DU CAP SMITH

4.3.1 Plutons du Groupe de Watts

Plusieurs larges plutons sont encaissés par les roches mafiques et ultramafiques stratifiées et les coulées de basalte coussinnées et massives du Groupe de Watts (Lamothe *et al.*, 1983; St-Onge *et al.*, 1987, 1988; Scott *et al.*, 1989, 1991). Ce rapport se concentre sur quelques intrusions mieux définies. Situé dans le secteur nord-est de la région étudiée (Fig. 2), le pluton **D3** d'approximativement 5 km de diamètre, est composé de monzodiorite quartzique à granodiorite à grains fins à moyens (notez que tous les plutons étant métamorphisés au des faciès shistes verts, le préfixe "méta" a été abandonné). Des dykes de diabase recoupés par des dykes granitiques (pegmatitiques?) sont notés à l'intérieur du pluton. Ce pluton possède une foliation bien développée, surtout le long de son contact avec la lithologie encaissante, les gabbros stratifiés du Groupe de Watts. L'étude des lames minces révèle la présence de grains de quartz déformés (10 à 20%) mais ne montre pas la foliation observée sur le terrain. Les minéraux mafiques sont représentés par 10% de biotite et par 5 à 20% d'actinote et d'épidote remplaçant la hornblende et la biotite. Le plagioclase (40 à 65%) et le feldspath potassique (10 à 15%) ont aussi subi l'effet de l'altération sous forme de séricitisation. Les minéraux accessoires incluent la sphène, l'apatite, et des oxydes opaques.

Plus à l'ouest, à proximité du lac Perrault, le pluton **CP5** bordé des failles fait aussi intrusion dans les gabbros stratifiés du Groupe de Watts (Fig. 2). Le pluton est faiblement folié, sauf à son contact avec le Groupe de Watts où une déformation plus prononcée est enregistrée. D'ouest en est, les échantillons du pluton passent d'une composition dioritique à monzodioritique (plus potassique) et sont moyennement à grossièrement grenus. La diorite est composée de 40% de plagioclase maclé et de 60% de minéraux mafiques, surtout du clinopyroxène et de l'orthopyroxène, ainsi qu'une quantité mineure d'hornblende, d'épidote, de chlorite et d'opaques. Les échantillons monzodioritiques contiennent de larges cristaux de perthite, de plus petits cristaux de microcline et de plagioclase (un total de 40 à 60%) et moins de minéraux mafiques que les gabbros. Le pyroxène demeure le minéral mafique principal mais la biotite et l'hornblende sont aussi notés. CP5 a été daté par Machado *et al.* (1993) à 1858 ± 2 Ma.

À l'extrémité nord-est de la région étudiée, un troisième pluton faisant intrusion dans les gabbros stratifiés du Groupe de Watts a été daté par Parrish (1989; échantillon Z2-84) à 1848 +6/-5 (Fig. 2) mais n'a pas été échantillonné par le MERQ. Feininger (1986) a décrit ce pluton comme une alaskite. Dunphy, par ailleurs, (données non-publiées) l'identifie comme une tonalite - granodiorite à clinopyroxène, massive et non-foliée.

Un pluton situé directement au sud de Z2-84 a été daté à env. 1840 Ma (Fig. 2), aussi par Parrish (1989; échantillon Z4-84). Cet échantillon est décrit comme une diorite quartzique massive par Parrish (1989) mais Lamothe (en rédaction) décrit plutot l'intrusion comme un assemblage de syénite et de granite.

Deux plutons tonalitiques se situent près du lac Watts à l'est de la figure 2. Tous les deux recoupent des failles de chevauchement précoces mais sont eux-mêmes charriés par des chevauchements tardifs. Les plutons sont âgés de 1876 ± 1,5 et 1870 à 1880 Ma (Parrish (1989); échantillons S162A-86 et P23-87) et un semble avoir hérité de zircons plus anciens.

4.3.2 Plutons du Groupe de Spartan

Structuralement sous-jacent au Groupe de Watts est le Groupe de Spartan, y composé d'une séquence clastique sédimentaire de pélites interlitées avec des semipélites et une moindre quantité de grès (Lamothe *et al.*, 1983; St-Onge *et al.*, 1988). Seules trois des intrusions présentes dans le Groupe de Spartan sont discutées en détail dans ce rapport dû au peu de données disponibles. Le pluton **D2** est composé de petites intrusions kilométriques possédant chacune des contacts nets avec la roche encaissante (Fig. 2). À l'échelle de l'affleurement, le pluton est décrit comme une diorite massive et non-foliée, composée d'amphibole (50 à 60%) et de plagioclase (35 à 40%). L'amphibole n'est pas distribuée de façon uniforme dans la roche, mais est regroupée localement en amas. Aucune lame mince n'est disponible pour une étude pétrographique.

À l'ouest de D2 on retrouve **D4** , un petit pluton (d'approximativement 4 km de diamètre) de tonalite à diorite quartzique intrusif dans les grauwackes du Groupe de Spartan (Fig. 2). Les sédiments ont été affectés par un métamorphisme de contact. L'étude de terrain a révélée que la bordure du pluton est plus mafique (dioritique) et que le noyau est plus felsique (tonalitique). Les lames minces provenant des échantillons de D4 sont pour la plupart des diorites quartziques, à l'exception d'un échantillon tonalitique. Les lames minces contiennent 10 à 25% de quartz déformé, 35 à 50% de plagioclase fortement altéré, et 25 à 40% de minéraux mafiques primaires telles que l'hornblende et la biotite, quoiqu'un remplacement secondaire par l'actinote, l'épidote et la chlorite est commun. Les minéraux accessoires incluent le sphène, les oxydes opaques, l'apatite et la calcite. La plupart des roches présentent, en lame mince, des textures finement à moyennement grenues et une foliation faible et variable dans certains échantillons.

Le troisième pluton étudié dans le Groupe de Spartan est **CP3**, décrit comme un pluton de tonalite massive à grains moyens. Ce pluton d'approximativement 10 km de large est situé près du lac Belleau au centre de la région cartographiée (Fig. 2). Les lames minces provenant de ce pluton révèlent une faible foliation définie par l'alignement de la biotite (15 à 20%) avec des quantités mineures de chlorite et d'épidote. Deux cristaux de grenat syn- à post-métamorphiques ont aussi été trouvés. Les minéraux leucocratiques sont représentés par de 30 à 35% de quartz déformé, 35% de plagioclase fortement séricitisé et dans un cas, 5% de feldspath potassique sous forme de microcline et de microperthite. Le sphène, l'apatite et la muscovite sont trouvées accessoirement dans tous les échantillons.

Une date imprécise minimum de 1850 Ma a été obtenue par Parrish (1989) pour un autre pluton de tonalite massive (Z5-84) situé dans le Groupe de Spartan.

4.3.3 Plutons du Groupe de Parent

Dans les roches volcaniques, volcaniclastiques et sédimentaires du Groupe de Parent (Tremblay, 1989; Picard *et al.*, 1990; St-Onge et Lucas, 1992; Picard, en préparation) cinq larges intrusions sont identifiées. **D5** et **CP6** font partie d'un cortège d'intrusions lenticulaires (Fig. 2) et foliées qui s'étendent sur 4 à 5 km et qui pourraient représenter des fragments d'un pluton déformé et boudiné. La composition des intrusions varie entre la diorite - diorite quartzique et la tonalite. La diorite est composée de 60 à 70% de minéraux mafiques secondaires (épidote et chlorite) et 30% de plagioclase fortement saussuritisé. La diorite quartzique contient 10% de quartz recristallisé, 40% de plagioclase maclé faiblement altéré, 25% de biotite avec un remplacement variable par la chlorite, 15% d'hornblende remplacée par l'actinote et la chlorite, et 1% de grenat fragmenté (aucune couronne de réaction n'a été observée). La tonalite est caractérisée par l'abondance de plagioclase altéré (60%), de quartz déformé (25%) et 10% de chlorite remplaçant la biotite. L'étude des lames minces de ce pluton n'a pas révélé de foliation, quoique la présence de quartz déformé suggère l'effet d'une déformation. La géochronologie U-Pb d'une petite intrusion de tonalite au nord-est de D5 et CP6 (mais qui n'apparaît pas sur la carte) par Parrish (1989; échantillon Z3-84) a produit un âge de 1898 +12/-9 Ma. Cette tonalite a été interprétée comme une partie de l'intrusion composée D5 et CP6 par Lamothe et Picard (en rédaction; Fig. 2) et par cette relation, pourrait contraindre l'âge du Groupe de Parent. Cependant, l'interprétation des roches de cette région par Parrish (1989) et St-Onge (communication personnelle, 1994), indiquerait plutôt que cette tonalite est encaissée par les lithologies du Groupe de Watts et non du Groupe de Parent.

Situé plus à l'ouest et à proximité du lac Étrange, le large pluton mafique, **CP2**, d'approximativement 10 km de diamètre, est intrusif dans les roches pyroclastiques du Groupe de Parent (Fig. 2). Il est composé de diorite massive, non-foliée, à grains moyens, et comprend 40% à 80% d'hornblende et 15 à 50% de plagioclase. Les lames minces montrent une composition généralement dioritique et des proportions plus ou moins égales de plagioclase fortement séricitisé, d'hornblende (partiellement altéré en chlorite et en épidote) et des quantités mineures de biotite, d'apatite, d'opaques et de sphène. Deux lames minces contiennent aussi une faible quantité d'orthoclase (<10%). Localement, la diorite contient des enclaves contenant jusqu'à 80% d'hornblende et des quantités mineures d'épidote et de chlorite. Par endroits, le pluton est envahi par un mésogabbro à grains grossiers. CP2 a été daté à 1860 ±2 Ma par Machado *et al.* (1993).

Au nord de CP2, et toujours à l'intérieur des roches pyroclastiques du Groupe de Parent, se trouve **DL8**, un petit pluton de tonalite et de diorite faiblement foliées à massives (Fig.2). La tonalite est composée de 20 à 25% de quartz déformé, 60 à 70% de plagioclase altéré et 10 à 15% d'hornblende remplacée par la chlorite et l'épidote. La diorite est essentiellement dépourvue de quartz. Un échantillon de diorite leucocratique contient 60% de plagioclase (contenant beaucoup de séricite, d'épidote, de clinozoisite, de chlorite, etc.) et 35% d'hornblende (et un peu de chlorite); un second échantillon plus mélanocratique se compose de 60% de chlorite et d'épidote et 40% de plagioclase fortement altéré. Le pluton est recoupé par un dyke de gabbro quartzique à grains fins.

Un quatrième pluton dans le Groupe de Parent, daté à 1874 +4/-3 Ma par Machado *et al.* ((1993); échantillon DL5) est trouvé au sud-ouest (Fig. 2) et est décrit comme une diorite. Aucune lame mince ou donnée géochimique n'est disponible pour ce pluton.

Un autre pluton du Groupe de Parent, **DL1**, est situé plusieurs dizaines de kilomètres plus à l'ouest, près du lac Lanyan (Fig. 2). Ce large pluton de granodiorite fait intrusion dans les tufs du Groupe de Parent et recoupe une intrusion dioritique le long de sa marge sud-ouest: des enclaves de diorite sont trouvées dans la granodiorite près de leur contact. L'étude de terrain a démontré que la granodiorite et la diorite sont toutes deux déformées et possèdent une foliation semblable. Au contact légèrement cisaillé des deux lithologies, la hornblende a réagi pour former de la biotite, associée au développement du grenat. Les lames minces dioritiques contiennent 50 à 70% d'hornblende altérée en actinote et en épidote, 20 à 45% de plagioclase non-altéré, et des quantités mineures d'oxydes opaques, de biotite, de sphène et de calcite. Les échantillons de granodiorite contiennent de 40 à 50% de quartz, 20 à 30% de plagioclase maclé, 10 à 20% de microcline et de perthite ainsi que 10 à 25% de minéraux mafiques (biotite, épidote, chlorite et sphène) et de muscovite. Dans ces lames minces, l'alignement de la biotite et de la muscovite définit une faible foliation; par contre, aucune foliation n'est remarquée dans les lames minces de diorite. DL1 a été daté par Machado *et al.* (1993) à 1845 ± 6 Ma.

4.4 GÉOCHIMIE

4.4.1 Méthodes analytiques

Les analyses géochimiques ont été entreprises au Centre de Recherche Minérale du Québec par fluorescence à rayons-X (XRF) pour les éléments majeurs, par spectrométrie d'émission atomique au plasma (SEAP) pour les éléments trace, et par activation neutronique (AN) pour les terres rares. Des variations irrégulières dans la concentration de terres rares plus lourdes tels le Ho et le Tm (*e.g.* figures 6c et 7f) sont attribuées à des difficultés analytiques. Les résultats d'analyses géochimiques et les concentrations d'éléments majeurs recalculées (100% anhydre) sont présentés en Annexes I et II (de cette chapitre).

4.4.2 Généralités

Les plutons de la Fosse de l'Ungava possèdent des caractéristiques géochimiques calco-alcalines et tholéiitiques telles que définies par le diagramme AFM de Irvine et Baragar (1971; Fig. 3a). Le contenu de potassium est intermédiaire dans la plupart des plutons, comme on l'observe sur le diagramme de K_2O en fonction de SiO_2 de Le Maitre (1989; Fig. 3b). La majorité des plutons sont sub-alcalins et métalumineux, bien que plusieurs échantillons (provenant surtout d'un seul et même pluton) sont légèrement peralumineux (Fig. 3c et d). Dans tous les plutons, une haute concentration de SiO_2 est accompagnée par une faible concentration de Fe_2O_3, de MgO, de CaO, de TiO_2 et de MnO et par un accroissement en Na_2O et en K_2O (Fig. 4). La déviation de certains échantillons de la tendance générale est due à l'altération. Celle-ci se manifeste par la présence de minéraux secondaires tels que l'épidote, la chlorite, l'actinote et la séricite et est particulièrement évidente dans les échantillons plus mafiques. En ce qui concerne les éléments trace, la plupart des échantillons contiennent moins de 200 ppm de Cr, 60 ppm de Ni, 200 ppm de Cu et 1200 ppm de Ba (annexe 2); les exceptions sont notées dans les paragraphes suivants. Les données géochimiques sont présentées dans des diagrammes étendus d'éléments trace normalisés aux chondrites (Sun et McDonough, 1989). Pour les échantillons sans analyse complète de terres rares, les concentrations de La et de Ce déterminées par SEAP ont été utilisées.

Les calculs normatifs CIPW et les classifications d'échantillons dans l'espace quartz -plagioclase - fedspath-alcalin s'accordent généralement avec les descriptions de terrain et de lames minces (Fig. 5a). Par contre, le pluton D2 est plus leucocratique d'après la classification normative (tonalite et diorite quartzique à monzodiorite) qu'il ne l'est d'après la description de terrain (diorite). Cependant, il n'est pas possible de clarifier davantage car aucune lame mince n'est disponible. Les compositions calculées de plagioclase varient d'approximativement An_{20} à An_{50} (Fig. 5b), ce qui est typique pour des roches dioritiques à granodioritiques. En raison du faible contenu d'anorthite (An) dans ces roches et particulièrement pour les échantillons plus mafiques, le terme diorite est utilisé au lieu de gabbro. Néanmoins, il est accepté que le caractère albitique du plagioclase pourrait avoir été le résultat de réactions métamorphiques dans un protolithe de composition gabbroïque.

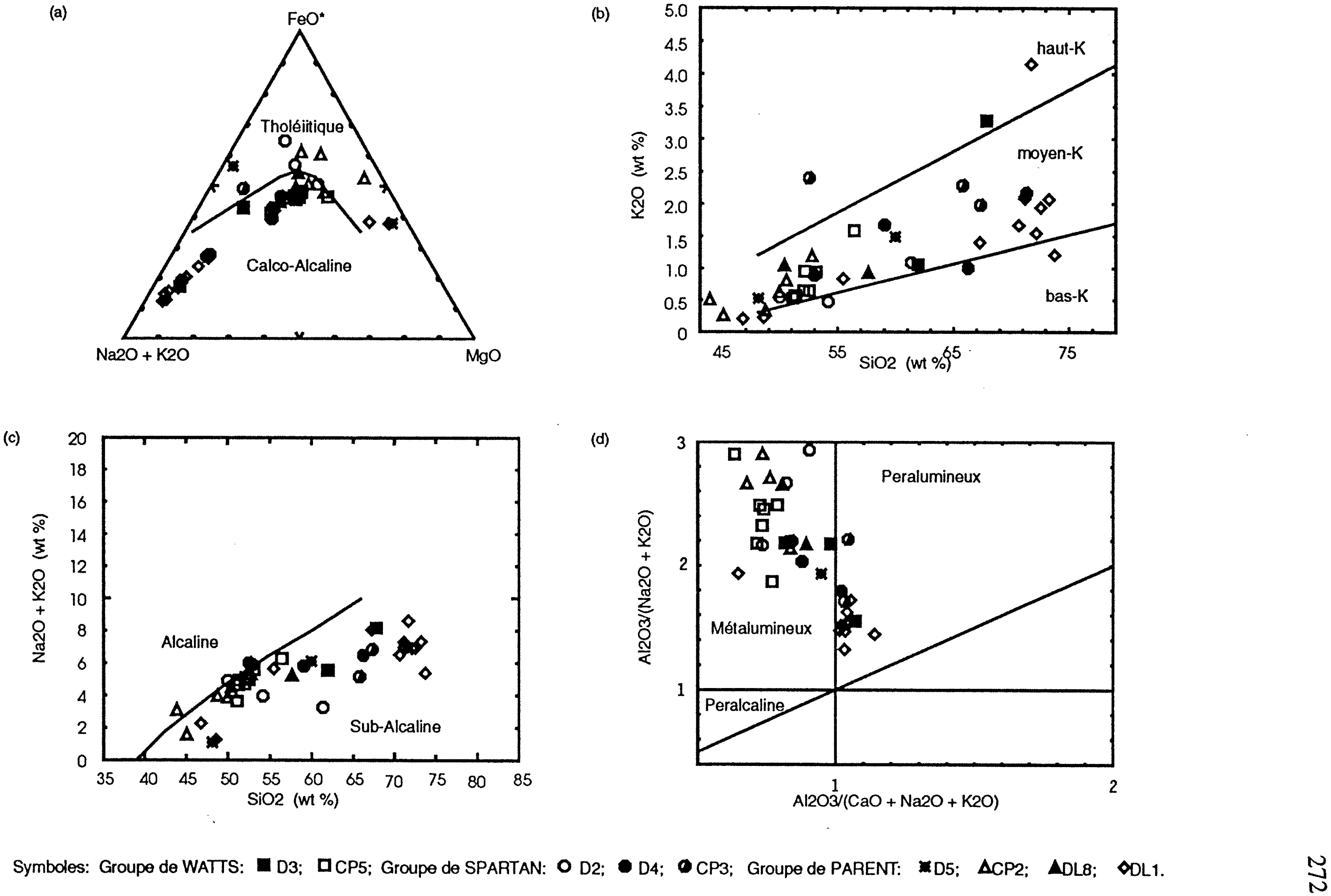


Figure 4.3. Diagramme de distribution d'éléments majeurs des plutons de la Fosse de l'Ungava; (a) diagramme AFM (Irvine et Baragar, 1971); (b) K_2O en fonction de SiO_2 (champs de Le Maitre, 1989); (c) Na_2O + K_2O en fonction de SiO_2 (classification alcaline/sub-alcaline de Irvine et Baragar, 1971); (d) indice de saturation d'alumine (Maniar et Piccoli, 1989).

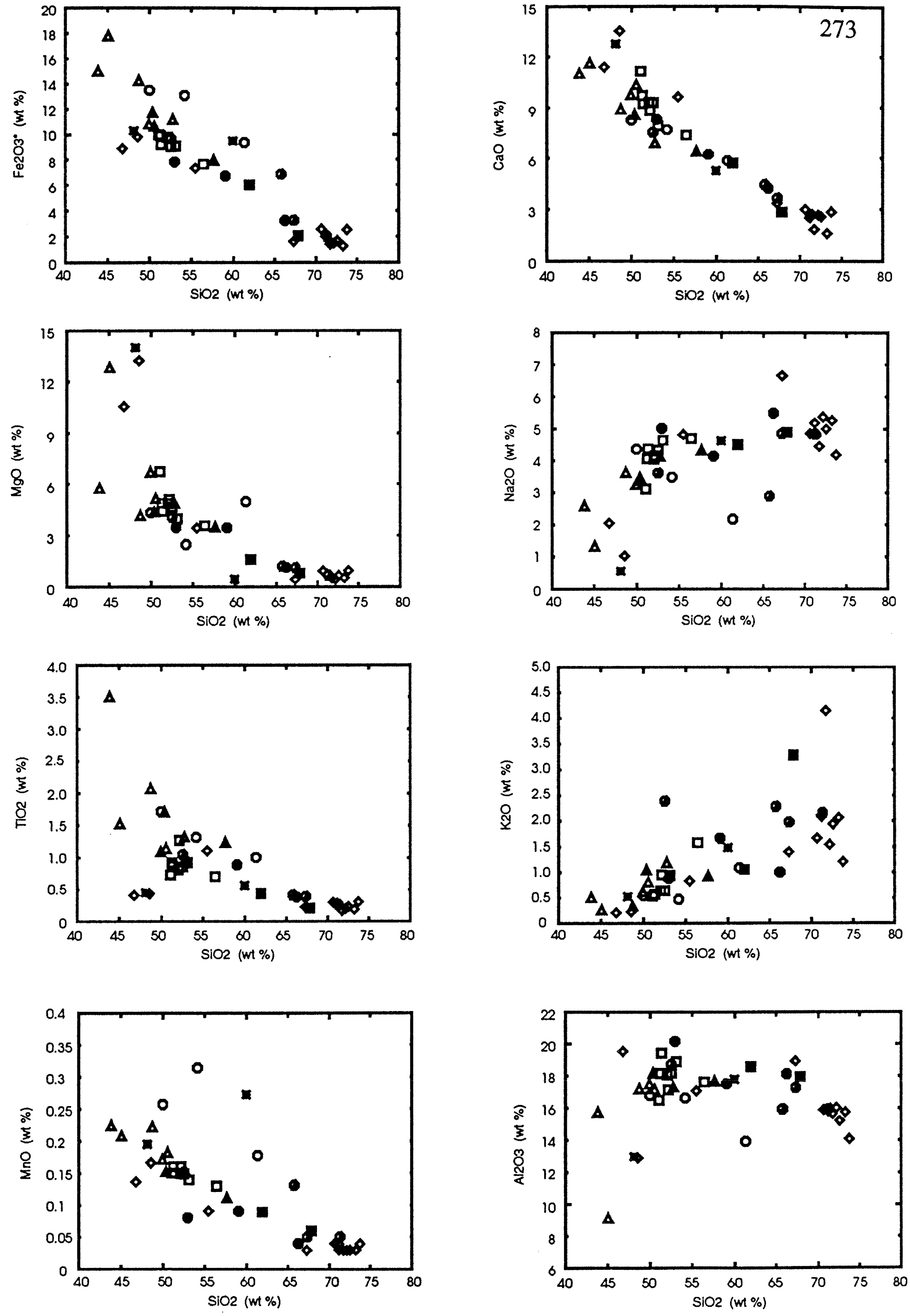
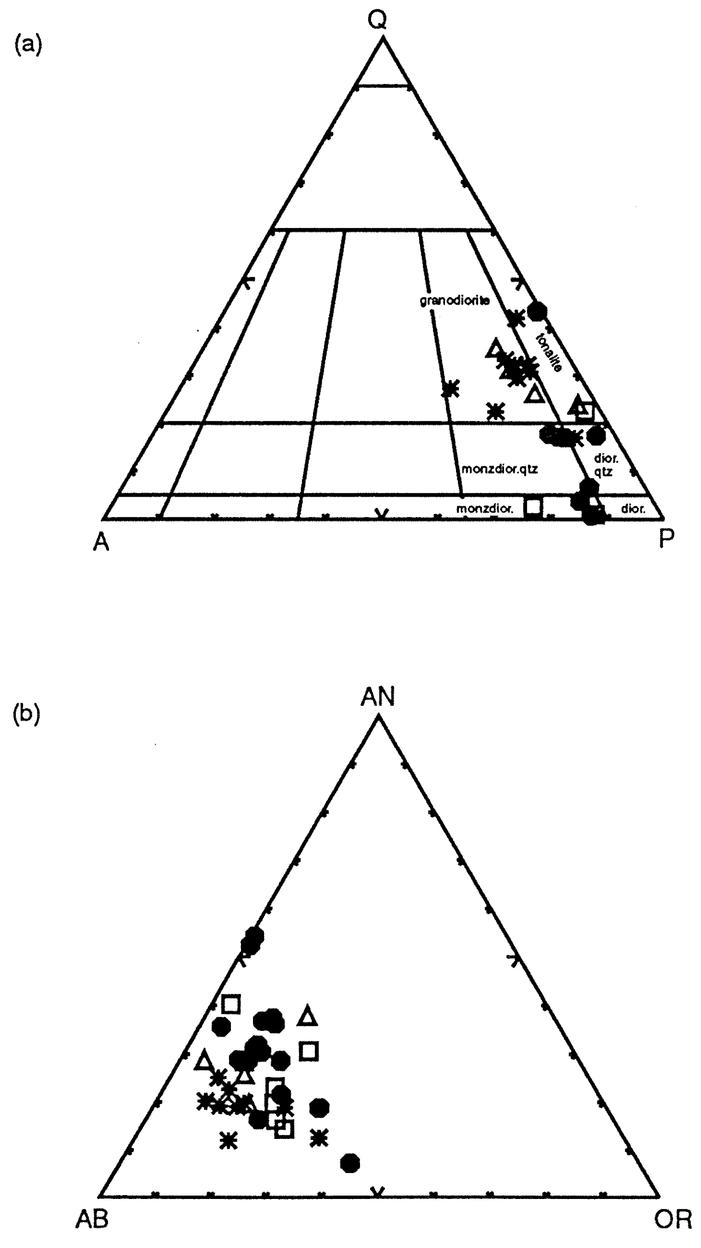


Figure 4.4. Diagrammes de variation d'éléments majeurs (diagrammes de Harker) pour les plutons de la Fosse de l'Ungava. Pour les symboles voir la figure 3.



Symboles: ● diorite, diorite qtz; □ monzdior., monzdior qtz; △ tonalite; ✱ granodiorite.

Figure 4.5. Classification normative CIPW des plutons de la Fosse de l'Ungava; (a) diagramme quartz - feldspath alcalin - plagioclase (QAP) (Streckeisen, 1976); (b) compositions normatives des plagioclases dans l'espace albite - anorthosite - orthoclase.

4.4.3 Plutons du Groupe de Watts

L'analyse géochimique des plutons du Groupe de Watts obtient des compositions de 50,8 à 67,8 % de SiO_2 en poids (anhydre; annexe 1). Des variations régulières d'éléments majeurs avec la silice sont observées, à l'exception d'une analyse du pluton D3 ayant une teneur en K_2O élevée (3,28 % en poids; Fig. 4).

Le pluton **D3** est une monzodiorite quartzique à granodiorite calco-alcaline (Fig. 3a) avec une faible teneur en Ti (0,21 à 0,44 % en poids d'oxyde) mais contenant des concentrations inhabituellement élevées de Cr (1200 ppm) et de Ba (2700 ppm). Les concentrations de K et de Rb sont relativement élevées, variant de 10 à 50X chondrite (Fig. 6a). Aucune analyse d'éléments de terres rares n'est disponible pour D3.

Le pluton **CP5** est aussi calco-alcalin (Fig. 3a) mais les échantillons de monzodiorite et de diorite possèdent tous une teneur en Ti plus élevée (0,7 à 1,27 % en poids d'oxyde) et des teneurs en Cr (53 à 132 ppm) et en Ba (304 à 600 ppm) plus faibles que D3. La teneur en Rb est très faible (5 à 31 ppm; un rapport de Rb/Sr maximum de 0,039) ainsi que la teneur en Th (surtout inférieure à 10X chondrite). Des anomalies positives prononcées de Sr (90 à 105X chondrite) et de Eu (25X chondrite) et de légères anomalies négatives de Zr, de Ti et de Nb sont notées dans les courbes des diagrammes étendus d'éléments trace (Fig. 6b). Les terres rares sont légèrement fractionnées dans CP5 (Fig. 6c), avec des concentrations de terres rares légères jusqu'à 65 à 90X chondrite (La_n/Yb_n = 7,6 à 8,1).

4.4.4 Plutons du Groupe de Spartan

Les plutons du Groupe de Spartan varient de composition dioritique à tonalitique (SiO_2 = 50,0 à 71,3 % en poids anhydre) et ont en général des tendances d'éléments majeurs régulières. Quelques échantillons altérés de D2 et CP3 font l'exception avec des teneurs plus faibles que la normale en Na_2O et Al_2O3 et plus élevées en K_2O, MgO et Fe_2O_3 (Fig. 4).

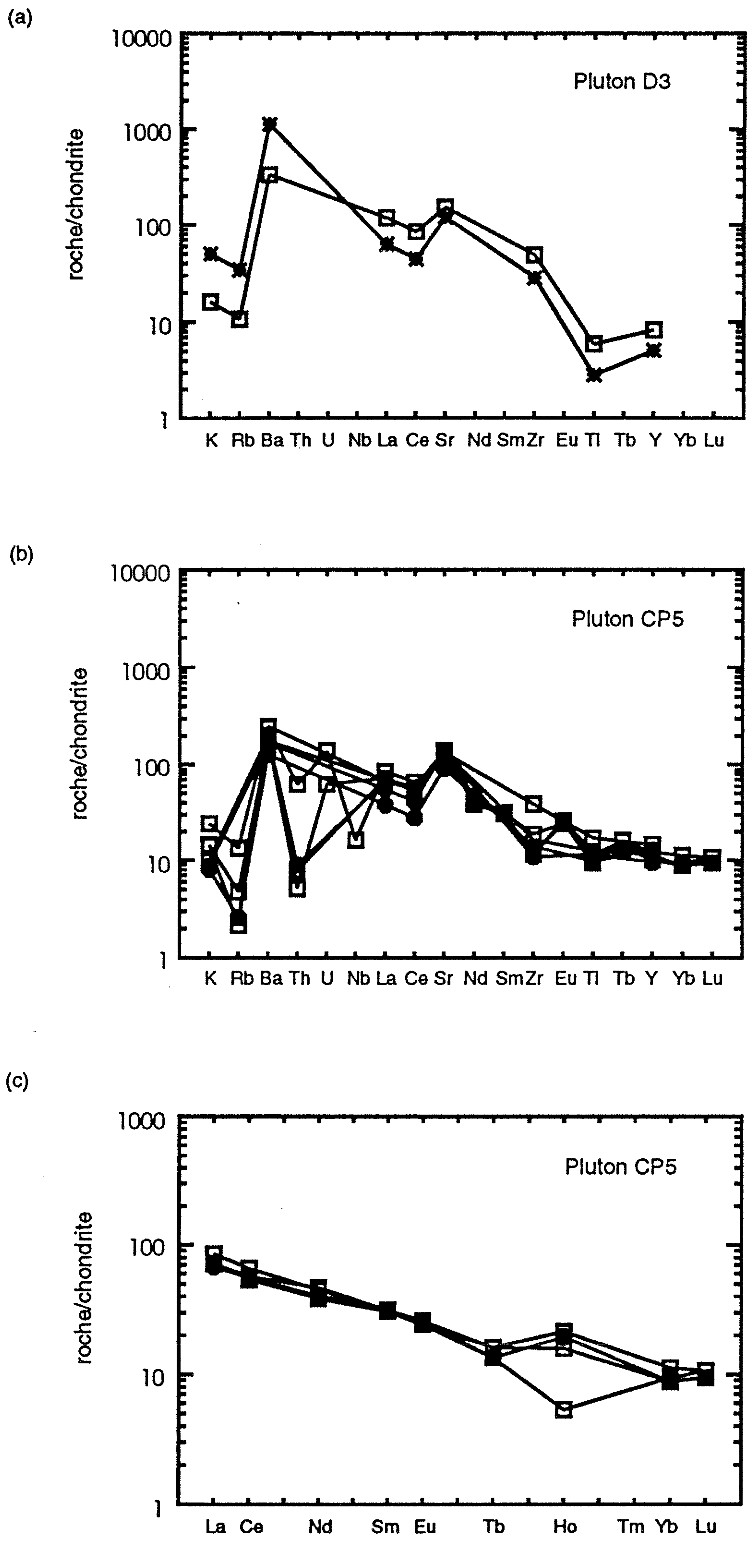


Figure 4.6. Diagrammes étendus d'éléments trace et de terres rares normalisés aux chondrites pour les plutons du Groupe de Watts (facteurs de normalisation d'après Sun et McDonough, 1989) (pour les symboles voir la figure 5); (a) pluton D3; (b) et (c) pluton CP5.

Le spectre de terres rares du pluton dioritique tholéiitique **D2** est relativement atténué, avec des rapports de La_n/Yb_n entre 2,5 et 3,2 (Fig. 7a). Les concentrations d'éléments hygromagmaphiles sont tous inférieures à 110X chondrite (Fig. 7b) et celles de K et de Rb généralement inférieures à 10X chondrite (K/Rb = 360 à 480).

Des concentrations semblables d'éléments hygromagmaphiles sont obtenues pour **D4** (Fig. 7c), un pluton calco-alcalin de diorite quartzique à tonalite, lui aussi intrusif dans le Groupe de Spartan. Cependant, une anomalie positive de Sr est observée dans D4 (approximativement 100X chondrite, en comparaison avec 50 à 60X chondrite pour D2) ainsi qu'une anomalie négative prononcée de Nb (La_n/Nb_n = 4,15). L'enrichissement en terres rares légères résulte en une courbe de terres rares fractionnées et des rapports de La_n/Yb_n atteignant 14,4 (Fig, 7d) donc, des valeurs beaucoup plus élevées que dans D2. Dans l'ensemble, les concentrations de terres rares sont moins élevées dans l'échantillon de tonalite que dans les échantillons de diorite.

Le troisième pluton trouvé dans le Groupe de Spartan est **CP3**, une intrusion principalement tonalitique, quoiqu'une analyse géochimique réflète plutôt une composition monzodioritique. Les échantillons de tonalite sont tous calco-alcalins tandis que la monzodiorite est tholéiitique en ce qui concerne la géochimie d'éléments majeurs (Fig. 3a). Seuls deux spectres de terres rares pour CP3 sont disponibles (une de la tonalite et une de la monzodiorite) et présentent des courbes fractionnées (Fig. 7e). La tonalite a un rapport de La_n/Yb_n de 32,3 tandis que la monzodiorite est moins fractionnée avec un rapport de La_n/Yb_n de 5,6. La concentration exceptionnellement faible de La dans l'échantillon monzodioritique suggère des problèmes analytiques qui ont résulté en un rapport de La_n/Yb_n faible. Les concentrations de K et de Rb sont plus élevées dans CP3 que dans les plutons mentionnés ci-haut, variant de 20 à 35X chondrite (Fig. 7f). La concentration de Ba est aussi relativement élevée (399 à 851 ppm) tandis que celles de Th et de U le sont moins. Les concentrations de TiO_2 pour la tonalite sont inférieures à 0,42 % en poids tandis que la monzodiorite contient un minimum de 1 % en poids de TiO_2, ce qui est comparable aux autres échantillons de diorite/monzodiorite.

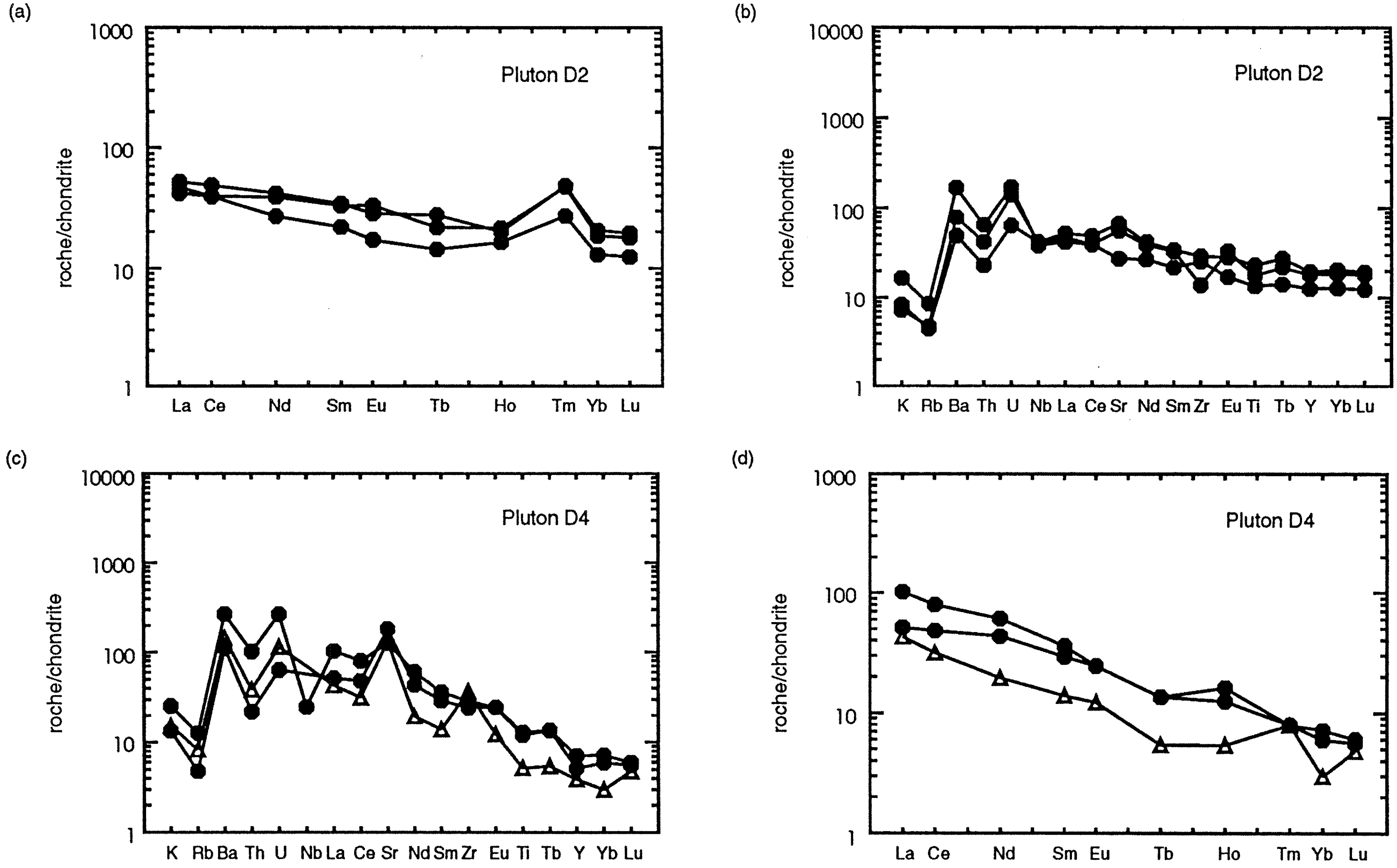


Figure 4.7. Diagrammes étendus d'éléments trace et de terres rares normalisés aux chondrites pour les plutons du Groupe de Spartan (pour les symboles voir la figure 5); (a) et (b) pluton D2; (c) et (d) pluton D4; (e) et (f) pluton CP3.

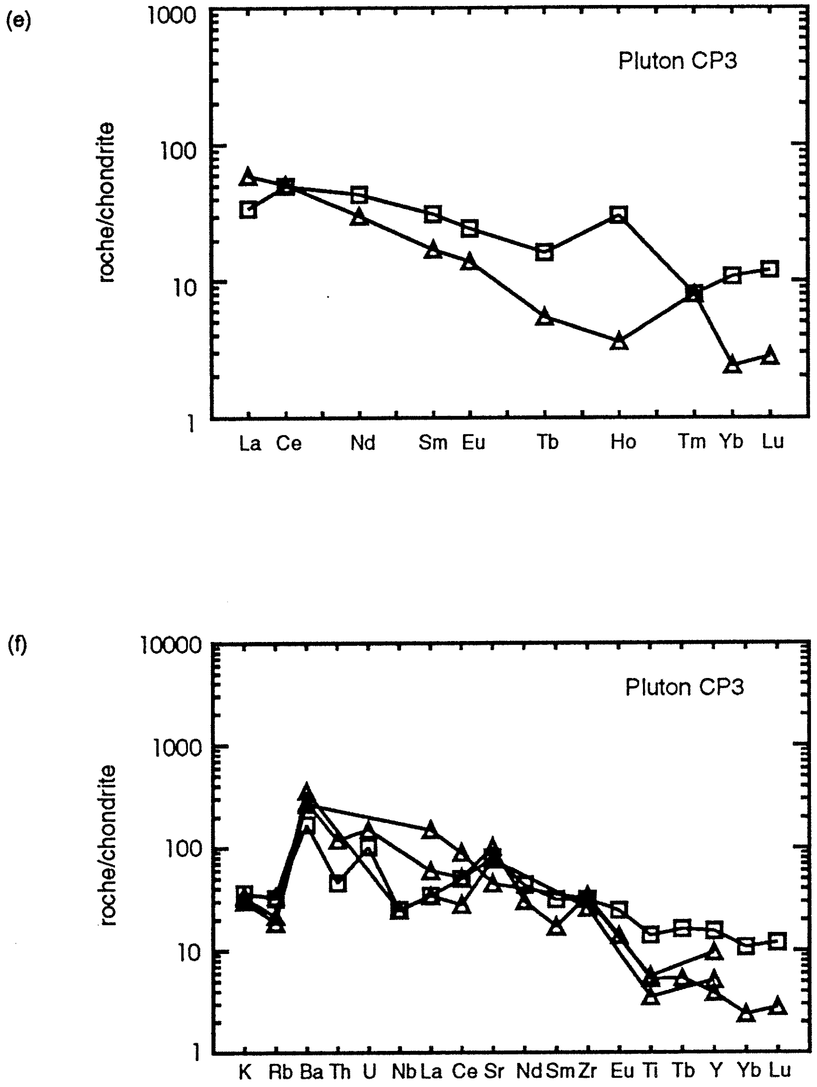


Figure 4.7 (suite).

4.4.5 Plutons du Groupe de Parent

Des compositions trés variables de plutons sont trouvées à l'intérieur du Groupe de Parent avec des concentrations de SiO_2 variant de 43,9 à 73,8 % en poids anhydre. Les effets de l'altération dans quelques échantillons résultent en concentrations atypiques de certains éléments tels le Na_2O, le Al_2O, le K_2O, le MgO et le Fe_2O_3 (Fig. 4). Un échantillon provenant d'un pluton du Groupe de Parent (CP2) a été rejeté de cette discussion dû à son contenu de volatiles excessivement élevé (PAF>8%).

Le pluton dioritique **CP2** possède des caractéristiques géochimiques calco-alcalines et tholéiitiques (Fig. 3a). La signature géochimique du pluton est très variable (Fig. 8a), ses concentrations de K et de Rb variant de 2 à 20 X chondrite, des rapports de Rb/Sr de 0,012 à 0,181, et des concentrations de Ba de 100 à 334 ppm. De plus, la concentration de 317 ppm de Ni dans un des échantillons de cette intrusion est élevée. Seule une analyse de terres rares est disponible et démontre un léger enrichissement en terres rares légères en relation aux terres rares lourdes et un rapport de La_n/Yb_n de 5,0 (Fig. 8b). En général, la composition géochimique de CP2 est similaire à celles provenant d'échantillons de diorite-gabbro mentionnés antérieurement.

Des échantillons de tonalite, de diorite quartzique et de diorite ont été prélevés du pluton **D5** mais seulement deux analyses géochimiques sont disponibles et révèlent une composition mafique. Les échantillons sont tholéiitiques et présentent des spectres d'éléments trace similaires. Des courbes très peu fractionnées de terres rares légères et des courbes atténuées de terres rares lourdes sont typiques (Fig. 8c); les rapports de La_n/Yb_n sont de 3,5 pour la diorite et 3,9 pour la diorite quartzique. Les deux échantillons possèdent une légère anomalie positive de Sr et des anomalies négatives plus importantes de Nb et de Ti (Fig. 8d); un enrichissement en Eu est remarqué dans la diorite quartzique. Des concentrations élevées de Ni de 204 ppm sont trouvées dans la diorite. Généralement, les spectres d'éléments trace pour ce pluton ressemblent de près à ceux de D2, une diorite dans le Groupe de Spartan.

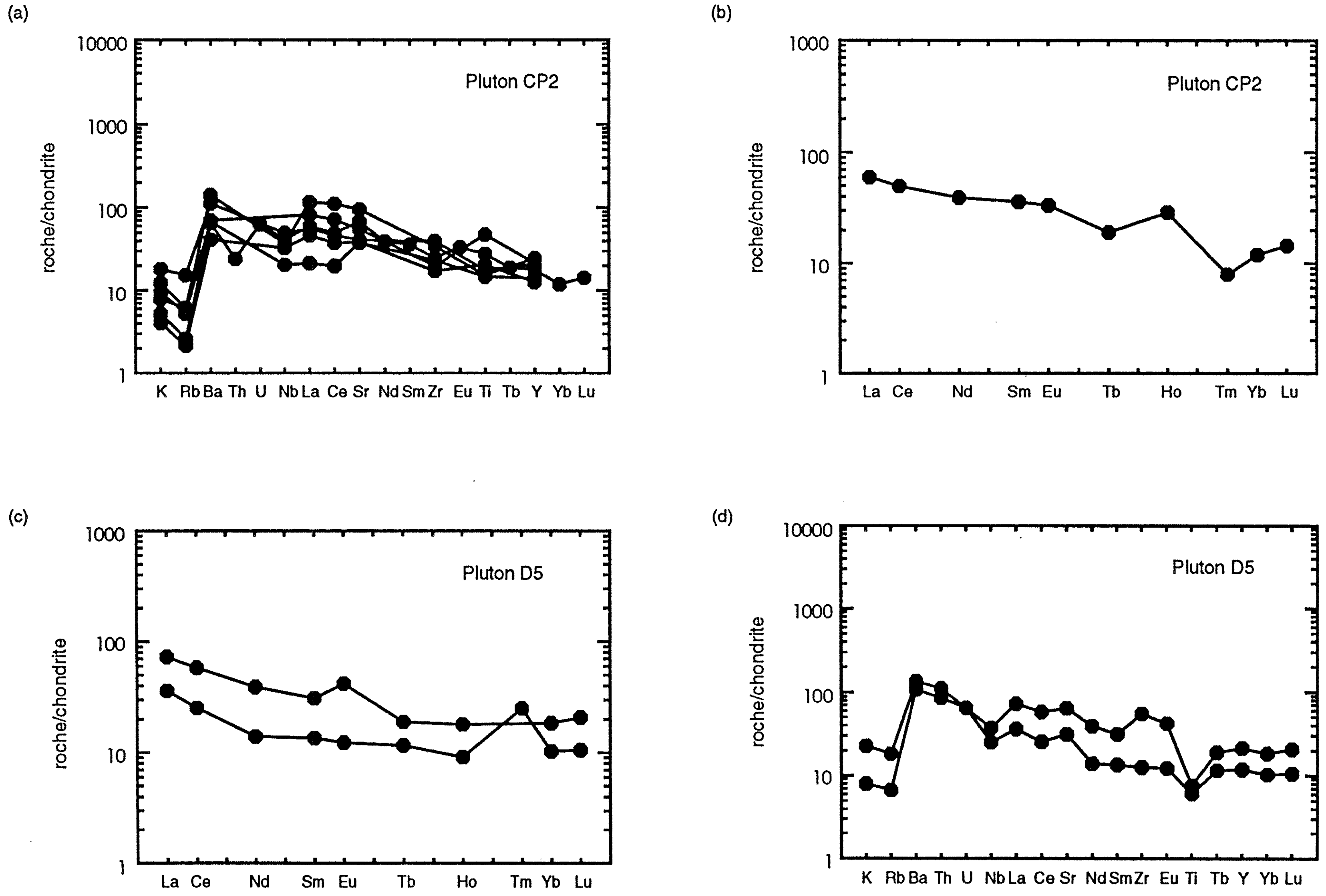


Figure 4.8. Diagrammes étendus d'éléments trace et de terres rares normalisés aux chondrites pour les plutons du Groupe de Parent. Pour les symboles voir la figure 5; (a) et (b) pluton CP2; (c) et (d) pluton D5; (e) et (f) pluton DL8; et (g) pluton DL1.

Une troisième intrusion, **DL8**, est représentée par deux échantillons dioritiques, tous deux calco-alcalins. Ce pluton est caractérisé par des spectres de terres rares parallèles et le fractionnement des terres rares légères (La_n/Yb_n = 7,2 à 8,2; Fig. 8e). La courbe d'éléments trace pour DL8 chevauche celle d'autres échantillons de diorite analysés, avec des concentrations typiquement faibles de K et de Rb et plus enrichies en Ba, Th et U (Fig. 8f). Les concentrations de Cu pour DL8 sont considérablement plus élevées (496 ppm) que la valeur maximum (200 ppm) obtenue dans la plupart des autres analyses.

Le pluton principalement felsique **DL1** est entièrement calco-alcalin et peralumineux, sauf pour deux des analyses de diorites (qui provient de la intrusiondioritique coupé par DL1) qui sont tholéiitiques (Fig. 3a). Sur un diagramme étendu d'éléments trace, les courbes des diorites sont très atténuées (Fig. 8g), ces roches possédant généralement une faible abondance d'éléments hygromagmaphiles et des concentrations élevées de Ti (0,41 à 1,11 % en poids d'oxyde) et de Ni (jusqu'à 220 ppm). En comparaison, les granodiorites possèdent des valeurs de K et Rb élevées (10 à 60X chondrite) et de Ba variant de 494 à 1300 ppm. Ces granodiorites possèdent des teneurs très faibles en Cu et Ni, inférieurs à 5 ppm et 9 ppm respectivement. Les valeurs de TiO_2 sont aussi plus basses que dans les diorites, moyennant 0,24 % en poids.

4.5 RELATIONS ENTRE LE PLUTONISME DES GROUPES DE WATTS, DE SPARTAN ET DE PARENT

La composition, la chimie et l'état de déformation des plutons des groupes de Watts, de Spartan et de Parent sont diverses. Ces plutons varient en composition du gabbro à la tonalite ou granodiorite et de texture massive à bien foliée. L'âge de ces plutons a été déterminé entre env. 1880 et 1840 Ma dans le Groupe de Watts et entre 1898 et 1845 Ma dans le Groupe de Parent.

Tous les plutons ont été affectés par un métamorphisme régressif qui a transformé l'hornblende et le pyroxène primaires, par degrés variables, en minéraux secondaires tels

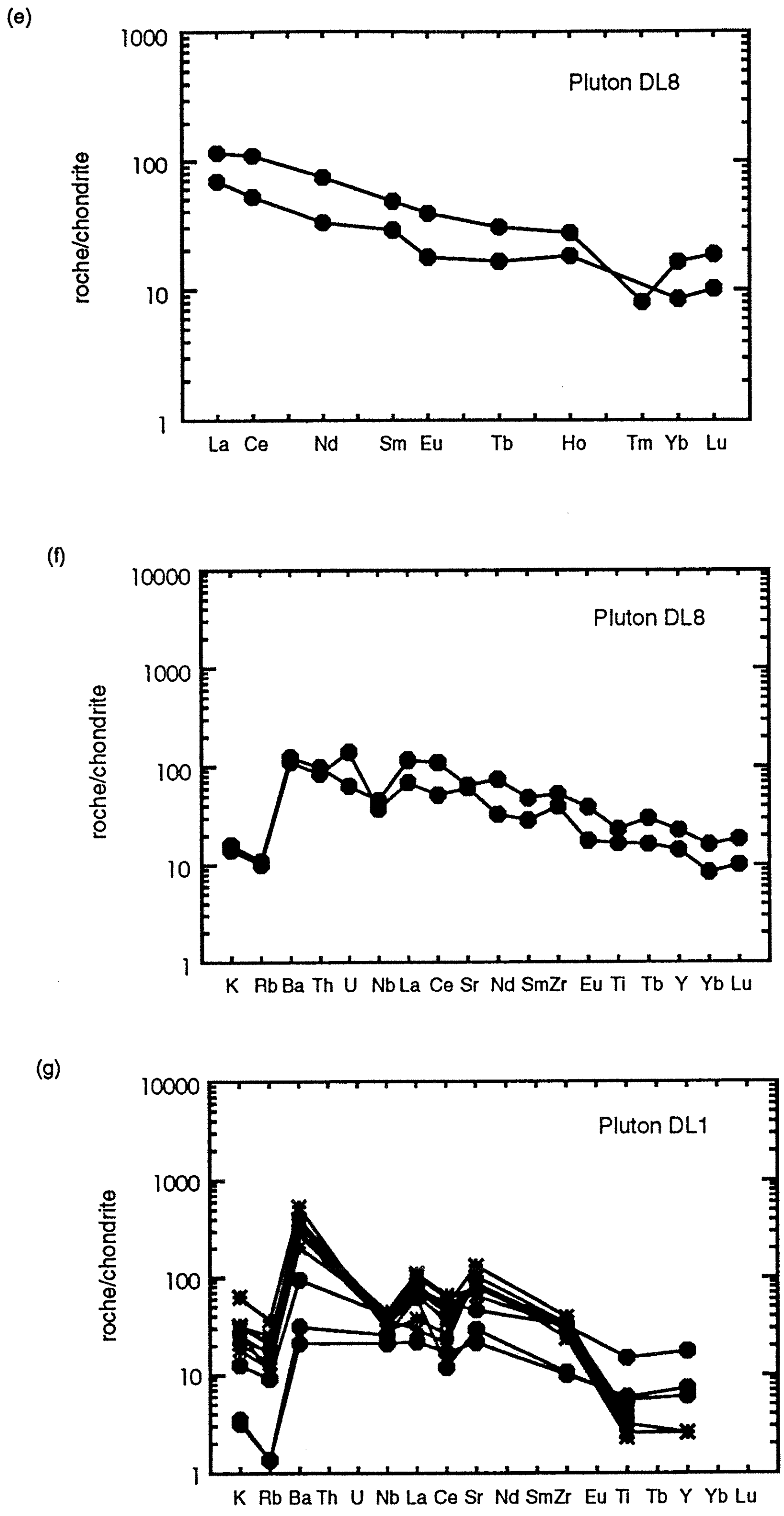


Figure 4.8 (sutie).

que la chlorite, l'épidote et l'actinote, et qui a produit des quantités appréciables de séricite et de clinozoisite sur les cristaux de plagioclase.

Les signatures géochimiques des échantillons tonalitiques et granodioritiques (provenant de plutons CP3 et DL1) sont généralement très similaires, ce qui est particulièrement évident dans les diagrammes de variation multi-éléments trace (Fig. 9a). Tous les échantillons possèdent de fortes concentrations d'éléments hygromagmatophiles mais très peu de Ti et de terres rares lourdes. Cependant, les échantillons plus mafiques, surtout les échantillons de monzodiorite et de monzodiorite quartzique, démontrent une plus grande variabilité de concentrations d'éléments majeurs et trace (Fig. 9b). En général, les courbes des échantillons felsiques chevauchent celles des échantillons mafiques les plus enrichis sur les diagrammes étendus d'éléments trace. Ceci est le cas pour presque tous les éléments sauf le Ti et les terres rares lourdes, dont les teneurs sont plus faibles dans les tonalites et les granodiorites. Le grenat aurait donc pu jouer un rôle important dans la pétrogenèse des échantillons felsiques mais toutefois moins important dans l'évolution des échantillons dioritiques. Tous les échantillons analysés ont une anomalie négative de Nb ($(La/Nb)_n>1$) et tombent dans le champ des arcs volcaniques (Fig. 10) sur les diagrammes de discrimination d'éléments trace de Pearce *et al.* (1984).

Les études isotopiques Sm/Nd et Rb/Sr de plusieurs plutons dans la Fosse de l'Ungava, par Dunphy (en préparation), démontrent que tous les plutons analysés sont juvéniles et possèdent des signatures isotopiques similaires, quel que soit leur âge ou leur composition. Les données isotopiques de Nd pour les plutons correspondent à celles qui sont disponibles pour les groupes de Watts, de Chukotat et de Povungnituk (Hegner et Bevier, 1989, 1991) ainsi que celles des magmas les plus appauvris de l'arc de Narsajuaq (Dunphy, en préparation).

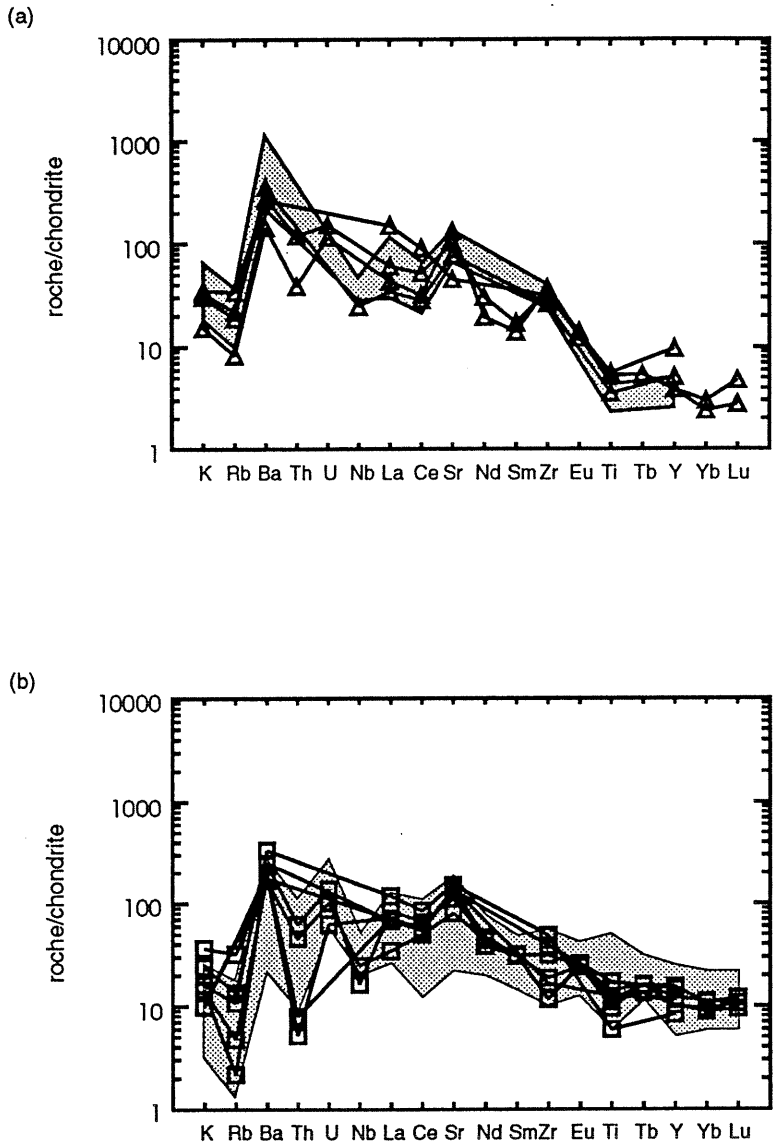
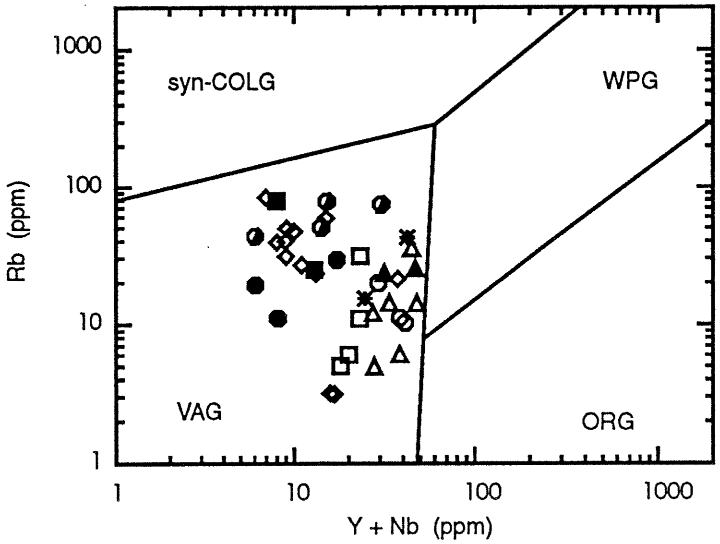


Figure 4.9. Comparaison de la signature géochimique (a) des échantillons de granodiorite (champ hachuré) et de tonalite (triangles ouverts); (b) échantillons de diorite (champ hachuré) et de monzodiorite (carrés ouverts).



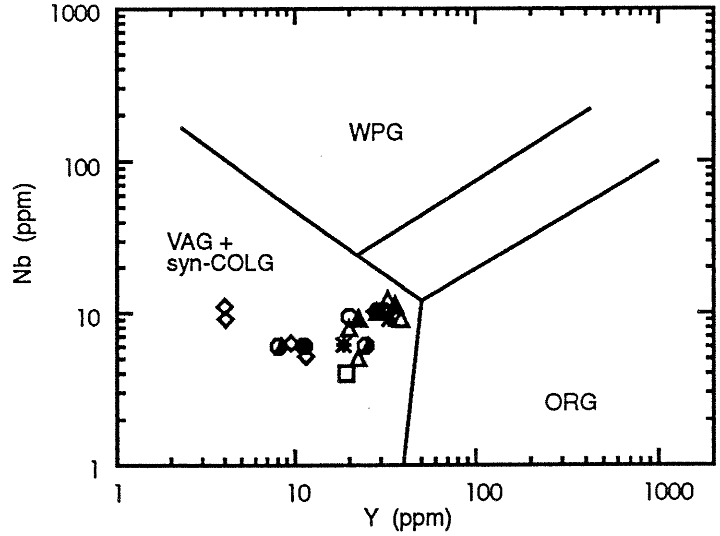


Figure 4.10. Diagrammes de discrimination d'environnement tectonique utilisant les éléments trace (Pearce *et al.*, 1984) pour tous les échantillons (pour les symboles voir la figure 3.). Abréviations: WPG = granites intracontinentaux; ORG = granites de dorsale médio-océanique; COLG = granites de milieu de collision; VAG = granites d'arc volcanique.

4.6 RELATIONS ENTRE LES PLUTONS DE LA FOSSE DE L'UNGAVA ET LES PLUTONS DE L'ARC DE NARSAJUAQ

4.6.1 Plutonisme de l'arc de Narsajuaq

Afin de comparer adéquatement les intrusions de la Fosse de l'Ungava aux plutons trouvés plus au nord, dans le terrain de l'arc de Narsajuaq, une brève introduction et description du magmatisme Narsajuaq est présentée ici. L'arc de Narsajuaq est une série complexe de plutons au faciès des granulites et de quantités mineures de roches supracrustales accrétés au socle de la Province du Supérieur (St-Onge et Lucas, 1990a, 1992; Lucas et St-Onge, 1991; St-Onge *et al.*, 1992). La lithologie principale de l'arc de Narsajuaq est une séquence bien stratifiée (à l'échelle centimétrique à décimétrique) de tonalite et de diorite quartzique, interlitée et pénétrée par des filons granitiques variablement déformés et métamorphisés. L'unité contient aussi des blocs et des enclaves de roches mafiques et ultramafiques, plus anciennes à comagmatiques, et de sédiments plus anciens (Groupe de Sugluk). On y retrouve aussi localement des séquences concordantes de diorite quartzique, d'amphibolite, de pyroxénite et de péridotite métamorphisées. Cette unité a été datée entre 1863 et 1861 Ma par Parrish (St-Onge *et al.*, 1992). Intrusive dans cette unité de base est une suite plus jeune de plutons kilométriques de diorite quartzique, de tonalite et de granite métamorphisées et variant en âge de 1844 à 1826 Ma (St-Onge *et al.*, 1992). Localement, dans cette suite plus jeune, on retrouve des enclaves de lithologies appartenant à la suite plus ancienne ainsi que de la pyroxénite, de la diorite, de la tonalite comagmatiques et des inclusions métasédimentaires. En général, les plutons de la suite plus ancienne sont beaucoup plus déformés que ceux de la suite plus jeune. Des granites d'environnement de collision encore plus jeunes (env. 1803 Ma) envahissent les deux suites magmatiques sous forme de plutons tabulaires kilométriques (Dunphy et Ludden, 1992; St-Onge et Lucas, 1992).

Géochimiquement, l'affinité calco-alcaline est prédominante dans les roches du terrane Narsajuaq; quelques plutons plus mafiques sont tholéiitiques (Dunphy et Ludden, 1992). En général, les concentrations de Fe_2O_3, de MgO, de CaO et de TiO_2 décroissent avec l'augmentation de SiO_2. Ces roches plutoniques sont caractérisées par de fortes concentrations d'éléments hygromagmaphiles tels le Ba, le Th et le U (jusqu'à 2800, 17 et 5,2 ppm respectivement), des anomalies négatives prononcées de Nb et de Ti et des terres rares variablement fractionnées ($(La/Yb)_n$ = 2,8 à 73,3) (Dunphy et Ludden, 1992). Les valeurs initiales de ε_{Nd} et $^{87}Sr/^{86}Sr$ pour les magmas de Narsajuaq varient d'appauvries (+4,0 et 0,7020) à très enrichies (-18,4 et 0,7062) (Dunphy *et al.*, 1993). Les valeurs juvéniles correspondent à celles rapportées pour les roches volcaniques de la Fosse de l'Ungava adjacente (Hegner et Bevier, 1989, 1991).

La chimie des éléments majeurs et trace des roches du terrane de Narsajuaq est similaire à celles d'autres granites d'arcs volcaniques rapportées antérieurement (Dunphy et Ludden, 1992). De plus, les roches granitiques du complexe plutonique Narsajuaq ont des compositions et des volumes comparables à ceux des granites exposés dans les batholites massifs Mésozoïques situés le long de la marge occidentale des Amériques du Nord et du Sud, de l'Alaska à l'Antarctique (Brown, 1982; Mahlburg Kay et Rapela, 1990). La contamination de magmas appauvris par des composantes crustales et associées à la subduction est signalée par les signatures isotopiques des plutons de l'arc Narsajuaq (Dunphy *et al.*, 1993). Les roches de l'arc de Narsajuaq ont été interprétées comme le noyau plutonique d'un arc magmatique (Dunphy et Ludden, 1992; Lucas *et al.*, 1992; St-Onge *et al.* 1992).

4.6.2 Comparaison des plutons de la Fosse de l'Ungava avec les plutons de l'arc de Narsajuaq

Les variations de composition des plutons de la Fosse de l'Ungava sont similaires à celles des roches de l'arc de Narsajuaq puisque toutes deux varient entre calco-alcalines et tholéiitiques. Cependant, les plutons mafiques sont plus communs à l'intérieur de la ceinture mobile de la bande du Cap Smith tandis que la composition tonalitique prédomine dans l'arc Narsajuaq. En général, les plutons de la Fosse de l'Ungava et leur équivalents en composition globale dans l'arc de Narsajuaq ont des signatures géochimiques semblables. Par exemple, les diorites provenant des plutons D2 et D4 ont des courbes comparables au diorites de l'arc de Narsajuaq sur les diagrammes étendus d'éléments trace (Fig. 11a). Quelques différences sont notées, tel qu'un faible contenu de terres rares lourdes de D4 relativement aux diorites Narsajuaq. Un seul échantillon de diorite provenant de DL1 du Groupe de Parent suit une tendance semblable à celle des diorites de l'arc de Narsajuaq; les deux autres diorites DL1 ont des concentrations généralement plus faibles d'éléments trace (non représentées sur les figures). Les spectres de terres rares de D5 et CP2 sont comparables à ceux des diorites Narsajuaq. Cependant, on note une concentration de Ba plus faible dans D5 et CP2 ainsi qu'une concentration de Ti plus faible et de Th plus élevée dans D5 (non représentées sur les figures). En ce qui concerne les roches de composition plus felsique, les tonalites de CP3 et de l'arc de Narsajuaq possèdent des contenus d'éléments hygromagmaphiles semblables mais les contenus de terres rares (particulièrement les terres rares lourdes) et de Ti sont relativement bas dans CP3 (Fig. 11b). Sauf pour des contenus de Nb et Ti plus élevés, la granodiorite DL1, située dans le Groupe de Parent, a un contenu d'éléments trace très similaire à celui des filons granitiques qui sont communs dans l'arc de Narsajuaq (Fig. 11c).

La similarité générale de la composition et de la signature géochimique des plutons du Cap Smith et ceux de l'arc de Narsajuaq suggère que ces intrusions ont subi une pétrogenèse semblable. De plus, le plutonisme dans l'arc de Narsajuaq est synchrone au plutonisme de la Fosse de l'Ungava, ce qui a mené St-Onge *et al.* (1992) à proposer une corrélation entre les deux événements. En ce qui concerne les compositions isotopiques de Nd et Sr, les magmas du terrane de Narsajuaq sont très divers tandis que les plutons du Cap Smith ont des compositions relativement restreintes (Dunphy, en préparation). Les données isotopiques pour les plutons du Cap Smith indiquent qu'ils proviennent d'une source mantellique appauvrie et légèrement contaminée. Les données disponibles indiquent que les plutons du Cap Smith ont été générés dans un environnement de subduction à peu près en même temps que les magmas Narsajuaq.

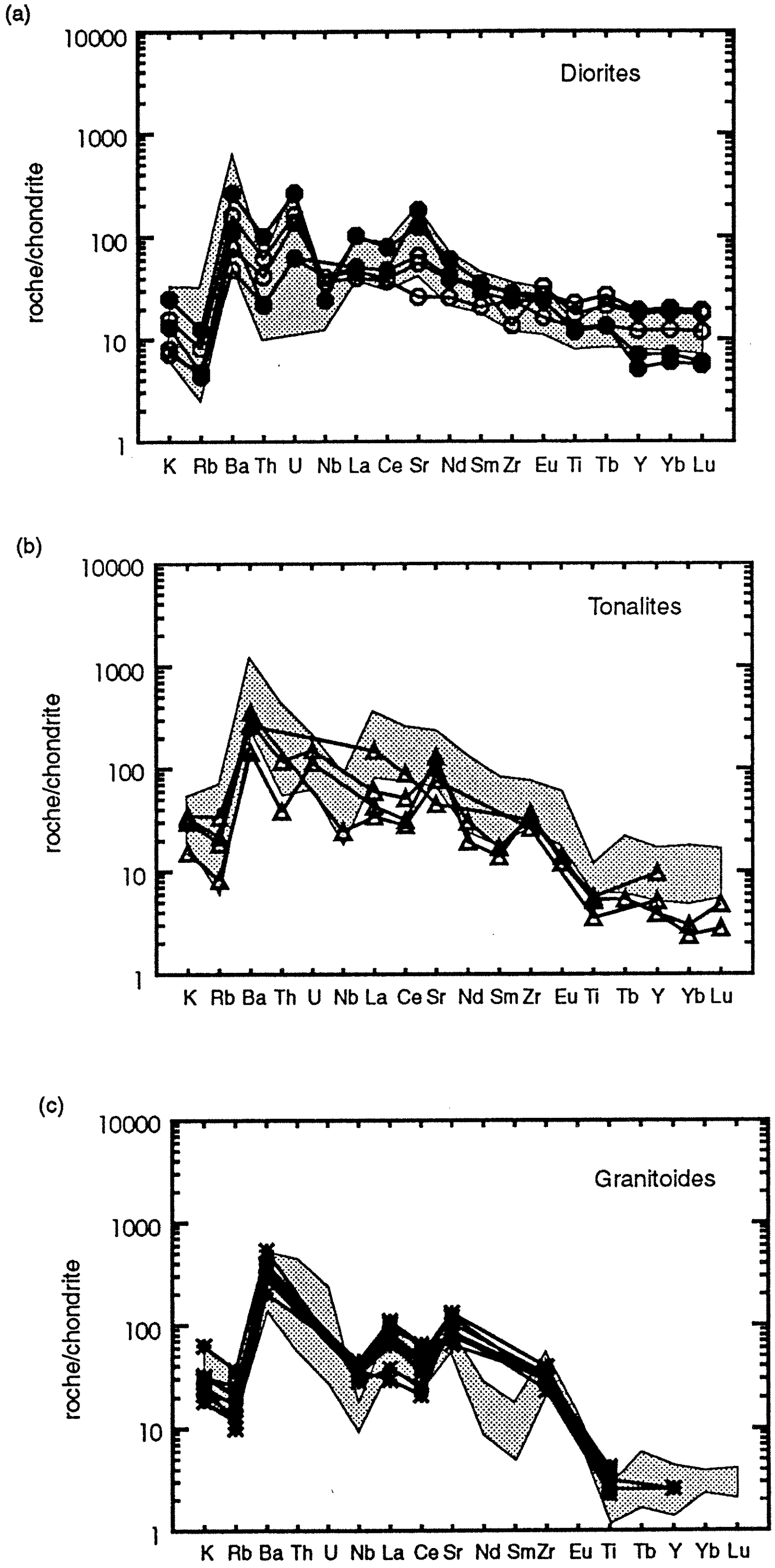


Figure 4.11. Diagrammes étendus d'éléments trace et de terres rares normalisés aux chondrites comparant les lithologies du terrain de Narsajuaq (champ hachuré; données non-publiées de Dunphy) aux plutons de composition semblable de la bande du Cap Smith: (a) diorites du terrain de Narsajuaq et des plutons D2 et D4; (b) tonalites du terrain Narsajuaq et du pluton CP3; (c) échantillons de granodiorite de DL1 et du terrane de Narsajuaq.

4.7 RELATION ENTRE LE PLUTONISME ET LE VOLCANISME DANS LE GROUPE DE PARENT

4.7.1 Description du Groupe de Parent

Le Groupe de Parent (cartographié originalement comme une partie de la Formation Parent du Groupe de Watts (Tremblay, 1989) est une séquence de roches volcaniques et sédimentaires structuralement intercalées au-dessus du Groupe de Chukotat et sous le Groupe de Watts. Il est exposé dans les régions occidentales et centrales de la Fosse de l'Ungava et consiste en des tufs volcaniclastiques et des laves basaltiques à rhyolitiques (le basalte andésitique étant le plus commun) intercalés avec des shales, siltstones et grauwackes (Tremblay, 1989, 1991; Picard *et al.*, 1990; St-Onge et Lucas, 1992; Picard, en préparation). Une description pétrologique complète des types de roches du Groupe de Parent est présentée dans Picard (en préparation). La signature géochimique des roches volcaniques/volcaniclastiques du Groupe de Parent a mené Picard *et al.* (1990) à suggérer que ces roches ont été mises en place dans un environnement d'arc magmatique. Les roches magmatiques du Groupe de Parent possèdent des affinités calco-alcalines et tholéiitiques et sont sub-alcalins et métalumineux. Les roches volcaniques ont des spectres de terres rares légèrement fractionnés (La_n/Yb_n = 2,38 à 5,53) et des anomalies de Nb négatives (Picard, en préparation). Sur les diagrammes de classification géotectonique de Pearce *et al.* (1984), les roches volcaniques du Groupe de Parent sont trouvées dans le champ des arcs volcaniques. Un âge de 1860 Ma a été obtenu par Machado *et al.* (1993; échantillon DL6) pour une rhyolite du Groupe de Parent. De plus, une rhyodacite (échantillon DL3) a aussi été datée, donnant un âge maximum de 1917 Ma avec des zircons hérités à 1968 et 2423 Ma. Étant donné les zircons hérités et les contraintes pétrographiques, Machado *et al.* (1993) suggèrent que cet échantillon est un métasédiment terrigène et ne restreint donc pas l'âge du volcanisme dans le Groupe de Parent.

4.7.2 Comparaisons des roches volcaniques et plutoniques du Groupe de Parent

En général, les plutons mafiques du Groupe de Parent et les roches volcaniques encaissantes possèdent une signature géochimique similaire. La similarité des courbes de ces roches sur les diagrammes étendus d'éléments trace démontre qu'elles sont équivalentes (Fig. 12). Par contre, la géochimie du pluton granodioritique DL1 se corrèle difficilement avec celle des roches volcaniques mais ceci pourrait être dû au fait qu'aucune analyse d'éléments trace de roche de composition similaire (*e.g.* andésite, rhyolite) ne soit disponible. Les roches volcaniques et plutoniques du Groupe de Parent possèdent une proche association géographique et temporelle ainsi que des compositions plus ou moins similaires. Ceci indiquerait que ces roches sont des expressions de la même activité magmatique. Cependant, plus de données géologiques et géochimiques détaillées sont requises afin de comprendre à fond la relation exacte des activités extrusives et intrusives. Les études géochimiques et isotopiques d'une suite d'échantillons du Groupe de Parent par J. Dunphy (en cours) aideront à clarifier ces questions.

4.8 CHRONOLOGIE DU PLUTONISME ET DE LA DÉFORMATION

4.8.1 Introduction

À partir des études de terrain, des études de lames minces et des analyses géochronologiques mentionnées ci-haut, une chronologie relative et absolue d'intrusion et de déformation peut être établie (résumée dans le Table 1). Hervet (1985) distingue trois phases d'intrusion plutonique, en se basant principalement sur l'état de déformation des plutons: les phases Povungnituk et Chukotat seraient pré- à syn-tectoniques tandis que la dernière phase Narsajuaq serait tardivement syn-tectonique à post-tectonique. Lamothe et Picard (en rédaction) ont interprété la plupart des plutons de la phase Povungnituk de Tremblay (1989) comme étant membres de la séquence mafique / ultramafique de gabbros stratifiés et de cumulats du Groupe de Watts. L'auteure de ce rapport se réfère aux plutons pré-, syn- et post-tectoniques plutôt qu'aux phases de plutonisme Povungnituk, Chukotat et Narsajuaq. Ceci évite l'implication d'équivalences génétiques qui n'ont pas nécessairement été établies mais qui seraient indiquées par l'usage des mêmes noms pour les phases plutoniques ainsi que les séquences volcaniques.

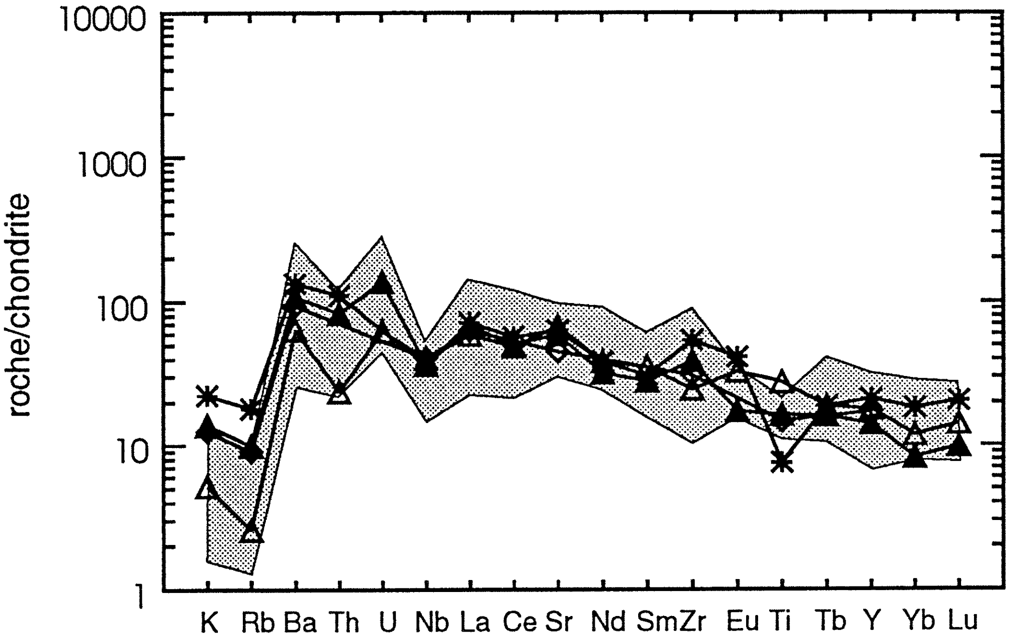


Figure 4.12. Comparaison géochimique des roches volcaniques du Groupe de Parent (champ hachuré; Picard, en préparation) et d'échantillons représentatifs des plutons du Groupe de Parent. Pour les symboles voir la figure 3.

Table I. Caractéristiques générales des plutons de la Fosse de l'Ungava.

Lithologie	Nom du pluton	Composition	Age (Ma)	État de la déformation du pluton
GROUPE DE WATTS	P23-87	tonalite	1870-1880 (P)	folié
	S162A-86	tonalite	1876 (P)	folié
	CP5	diorite, monzodiorite	1858 (M)	peu folié
	D3	monzodior.qtz-granodiorite		folié
	Z2-84	tonalite - granodiorite	1848 (P)	non-déformé
	Z4-84	diorite qtz		non-déformé
GROUPE DE SPARTAN	D4	tonalite - diorite qtz		peu folié
	CP3	tonalite		peu folié
	D2	diorite		non-déformé
	Z5-84	tonalite	>1850 (P)	non-déformé
GROUPE DE PARENT	D5, CP6 Z3-84	diorite, tonalite diorite qtz	1898 (P)	folié
	D5	diorite	1874 (M)	????
	CP2	diorite	1860 (M)	non-déformé
	DL1	granodiorite, diorite	1845 (P)	folié
	DL8	tonalite diorite		peu folié

(P) = géochronologie par Parrish (1989)
(M) = géochronologie par Machado et al. (1993)

4.8.2 Groupe de Watts

La déformation dans le Groupe de Watts était commencée à partir de 1876 à 1880 Ma, soit l'âge des tonalites déformées (P23-87, S162A-86; Parrish, 1989) qui recoupent des failles de chevauchement précoces mais qui ont été déplacées par des chevauchements tardifs. La déformation paraît avoir subi un déclin à partir de 1858 Ma, l'âge du pluton CP5 (Machado *et al.*, 1993), un gabbro-monzodiorite faiblement folié (syn-tectonique). Le pluton D3, une monzodiorite quartzique à granodiorite, foliée et pré-tectonique, a donc dû faire intrusion avant 1858 Ma. Un pluton de tonalite-granodiorite non-déformé post-tectonique (Z2-84) limite la fin de la déformation dans le Groupe de Watts avant 1848 Ma (Parrish, 1989).

4.8.3 Groupe de Spartan

Dans le Groupe de Spartan, une chronologie relative de la mise en place des plutons peut être établie en se basant sur leur état de déformation. Les plutons D4 et CP3, faiblement foliés, ont été mis en place avant D2, qui ne possède aucune foliation et qui est donc post-tectonique. Un petit pluton de tonalite massive du Groupe de Spartan, situé au nord de D4 (non inclus sur la carte de compilation), a été daté et un âge minimum de 1850 Ma a été obtenu (Parrish, 1989; échantillon Z4-84). Ceci indique que la déformation dans le Groupe de Spartan avait cessé à partir de cette date.

4.8.4 Groupe de Parent

En se basant sur l'âge de 1898 Ma de Parrish (1989), une série de plutons foliés et déformés (D5, CP6, Z3-84) sont interprétés comme les plus anciennes intrusions dans le Groupe de Parent. Un second pluton déformé (DL1) du Groupe de Parent a été daté par Machado *et al.* (1993) à 1845 Ma, ce qui suggèrerait une prolongation de la déformation au moins de 1898 à 1845 Ma. La présence d'un pluton de gabbro/diorite non-déformé post-tectonique (CP2, aussi du Groupe de Parent) contredit cette interprétation puisqu'il a été daté à 1860 Ma (Machado *et al.*, 1993). Cette contradiction peut être levée car les plutons D5, CP6, Z3-84 et CP2 sont trouvés dans la même nappe de charriage tandis que DL1 se trouve dans une nappe située plus à l'ouest. Ils pourraient donc avoir subi la déformation à des temps différents. Donc, dans la région centrale de la Fosse de l'Ungava, la déformation du Groupe de Parent était en cours par 1898 Ma et était terminée par 1860 Ma. À l'ouest, la déformation dans le Groupe de Parent a continué jusqu'à 1845 Ma.

4.9 PÉTROGENÈSE DES PLUTONS DE LA FOSSE DE L'UNGAVA

La composition des plutons surtout calco-alcalins et métalumineux de la Fosse de l'Ungava varie de dioritique à granodioritique. Les plutons de composition similaire possèdent des signatures géochimiques comparables et tous les plutons sont caractérisés par un enrichissement variable en éléments hygromagmaphiles, par les terres rares fractionnées et par les anomalies négatives de Nb et Ti. En général cependant, le plus bas contenu de terres rares lourdes dans les plutons plus felsiques suggère que le fractionnement du grenat a joué un rôle important dans la pétrogenèse des plutons de tonalite et de granodiorite. Les plutons sont tous isotopiquement juvéniles et sont dérivés d'une source mantellique appauvrie ayant subi une contamination mineure. La similarité générale des compositions ainsi que la disposition géographique et temporelle de ces plutons, des roches du terrane de l'arc Narsajuaq et des roches volcaniques du Groupe de Parent suggèrent que ces unités ont subi une pétrogenèse semblable. Ces similarités ainsi que les caractéristiques minéralogiques et géochimiques des plutons mentionnées ci-haut indiquent fortement que ces plutons ont été générés dans un milieu de subduction tel qu'un arc magmatique.

Approximativement 30 Ma avant la construction de l'arc de Narsajuaq, le plutonisme a commencé dans les groupes de Watts, de Parent et de Spartan dans la Fosse de l'Ungava. Entre env. 1860 et 1840 Ma, l'activité plutonique avait lieu concurremment après quoi, elle a cessé dans les lithologies de la Fosse de l'Ungava et a continué pour au moins 14 Ma additionnels dans l'arc de Narsajuaq. Les implications de ces restrictions géochronologiques et d'un modèle général pour la pétrogenèse des plutons sont discutées dans la section suivante.

4.10 INTERPRÉTATION TECTONOSTRATIGRAPHIQUE DE L'OROGÈNE DE L'UNGAVA

L'origine de la Fosse de l'Ungava a longtemps été controversée. Les premiers modèles suggéraient que la Fosse de l'Ungava était une ceinture de plissement ensialique (Dimroth *et al.*, 1970; Baer, 1977; Baragar et Scoates, 1981). Une autre interprétation était que la bande représente le site d'une géosuture de collision majeure (Gibb et Walcott, 1971; Burke *et al.*, 1977; Thomas et Gibb, 1977). Hynes et Francis (1982), Francis *et al.* (1981, 1983) ainsi que Picard *et al.* (1990) suggéraient que la bande était le site d'un petit bassin océanique éphémère. Hoffman (1985) proposait que la bande était essentiellement un klippe et que la géosuture était située plus au nord et séparée de la bande par un antiforme qui post-datait le chevauchement.

Les travaux récents du MERQ dans l'ouest et de la CGC dans l'est et l'arrière-pays de la bande ont engendré des découvertes importantes sur la nature de la région du Cap Smith: (1) la présence d'affleurement continu du socle de la Province du Supérieur autour de l'extrémité est de la Fosse de l'Ungava (St-Onge et Lucas, 1993); (2) l'existence de deux demi-fenêtres de socle autochtone au coeur d'un antiforme à grande échelle, l'antiforme Kovik (Hoffman, 1985), au nord de la bande (Lucas et St-Onge, 1991; St-Onge et Lucas, 1992); (3) la préservation de l'ophiolite Purtuniq de env. 2,00 Ga dans la Fosse de l'Ungava (St-Onge *et al.*, 1988; Scott *et al.*, 1989, 1991); (4) l'interprétation de l'origine des roches volcaniques et volcaniclastiques du Groupe de Parent dans un milieu d'arc de subduction (Picard *et al.*, 1990; Picard, en préparation); et (5) la caractérisation de l'arrière-pays de la Fosse de l'Ungava comme terrane d'arc magmatique de 1,86 à 1,83 Ga (St-Onge et Lucas, 1990a, 1992; Lucas et St-Onge, 1991).

Ces nouvelles découvertes se sont ajoutées aux travaux antérieurs dans la région et ont permis à St-Onge *et al.* (1992) de développer un modèle de l'évolution de l'orogène de l'Ungava pendant le Protérozoïque inférieur (voir aussi Lucas et St-Onge, 1992; Lucas *et al.*, 1992). Le développement d'un rift dans la croûte continentale de la Province du Supérieur paraît avoir été en cours vers 2038 Ma (Machado *et al.*, 1993), ceci étant l'âge d'un filon-couche de gabbro faisant intrusion dans les roches sédimentaires du Groupe de Povungnituk inférieur. La sédimentation et le magmatisme Povungnituk ont continué au moins jusqu'à 1959 Ma (Parrish, 1989) si on se base sur l'âge d'une rhyolite au sommet du Groupe de Povungnituk. L'activité d'un point chaud aurait généré la volumineuse séquence volcanique du Groupe de Povungnituk pendant le Protérozoïque inférieur et aurait provoqué le développement du rift (Lucas *et al.*, 1992; Scott *et al.*, 1992; St-Onge *et al.*, 1992; Dunphy *et al.*, 1994). La similarité des basaltes Povungnituk et des tholéiites continentales modernes, la présence de coulées de basalte et de rhyolite interlitées et celles de basaltes alcalins près du sommet de la séquence volcanique soutiennent l'hypothèse du dépôt des basaltes Povungnituk à la marge d'un rift (Hynes et Francis, 1982; Francis *et al.*, 1983; Gaonac'h *et al.*, 1989, 1992; Picard, 1989b; Picard *et al.*, 1990; St-Onge et Lucas, 1990b).

Les produits du magmatisme Chukotat paraissent enregistrer la transition entre l'accumulation de roches volcaniques continentales de rift et la formation d'une croûte océanique transitionnelle (Francis et Hynes,1979; Francis *et al.*, 1981, 1983; Hynes et Francis, 1982; Picard, 1989a, 1989b; Picard *et al.*, 1990). Un filon-couche de ferrogabbro quartzique intrusif dans le Groupe de Povungnituk inférieur permet de dater le Groupe de Chukotat à 1918 Ma (Parrish, 1989). Ce filon-couche est considéré consanguin avec des filons-couches similaires trouvés dans le Groupe de Chukotat et pourrait représenter le système d'alimentation magmatique de celui-ci (Bédard *et al.*, 1984; Thibert *et al.*, 1989; Barnes et Giovenazzo, 1990). L'apparente lacune de magmatisme entre 1959 et 1918 Ma n'a pas encore été bien comprise.

L'ophiolite du Groupe de Watts a été datée à 1998 Ma (Parrish, 1989), un âge considérablement plus ancien que celui des roches du Groupe de Chukotat, qui sont structuralement sous-jacentes. Le caractère géochimique et isotopique des roches du

Groupe de Watts indique que l'ophiolite a été formée à une dorsale médio-océanique (suite MORB) dans laquelle ont fait intrusion des roches d'affinité de basaltes d'île océanique (suite OIB) (Hegner et Bevier, 1989, 1991; Scott, 1990; Scott *et al.*, 1991, 1992). Deux modèles tectoniques principaux ont été proposés pour expliquer l'origine du Groupe de Watts. Picard *et al.* (1989, 1990) suggèrent que le Groupe de Watts ait été formé dans un bassin plus ancien qui aurait précédé l'ouverture d'un second bassin au sud où se seraient accumulés les groupes de Povungnituk et de Chukotat. Un modèle alternatif n'impliquerait qu'un seul bassin et nécessiterait un système de rift propageant (analogue à la Mer Rouge actuelle) qui se développerait avec la juxtaposition éventuelle de la croûte océanique plus ancienne (Groupe de Watts) avec la croûte océanique transitionnelle plus jeune (Groupe de Chukotat) le long d'une marge de plaque transformante (Hegner et Bevier, 1989; Scott *et al.*, 1989, 1991; St-Onge *et al.*, 1989, 1992). Le problème avec le premier modèle est qu'aucune trace de la croûte qui aurait séparé les deux bassins n'a été trouvée (St-Onge *et al.*, 1989; Picard *et al,*, 1990; St-Onge et Lucas, 1993). Une difficulté avec le second modèle est l'absence d'évidence structurale pour la translation latérale du Groupe de Watts plus ancien, relativement au Groupe de Chukotat plus jeune (Picard *et al.*, 1990; St-Onge *et al.*, 1992).

Le Groupe de Spartan et le Groupe de Parent sont trouvés intercalés entre les nappes de charriage mafiques et ultramafiques du Groupe de Watts et la croûte océanique transitionnelle du Groupe de Chukotat. Les roches du Groupe de Parent sont interprétées comme étant des dépôts associés au magmatisme calco-alcalin d'une zone de subduction (Picard *et al.*, 1990; Picard, en préparation). Les travaux de Lamothe et Picard (en rédaction) indiquent une contamination crustale des magmas du Groupe de Parent. Les roches sédimentaires du Groupe de Spartan ont été interprétées comme sédiments pélagiques (St-Onge *et al.*, 1992). La similarité de la position structurale du Groupe de Spartan et du Groupe de Parent suggère qu'ils pourraient être des équivalents latéraux, le Groupe de Spartan représentant un prisme clastique en marge de l'arc de Parent (tel un dépôt de bassin avant-arc) (St-Onge *et al.*, 1992). Alternativement, les sédiments Spartan pourraient aussi bien représenter des dépôts distaux associés avec la marge nord du rift de la Province du Supérieur (St-Onge *et al.*, 1992). Le complexe plutonique Narsajuaq, situé dans l'arrière-pays de la Fosse de l'Ungava, a été interprété comme le noyau plutonique d'un arc magmatique (Dunphy et Ludden, 1992; Lucas *et al.*, 1992; St-Onge *et al.*, 1992). Les données préliminaires géochimiques et isotopiques ont mené Lucas *et al.* (1992) à suggérer que l'arc de Narsajuaq a été construit sur la croûte océanique Watts (lors de la subduction vers le nord de la croûte Chukotat) et que le Groupe de Parent pourrait être l'équivalent extrusif de cette activité. Des études subséquentes ont montré que la pétrogenèse des magmas Narsajuaq était encore plus complexe, nécessitant la contamination de magmas appauvris par des composantes crustales (Dunphy *et al.*, 1993) et indiquant la contribution de la croûte continentale (ou d'un fragment continental; Dunphy, en préparation).

Le magmatisme associé à la subduction a résulté en la mise en place d'une série de plutons entre 1898 et 1840 Ma dans les unités de la Fosse de l'Ungava. Ces plutons sont restreints aux Groupe de Watts, Spartan et Parent et leur absence dans le socle de la Province de Supérieur sous-jacent et dans les groupes de Povungnituk et Chukotat suggère que ces dernières unités faisaient partie de la plaque en subduction et n'étaient donc pas au large du système de fosses. Les variations isotopiques et géochimiques mineures observées dans les plutons ont été produites par une contamination mineure des magmas (provenant d'une source mantellique appauvrie), par des sédiments (entraînés par la subduction) et des processus magmatiques subséquents tels que le fractionnement.

Dans le modèle proposé par St-Onge *et al.* (1992; Fig. 13), la subduction vers le nord du bassin océanique aurait été suivie par un changement de polarité de subduction qui aurait eu comme résultat la formation d'une zone de subduction plus jeune, avec pendage vers le sud et la génération de la plus jeune suite plutonique de l'arc de Narsajuaq. Le changement de polarité est interprété comme résultant de la tentative de subduction de la lithosphère continentale de la Province du supérieur sous la croûte

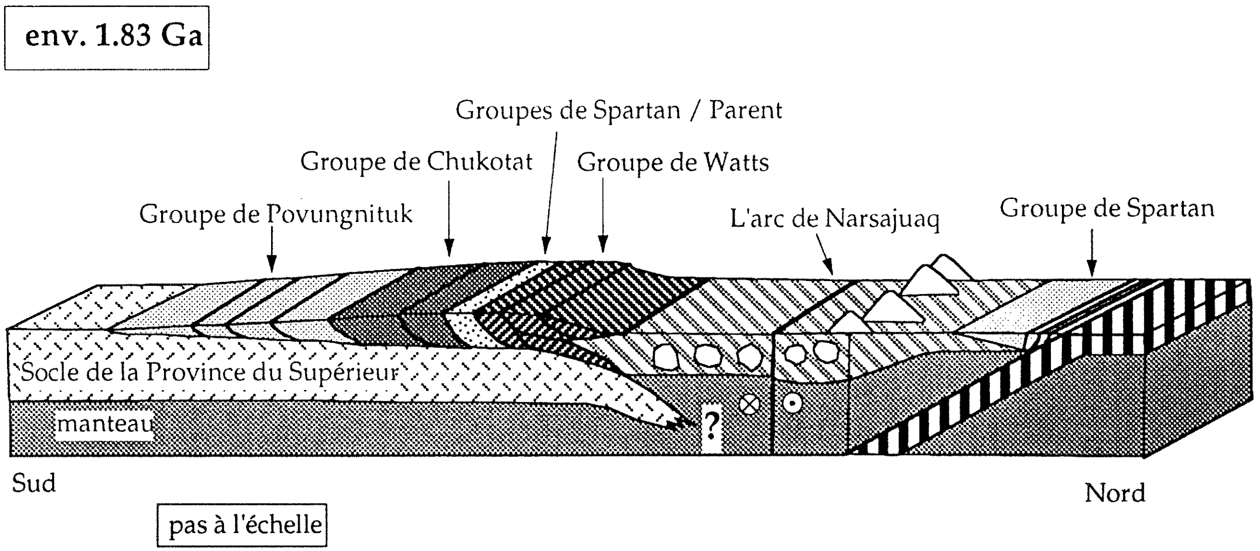


Figure 4.13. Diagramme schématique illustrant la formation de l'arc de Narsajuaq suivant le changement de polarité de subduction, d'après le modèle de St-Onge *et al.* (1992). À ce moment dans l'évolution de l'orogène, les unités de la bande du Cap Smith ont été accrétées au socle de la Province du Supérieur et charriées sur celui-ci.

océanique/d'arc sous-jacente. Le plutonisme Cap Smith aurait dû avoir lieu avant le changement de polarité puisque ces plutons sont les plus anciens ayant des caractéristiques d'environnement d'arc. Un modèle pour la pétrogenèse de ces plutons qui soit en harmonie avec le modèle de St-Onge *et al.* (1992), impliquerait la subduction vers le nord de la croûte océanique en-dessous du bassin océanique restant (Groupe de Watts et Groupe de Spartan). Cette subduction aurait alimenté les plutons dans ces unités et aurait aussi généré l'arc volcanique/plutonique Parent. Le changement de polarité du système de subduction aurait résulté en la construction des deux suites (ancienne et jeune) de l'arc de Narsajuaq en partie sur un fragment de croûte continentale (Dunphy, en préparation). Une autre hypothèse n'impliquant aucun changement de polarité du système de subduction est celle où le subduction serait dirigée de façon continue vers le nord sous une zone de transition entre un bassin océanique et la croûte continentale (Dunphy, en préparation). Une meilleure connaissance de la géologie de la région au nord du terrane Narsajuaq (i.e. sur l'Ile Baffin) est nécessaire pour mieux raffiner cette partie du modèle. L'accrétion finale du terrane d'arc au socle de la Province du Supérieur est présumément postérieure à 1826 Ma (l'âge du pluton de milieu d'arc le plus jeune) mais antérieure à 1758 Ma (l'âge de dykes de syénogranite pegmatitique qui recoupent toutes les roches de l'orogène) (St-Onge *et al.*, 1992; Lucas et St-Onge, 1992).

4.11 CONCLUSIONS

Le plutonisme dans la Fosse de l'Ungava s'est produit pendant le long intervalle d'env. 1880 et 1840 Ma dans le Groupe de Watts et 1898 et 1845 Ma dans le Groupe de Parent. Une grande variété de compositions de plutons, surtout dioritiques à granodioritiques, est observée quoique les compositions mafiques sont les plus communes. Les plutons contiennent généralement de la hornblende et/ou de la biotite bien qu'un métamorphisme régressif au faciès des schistes verts a produit d'abondants minéraux secondaires tels l'épidote et la chlorite qui remplacent les minéraux primaires.

La signature géochimique et isotopique des plutons ainsi que les caractéristiques de leur compositions suggèrent que ces plutons ont été générés dans un environnement de subduction tel qu'un arc magmatique. Les relations chronologiques et géographiques ainsi que les similarités géochimiques générales entre les plutons du Cap Smith et les produits de l'activité ignée dans le Groupe de Parent et le terrane de Narsajuaq suggèrent qu'ils sont reliés pétrogénétiquement.

Un modèle est proposé pour la pétrogenèse des plutons et impliquerait la subduction vers le nord de la croûte océanique sous le Groupe de Watts et Spartan, produisant les plutons retrouvés dans ces unités et créant l'arc de Parent. Le changement de polarité du système de subduction (peut-être aux marges d'un petit fragment continental) aurait généré le complexe plutonique Narsajuaq. Pendant cet intervalle, seul un plutonisme mineur se serait produit dans les unités de la Fosse de l'Ungava.

RÉFÉRENCES

BAER, A.J., 1977 -- Gravity anomalies and deep structure of the Cape Smith Fold Belt, northern Ungava, Quebec: comment and reply. Geology; volume 5, pages 651-653.

BARAGAR, W.R.A., 1974 -- Volcanic studies in the Cape Smith - Wakeham Bay Belt, New Quebec. *In:* Report of Activities, Part A. Geological Survey of Canada; Paper 74-1A, pages 155-157.

BARAGAR, W.R.A., 1984 -- Pillow formation and layered flows in the Circum-Superior Belt of eastern Hudson Bay. Canadian Journal of Earth Sciences; volume 21, pages 781-792.

BARAGAR, W.R.A. et SCOATES, R.F.J., 1981 -- The Circum-Superior Belt: a Proterozoic plate margin? *In* Precambrian Plate Tectonics (A. Kroner, (éditor)). Elsevier, New York, pages 297-330.

BARNES, S.-J. et GIOVENAZZO, D., 1990 -- Platinum-group elements in the Bravo intrusion, Cape Smith Fold Belt, northern Quebec. Canadian Mineralogist; volume

28, pages 431-449.

BARRETTE, P.D., 1988 -- Géologie de la région du lac Bilson, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 87-23.

BARRETTE, P.D., 1989 -- Géologie de la région du lac Bolduc, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 88-17.

BARRETTE, P.D., 1990 -- Géologie de la région du lac Bolduc (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 89-03.

BEALL, G.H., 1959 -- Cross Lake area, New Quebec. Quebec Department of Mines, Preliminary Report 396, Preliminary Map 1267.

BEALL, G.H., 1960 -- Laflamme Lake area, New Quebec. Quebec Department of Mines, Preliminary Report 435, Preliminary Map 1351.

BÉDARD, J.H., FRANCIS, D.M., HYNES, A.J. et NADEAU, S., 1984 -- Fractionation in the feeder system at a Proterozoic rifted margin. Canadian Journal of Earth Sciences; volume 21, pages 489-499.

BELL, R., 1885 -- Observations on the geology, mineralogy, zoology and botany of the Labrador Coast, Hudson Strait and Bay. Geological Survey of Canada, Report of Progress, 1882-84.

BERGERON, R., 1957 -- Cape Smith - Wakeham Bay belt, New Quebec, Quebec. Quebec Department of Mines, Preliminary Report 355, Preliminary Maps 1090, 1196,

BERGERON, R., 1959 -- Povungnituk range area, New Quebec. Quebec Department of Mines, Preliminary Report 392, Preliminary Map 1279.

BROWN, G.C., 1982 -- Calc-alkaline intrusive rocks: their diversity, evolution and relation to volcanic arcs. *In* Andesites (R.S. Thorpe, (éditors)). John Wiley and Sons, New York, pages 437-464.

BURKE, K., DEWEY, J.F. et KIDD, W.S.F., 1977 -- World distribution of sutures - the sites of former oceans. Tectonophysics; volume 40, pages 69-99.

DEMONTIGNY, P.A., 1959 -- Upper Deception River area, New Quebec. Quebec Department of Mines, Preliminary Report 398, Preliminary Map 1285.

DIMROTH, E., BARAGAR, W.R.A., BERGERON, R. et JACKSON, G.D., 1970 -- The filling of the Circum-Ungava geosyncline. *In* Symposium on basins and geosynclines of the Canadian Shield (A.J. Baer, (éditor)). Geological Survey of Canada; Paper 70-40, pages 45-142.

DUNPHY, J.M. et LUDDEN, J.N., 1992 -- The Narsajuaq terrane, northern Quebec: a mid-crustal section of an Early Proterozoic magmatic arc. Eos Transactions (American Geophysical Union), volume 73, page 318.

DUNPHY, J.M., LUDDEN, J.N. et FRANCIS, D., 1994 -- Geochemistry of mafic magmas from the Ungava orogen, Quebec, Canada: Implications for mantle reservoir compositions at 2.0 Ga. Chemical Geology; submitted.

DUNPHY, J.M., LUDDEN, J.N. et FRANCIS, D.M., 1993 -- Geochemistry of magmas from the Ungava orogen, Quebec, Canada: Implications for mantle evolution proir to 2.0 Ga. International Association of Volcanology and Chemistry of the Earth's Interior General Assembly 1993, Book of Abstracts, page 28.

FEININGER, T., 1986 -- An unusual alaskite located 40 km west of Asbestos Hill. *In* Exploration en Ungava: données récents sur la géologie et la gïtologie (D. Lamothe, R. Gagnon et T. Clark, (éditor)). Ministère de l'Energie et des Ressources du Québec; DV 86-16, pages 51-56.

FRANCIS, D.M. et HYNES, A.J., 1979 -- Komatiite-derived tholeiites in the Proterozoic of New Quebec. Earth and Planetary Science Letters; volume 44, pages 473-481.

FRANCIS, D.M., HYNES, A.J., LUDDEN, J.N. et BÉDARD, J., 1981 -- Crystal fractionation and partial melting in the petrogenesis of a Proterozoic high-MgO volcanic suite, Ungava Quebec. Contributions to Mineralogy and Petrology; volume 78, pages 27-36.

FRANCIS, D.M., LUDDEN, J.N. et HYNES, A.J., 1983 -- Magma evolution in a Proterozoic rifting environment. Journal of Petrology; volume 24, pages 556-582.

GAONAC'H, H., LUDDEN, J.N., PICARD, C. et FRANCIS, D.M., 1992 -- Highly alkaline lavas in a Proterozoic rift zone: Implications for Precambrian mantle metasomatic processes. Geology; volume 20, pages 247-250.

GAONAC'H, H., PICARD, C., LUDDEN, J.N. et FRANCIS, D.M., 1989 -- Alkaline rocks from a Proterozoic volcanic island in the Cape Smith Thrust Belt, New

Quebec. Geoscience Canada; volume 16, pages 137-139.

GÉLINAS, L., 1962 -- Watts Lake area, New Quebec. Quebec Department of Natural Ressources, Preliminary Report 471, Preliminary Map 1414.

GIBB, R.A. et WALCOTT, R.I., 1971 -- A Precambrian suture in the Canadian Shield. Earth and Planetary Science Letters; volume 10, pages 417-422.

GOLD, D.P., 1962 -- Brisebois Lake area, New Quebec. Quebec Department of Natural Resources, Preliminary Report 470, Preliminary Map 1413.

HEGNER, E. et BEVIER, M.L., 1989 -- Geochemical constraints on the origin of mafic rocks from the Cape Smith Belt. Geoscience Canada; volume 16, pages 148-151.

HEGNER, E. et BEVIER, M.L., 1991 -- Nd and Pb isotopic constraints on the origin of the Purtuniq ophiolite and Early Proterozoic Cape Smith Belt, northern Quebec, Canada. Chemical Geology; volume 91, pages 357-371.

HERVET, M., 1985 -- Région de la vallée de Narsajuaq, Ungava. Ministère de l'Energie et des Ressources du Québec, DP 85-05.

HOFFMAN, P.F., 1985 -- Is the Cape Smith Belt (northern Quebec) a klippe? Canadian Journal of Earth Sciences; volume 22, pages 1361-1369.

HOFFMAN, P.F., 1990 -- Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen. *In* The Early Proterozoic Trans-Hudson Orogen of North America (J.F. Lewry et M.R. Stauffer, (éditors)). Geological Association of Canada; Special Paper 37, pages 15-39.

HYNES, A.J. et FRANCIS, D.M., 1982 -- A transect of the Early Proterozoic Cape Smith foldbelt, New Quebec. Tectonophysics; volume 88, pages 23-59.

IRVING, T.N. et BARAGAR, W.R.A., 1971 -- A guide to the chemical classification of the common volcanic rocks. Canadian Journal of Earth Sciences; volume 8, pages 523-548.

LAMOTHE, D., GAGNON, R. et CLARKE, T. (éditors), 1986 -- Exploration en Ungava: données récents sur la géologie et la gîtologie. Séminaire d'information, Ministère de l'Energie et des Ressources du Québec, DV 86-16.

LAMOTHE, D. et PICARD, C., 1994 -- Synthèse géologique de la Fosse de l'Ungava. Ministère des Ressources Naturelles, MB 95-01.

LAMOTHE, D., PICARD, C. et MOORHEAD, J., 1983 -- Région du lac Beauparlant, Bande de Cap Smith - Maricourt. Ministère de l'Energie et des Ressources du Québec, DP 84-39.

LE MAITRE, R.W. (éditor), 1989 -- A Classification of Igneous Rocks and a Glossary of Terms. Blackwell, Oxford, 193 pages.

LEWRY, J.F. et COLLERSON, K.D., 1990 -- The Trans-Hudson Orogen: extent, subdivision and problems. *In* The Early Proterozoic Trans-Hudson Orogen of North America (J.F. Lewry et M.R. Stauffer, (éditors)). Geological Association of Canada; Special Paper 37, pages 1-14.

LOW, A.P., 1899 -- Report on an exploration of part of the south shore of Hudson Strait and Ungava Bay. Geological Survey of Canada, Annual Report, vol. XI, part L.

LOW, A.P., 1902 -- Report on an exploration of the east coast of Hudson Bay from Cape Wolstenholme to the south end of James Bay. Geological Survey of Canada, Annual Report, volume XIII, part D.

LUCAS, S.B. et ST-ONGE, M.R., 1991 -- Evolution of Archean and Early Proterozoic magmatic arcs in northeastern Ungava Peninsula, Quebec. *In:* Current Research, Part C. Geological Survey of Canada; Paper 91-1C, pages 109-119.

LUCAS, S.B. et ST-ONGE, M.R., 1992 -- Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 2: Structural and metamorphic history. Canadian Journal of Earth Sciences; volume 29, pages 765-782.

LUCAS, S.B., ST-ONGE, M.R., PARRISH, R.R. et DUNPHY, J.M., 1992 -- Long-lived continent-ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. Geology; volume 20, pages 113-116.

MACHADO, N., DAVID, J., SCOTT, D.J., LAMOTHE, D., PHILIPEE, S. et GARIÉPY, C., 1994 -- U-Pb geochronology of the western Cape Smith Belt, Canada: New insights on the age of initial rifting and arc magmatism. Precambrian Research; volume 63, pages ?

MAHLBURG KAY, S. et RAPELA, C.W. (éditors), 1990 -- Plutonism from Antarctica to Alaska. Geological Society of America; Special Paper 241, 263 pages.

MANIAR, P.D. et PICCOLI, P.M., 1989 -- Tectonic discrimination of granitoids.

Geological Society of America Bulletin; volume 101, pages 635-643.

MOORHEAD, J., 1985 -- Géologie de la région du lac Chukotat. Ministère de l'Energie et des Ressources du Québec, DP 86-01.

MOORHEAD, J., 1987 -- Géologie de la région du lac Hubert, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 86-31.

MOORHEAD, J., 1988 -- Géologie de la région du lac Vigneau, Nouveau Québec. Ministère de l'Energie et des Ressources du Québec, DP 88-05.

MOORHEAD, J., 1989 -- Géologie de la région du lac Chukotat (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 87-10.

PARRISH, R.R., 1989 -- U-Pb geochronology of the Cape Smith Belt and Sugluk block, northern Quebec. Geoscience Canada; volume 16, pages 126-130.

PEARCE, J.A., HARRIS, N.B.W. et TINDLE, A.G., 1984 -- Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. Journal of Petrology; volume 25, pages 956-983.

PICARD, C., 1989a -- Lithochimie des roches volcaniques protérozoïques de la partie occidentale de la Fosse de l'Ungava (région au sud du lac Lanyan). Ministère de l'Energie et des Ressources du Québec, ET 87-14.

PICARD, C., 1989b -- Pétrologie et volcanologie des roches volcaniques protérozoïques de la partie centrale de la Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, ET 87-07.

PICARD, C., in preparation -- Caractéristiques pétrographiques et géochimiques des roches volcaniques du Groupe de Parent. Ministère de l'Energie et des Ressources du Québec.

PICARD, C., GIOVENAZZO, D. et LAMOTHE, D., 1989 -- Geotectonic evolution by asymmetric rifting of the Proterozoic Cape Smith Belt, New Quebec. Geoscience Canada; volume 16, pages 130-134.

PICARD, C., LAMOTHE, D., PIBOULE, M. et OLIVER, R., 1990 -- Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt (New-Quebec). Precambrian Research; volume 47, pages 223-249.

ROY, C., 1989 -- Géologie de la région du lac Bélanger, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, MB 89-13.

SCHWARZ, E.J. et FUJIWARA, Y., 1977 -- Komatiitic basalts from the Proterozoic Cape Smith Range in northern Quebec, Canada. *In* Volcanic regimes in Canada (W.R.A. Baragar, L.C. Coleman et J.M. Hall, (éditors)). Geological Association of Canada; Special Paper 16, pages 193-201.

SCOTT, D.J., 1990 -- Geology and geochemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. Queens University (Kingston, Ontario); thèse de doctorat, inédit.

SCOTT, D.J., HELMSTAEDT, H. et BICKLE, M.J., 1992 -- Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. Geology; volume 20, pages 173-176.

SCOTT, D.J., ST-ONGE, M.R., LUCAS, S.B. et HELMSTAEDT, H., 1989 -- The 1998 Ma Purtuniq ophiolite: imbricated and metamorphosed oceanic crust in the Cape Smith Thrust Belt, northern Quebec. Geoscience Canada; volume 16, pages 144-147.

SCOTT, D.J., ST-ONGE, M.R., LUCAS, S.B. et HELMSTAEDT, H., 1991 -- Geology and chemistry of the Early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. *In* Ophiolite Genesis and Evolution of the Oceanic Lithosphere (T. Peters et al., (éditors)). Kluwer Academic Publishers, Amsterdam, pages 817-849.

ST-ONGE, M.R. et LUCAS, S.B. 1989 -- Geology, eastern portion of the Cape Smith Thrust-Fold Belt, parts of the Wakeham Bay, Cratère du Nouveau-Québec and Nuvilik Lakes map areas, northern Quebec. Geological Survey of Canada; Maps 1721A-1735A, scale 1:50,000 (carte annotée).

ST-ONGE, M.R. et LUCAS, S.B., 1990a -- Early Proterozoic collisional tectonics in the internal zone of the Ungava (Trans-Hudson) orogen, Lacs Nuvilik and Sugluk map areas, Quebec. *In:* Current Research, Part C. Geological Survey of Canada; Paper 90-1C, pages 119-132.

ST-ONGE, M.R. et LUCAS, S.B., 1990b -- Evolution of the Cape Smith Belt: Early

Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. *In* The Early Proterozoic Trans-Hudson Orogen of North America (J.F. Lewry et M.R. Stauffer, (éditors)). Geological Association of Canada; Special Paper 37, pages 313-351.

ST-ONGE, M.R. et LUCAS, S.B., 1992 -- New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. *In:* Current Research, Part C. Geological Survey of Canada; Paper 92-1C, pages 31-41.

ST-ONGE, M.R. et LUCAS, S.B., 1993 -- Geology of the eastern Cape Smith Belt: Parts of the Kangiqssujuaq, Cratère du Nouveau-Québec, and Lacs Nuvilik map areas, Quebec. *In:* Geological Survey of Canada; Memoir 438, pages 110.

ST-ONGE, M.R., LUCAS, S.B. et PARRISH, R.R., 1992 -- Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Canadian Journal of Earth Science; volume 29, pages 746-764.

ST-ONGE, M.R., LUCAS, S.B., SCOTT, D.J. et BÉGIN, N.J., 1986 -- Eastern Cape Smith Belt: an Early Proterozoic thrust-fold belt and basal shear zone exposed in oblique section, Wakeham Bay and Cratère du Nouveau-Québec map areas, northern Quebec. *In:* Current Research, Part A. Geological Survey of Canada; Paper 86-1A, pages 1-14.

ST-ONGE, M.R., LUCAS, S.B., SCOTT, D.J. et BÉGIN, N.J., 1987 -- Tectonostratigraphy and structure of the Lac Watts - Lac Cross - Rivière Déception area, central Cape Smith Belt, northern Quebec. *In:* Current Research, Part A. Geological Survey of Canada; Paper 87-1A, pages 619-632.

ST-ONGE, M.R., LUCAS, S.B., SCOTT, D.J. et BÉGIN, N.J., 1989 -- Evidence for the development of oceanic crust and for continental rifting in the tectonostratigraphy of the Early Proterozoic Cape Smith Belt. Geoscience Canada; volume 16, pages 119-122.

ST-ONGE, M.R., LUCAS, S.B., SCOTT, D.J., BÉGIN, N.J., HELMSTAEDT, H. et

CARMICHEAL, D.M., 1988 -- Thin-skinned imbrication and subsequent thick-skinned folding of rift-fill, transitional-crust and ophiolite suites in the 1.9 Ga Cape Smith Belt, northern Quebec. *In:* Current Research, Part C. Geological Survey of Canada; Paper 88-1C, pages 1-18.

STAM, J.C., 1961 -- On the geology and petrology of the Cape Smith - Wakeham Bay Belt, Ungava, New Quebec. Geologie en Mijnbouw; volume 40, pages 412-421.

STRECKEISEN, A.L., 1976 -- To each plutonic rock its proper name. Earth Science Reviews; volume 12, pages 1-33.

SUN, S.S. et MCDONOUGH, W.F., 1989 -- Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. *In* Magmatism in the ocean basins (A.D. Saunders et M.J. Norry, (éditors)). Geological Society; Special Publication 42, pages 313-345.

TAYLOR, F.C., 1974 -- Reconnaissane geology of a part of the Precambrian Shield, northern Quebec and Northwest Territories. *In:* Geological Survey of Canada; Paper 74-21, pages 10.

TAYLOR, F.C., 1982 -- Reconnaissance geology of a part of the Canadian Shield, northern Quebec and Northwest Territories. *In:* Geological Survey of Canada; Memoir 399, pages 32.

THIBERT, F., PICARD, C. et TRZCIENSKI, W., 1989 -- Pétrologie des filons-couches difféмnciés Roméo 1 et 2 dans la partie centrale de la bande du Cap Smith. Geoscience Canada; volume 16, pages 140-144.

THOMAS, M.D. et GIBB, R.A., 1977 -- Gravity anomalies and deep structure of the Cape Smith Fold Belt, northern Ungava, Quebec. Geology; volume 5, pages 169-172.

TREMBLAY, G., 1987 -- Géologie de la région du lac Lessard, Fosse de l'Ungava. Ministère de l'Energie et des Ressources du Québec, DP 86-30.

TREMBLAY, G., 1989 -- Géologie de la région du lac Vanasse (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 87-08.

TREMBLAY, G., 1991 -- Géologie de la région du lac Lessard (Fosse de l'Ungava). Ministère de l'Energie et des Ressources du Québec, ET 88-09.

ANNEXE 1

Compositions recalculées (100% anhydre) et géochimie des éléments majeurs des plutons de la Fosse de l'Ungava.

PLUTON	D3	D3	CP5	CP5	CP5	CP5	CP5	CP5	CP5	CP5
NUMERO	84015706	84019556	89016071	89016072	89016058	89016059	89016060	89016061	89016073	89016075
ESTANT	498400	497900	467631	467934	467934	471012	471319	471833	474950	
NORDANT	6853625	6855700	6842598	6842660	6842655	6842655	6842702	6842701	6841295	
ECHAN	MH84-3243A	MH-84-3245A	CP-89-P-2	CP-89-P-3A	CP-89-P-3B	CP-89-P-6A	CP-89-P-7	CP-89-P-9	CP-89-P-13	CP-89-P-15
LAME MINCE	3243A(J)	3245A	L-PER-2	L-PER-3A	L-PER-3B	L-PER-6A	L-PER-7	L-PER-9	L-PER-13	L-PER-15
LITHO	Monzdior.qtz	Monzdior.qtz	Diorite	Diorite	Diorite	Monzdior.	Monzdior.	Monzdior.	Monzdior.	Monzdior.
SIO2	68.00	62.00	50.80	51.00	51.10	56.00	52.60	51.90	52.90	52.00
Al2O3	18.00	18.60	16.40	19.30	18.10	17.50	18.20	18.00	18.80	17.10
Fe2O3T	2.07	6.01	9.94	9.14	9.88	7.57	9.03	9.71	9.05	9.75
MgO	0.81	1.60	6.69	4.39	4.86	3.54	4.65	4.84	3.96	5.08
CaO	2.84	5.70	11.10	9.18	9.70	7.33	9.32	9.32	7.90	8.82
Na2O	4.90	4.51	3.09	4.34	4.05	4.66	4.34	4.04	4.62	4.15
K2O	3.29	1.06	0.53	0.57	0.57	1.57	0.65	0.64	0.94	0.95
TIO2	0.21	0.44	0.73	0.86	0.92	0.70	0.86	0.81	0.92	1.27
MnO	0.06	0.09	0.15	0.15	0.16	0.13	0.15	0.15	0.14	0.16
P2O5	0.06	0.13	0.03	0.38	0.34	0.25	0.34	0.32	0.37	0.45
PAF	0.56	1.39	0.05	1.48	0.03	0.19	0.04	0.00	0.96	0.44
FeO	0.90	2.06								
Fe2O3		3.72								
TOTAL	100.80	101.53	99.51	100.79	99.71	99.44	100.18	99.73	100.56	100.17
Recalculé anhydre										
SIO2	67.84	61.91	51.08	51.35	51.26	56.42	52.53	52.04	53.11	52.14
Al2O3	17.96	18.57	16.49	19.43	18.16	17.63	18.17	18.05	18.88	17.15
Fe2O3T	2.07	6.00	9.99	9.20	9.91	7.63	9.02	9.74	9.09	9.78
MgO	0.81	1.60	6.73	4.42	4.88	3.57	4.64	4.85	3.98	5.09
CaO	2.83	5.69	11.16	9.24	9.73	7.39	9.31	9.35	7.93	8.84
Na2O	4.89	4.50	3.11	4.37	4.06	4.70	4.33	4.05	4.64	4.16
K2O	3.28	1.06	0.53	0.57	0.57	1.58	0.65	0.64	0.94	0.95
TIO2	0.21	0.44	0.73	0.87	0.92	0.71	0.86	0.81	0.92	1.27
MnO	0.06	0.09	0.15	0.15	0.16	0.13	0.15	0.15	0.14	0.16
P2O5	0.06	0.13	0.03	0.38	0.34	0.25	0.34	0.32	0.37	0.45
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

PLUTON	D2	D2	D2	D4	D4	D4	CP3	CP3	CP3
NUMERO	88019635	88019641	88019649	89016951	89016952	89016953	85017900	89016076	89016077
ESTANT	506557	500580	506859	493389	492907	492198	458150	462087	462082
NORDANT	6843386	6841505	6843761	6837458	6838306	6838740	6833000	6832954	6832953
ECHAN	MH-84-1229A	MH-84-1248B	MH-84-1228B	NT-89-19-A1	NT-89-22-A1	NT-89-23-A1	GT-85-289	CP-89-BEL-1A	CP-89-BEL-1B
LAME MINCE				19A1	22A	23A1	GT-85-289	L-BEL-1A	L-BEL-1B
LITHO	Diorite	Diorite	Diorite	Diorite qtz	Diorite qtz	Diorite qtz	Tonalite	Ton.-Grano.	Monzdior.
SIO2	58.50	48.50	53.40	58.30	52.30	65.40	64.80	70.60	51.60
Al2O3	13.30	16.30	16.40	17.30	19.90	17.90	15.70	15.80	18.40
Fe2O3T	8.92	13.10	12.90	6.61	7.72	3.21	6.74	2.00	9.52
MgO	4.74	4.20	2.43	3.40	3.42	1.12	1.19	0.67	3.97
CaO	5.59	8.05	7.60	6.14	8.22	4.15	4.35	2.64	7.41
Na2O	2.07	4.23	3.43	4.09	4.95	5.42	2.84	4.78	3.55
K2O	1.04	0.53	0.47	1.65	0.87	0.99	2.25	2.14	2.35
TIO2	0.96	1.67	1.30	0.88	0.94	0.38	0.41	0.26	1.03
MnO	0.17	0.25	0.31	0.09	0.08	0.04	0.13	0.05	0.15
P2O5	0.09	0.23	0.34	0.28	0.33	0.13	0.11	0.07	0.26
PAF	3.33	1.07	1.01	1.83	1.88	1.56	1.82	1.19	2.16
FeO									
Fe2O3									
TOTAL	98.71	98.13	99.59	100.57	100.61	100.30	100.34	100.20	100.40
Recalculé anhydre									
SIO2	61.33	49.97	54.17	59.04	52.97	66.23	65.77	71.31	52.52
Al2O3	13.94	16.79	16.64	17.52	20.16	18.13	15.94	15.96	18.73
Fe2O3T	9.35	13.50	13.09	6.69	7.82	3.25	6.84	2.02	9.69
MgO	4.97	4.33	2.47	3.44	3.46	1.13	1.21	0.68	4.04
CaO	5.86	8.29	7.71	6.22	8.33	4.20	4.42	2.67	7.54
Na2O	2.17	4.36	3.48	4.14	5.01	5.49	2.88	4.83	3.61
K2O	1.09	0.55	0.48	1.67	0.88	1.00	2.28	2.16	2.39
TIO2	1.01	1.72	1.32	0.89	0.95	0.38	0.42	0.26	1.05
MnO	0.18	0.26	0.31	0.09	0.08	0.04	0.13	0.05	0.15
P2O5	0.09	0.24	0.34	0.28	0.33	0.13	0.11	0.07	0.26
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

PLUTON	CP3	D5	D5	D8	D8	CP2	CP2	CP2	CP2
NUMERO	89016078	88019643	89016992	89016972	89016974	85017883	86024021	86024022	89016068
ESTANT	461973	489400	497507	455125	454379	458247	454696	454398	454875
NORDANT	6833656	6825985	6831630	6830648	6831345	6823347	6823710	6823323	6822142
ECHAN	CP-89-BEL-2	MH-84-235A2	NT-89-1006A	NT-89-67-B	NT-89-70-A1	GT-85-1205	CL-86-9310A3	CL-86-9124B3	CP-89-LAC-1B
LAME MINCE	L-BEL-2	235B	1006-A	67B	70-A1		47-9310A3 (SP)	43-9124 (SP)	L-ETR-1B
LITHO	Tonalite	Diorite	Diorite qtz	Diorite	Diorite	Diorite	Amphibolite	Diorite	Diorite
SiO2	66.60	46.70	59.30	49.30	56.40	42.90	45.30	45.60	49.30
Al2O3	17.10	12.60	17.60	17.80	17.30	15.40	9.19	15.80	17.30
Fe2O3T	3.23	10.00	9.37	11.50	7.77	14.70	17.90	12.00	10.70
MgO	1.10	13.60	0.45	4.29	3.42	5.64	12.90	4.53	6.62
CaO	3.60	12.40	5.20	8.43	6.27	10.80	11.70	6.61	9.66
Na2O	4.80	0.54	4.57	3.40	4.24	2.53	1.34	2.81	3.21
K2O	1.96	0.51	1.47	1.03	0.91	0.50	0.27	1.41	0.62
TiO2	0.39	0.44	0.56	1.68	1.21	3.43	1.54	1.99	1.08
MnO	0.05	0.19	0.27	0.15	0.11	0.22	0.21	0.16	0.17
P2O5	0.10	0.05	0.10	0.32	0.21	1.71	0.18	0.34	0.10
PAF	1.70	3.09	1.74	1.95	2.09	1.19	1.15	8.24	2.17
FeO			0.00		0.00				
Fe2O3			0.00		0.00				
TOTAL	100.63	100.12	100.63	99.85	99.93	99.02	101.68	99.49	100.93
Recalculé anhydre									
SiO2	67.32	48.13	59.97	50.36	57.65	43.85	45.06	49.97	49.92
Al2O3	17.28	12.99	17.80	18.18	17.68	15.74	9.14	17.32	17.52
Fe2O3T	3.26	10.31	9.48	11.75	7.94	15.03	17.81	13.15	10.83
MgO	1.11	14.02	0.46	4.38	3.50	5.77	12.83	4.96	6.70
CaO	3.64	12.78	5.26	8.61	6.41	11.04	11.64	7.24	9.78
Na2O	4.85	0.56	4.62	3.47	4.33	2.59	1.33	3.08	3.25
K2O	1.98	0.53	1.49	1.05	0.93	0.51	0.27	1.55	0.63
TiO2	0.39	0.45	0.57	1.72	1.24	3.51	1.53	2.18	1.09
MnO	0.05	0.20	0.27	0.15	0.11	0.22	0.21	0.18	0.17
P2O5	0.10	0.05	0.10	0.33	0.21	1.75	0.18	0.37	0.10
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

PLUTON	CP2	CP2	CP2	DL1	DL1	DL1	DL1	DL1	DL1
NUMERO	89016069	89016070	89016067	87022920	87022927	87022928	87024405	87024416	87024419
ESTANT	454758	453611	454878	381864	382004	382016	388788	385376	383782
NORDANT	6823462	6822649	6822139	6811074	6810491	6810933	6817843	6816664	6812993
ECHAN	CP-89-LAC-2	CP-89-LAC-3	CP-89-LAC-1	PB-87-3374A	PB-87-3379A	PB-87-3375A	PB-87-2075	PB-87-409	PB-87-3040
LAME MINCE	L-ETR-2	L-ETR-3	L-ETR-1	PB-3374A-87	PB-3379A-87	PB-3375A-87	PB-2075-3A-87		PB-3040
LITHO	Diorite	Dior.-Monzdior.	Diorite	Diorite	Diorite	Diorite	Granodiorite	Granodiorite	Granodiorite
SIO2	49.70	51.80	48.10	54.60	44.30	46.60	70.30	70.30	72.50
Al2O3	16.90	17.00	17.00	16.80	18.50	12.40	15.70	15.80	16.10
Fe2O3T	10.50	11.00	14.10	7.21	8.42	9.42	1.99	2.57	1.51
MgO	5.06	4.80	4.12	3.38	10.00	12.70	0.72	0.92	0.45
CaO	10.20	6.80	8.81	9.48	10.80	13.00	2.47	2.98	2.67
Na2O	3.33	4.06	3.59	4.74	1.94	0.99	5.12	4.83	5.40
K2O	0.80	1.17	0.34	0.82	0.20	0.22	2.09	1.66	1.55
TIO2	1.13	1.30	2.05	1.09	0.39	0.42	0.28	0.30	0.20
MnO	0.18	0.15	0.22	0.09	0.13	0.16	0.03	0.04	0.03
P2O5	0.53	0.11	0.40	0.22	0.04	0.03	0.07	0.09	0.06
PAF	1.81	2.16	0.17	0.66	3.76	3.24	0.66	0.45	0.56
FeO									
Fe2O3									
TOTAL	100.14	100.35	98.90	99.09	98.48	99.18	99.43	99.94	101.03
Recalculé anhydre									
SIO2	50.54	52.75	48.72	55.47	46.77	48.57	71.18	70.66	72.16
Al2O3	17.19	17.31	17.22	17.07	19.53	12.92	15.90	15.88	16.02
Fe2O3T	10.68	11.20	14.28	7.33	8.89	9.82	2.01	2.58	1.50
MgO	5.15	4.89	4.17	3.43	10.56	13.24	0.73	0.92	0.45
CaO	10.37	6.93	8.92	9.63	11.40	13.55	2.50	3.00	2.66
Na2O	3.39	4.13	3.64	4.82	2.05	1.03	5.18	4.85	5.37
K2O	0.81	1.19	0.34	0.83	0.21	0.23	2.12	1.67	1.54
TIO2	1.15	1.32	2.08	1.11	0.41	0.44	0.28	0.30	0.20
MnO	0.18	0.15	0.22	0.09	0.14	0.17	0.03	0.04	0.03
P2O5	0.54	0.11	0.41	0.22	0.04	0.03	0.07	0.09	0.06
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

PLUTON	DL1	DL1	DL1	DL1	DL1	DL1
NUMERO	87024420	87024425	87024426	87024430	87024434	87024443
ESTANT	382638	382301	385003	387326	387427	387522
NORDANT	6809039	6815328	6811733	6817239	6811548	6810519
ECHAN	PB-87-3493	PB-87-415	PB-87-3044	PB-87-3412	PB-87-3500	PB-87-3499
LAME MINCE			PB-3044-87			PB-3499
LITHO	Granodiorite	Granodiorite	Granodiorite	Granodiorite	Granodiorite	Granodiorite
SIO2	73.80	71.40	67.20	71.20	73.00	71.90
Al2O3	14.10	15.00	18.90	15.80	15.70	15.70
Fe2O3T	2.53	1.66	1.63	2.19	1.27	1.43
MgO	0.95	0.66	0.44	0.75	0.53	0.52
CaO	2.84	2.54	3.34	2.70	1.59	1.84
Na2O	4.19	4.92	6.65	4.89	5.25	4.47
K2O	1.21	1.91	1.40	2.08	2.06	4.16
TIO2	0.31	0.23	0.23	0.27	0.19	0.17
MnO	0.04	0.03	0.03	0.04	0.03	0.03
P2O5	0.09	0.07	0.07	0.08	0.05	0.05
PAF	0.53	0.78	0.66	0.40	1.03	0.44
FeO						
Fe2O3						
TOTAL	100.59	99.20	100.55	100.40	100.70	100.71
Recalculé anhydre						
SIO2	73.76	72.55	67.27	71.20	73.24	71.71
Al2O3	14.09	15.24	18.92	15.80	15.75	15.66
Fe2O3T	2.53	1.69	1.63	2.19	1.27	1.43
MgO	0.95	0.67	0.44	0.75	0.53	0.52
CaO	2.84	2.58	3.34	2.70	1.60	1.84
Na2O	4.19	5.00	6.66	4.89	5.27	4.46
K2O	1.21	1.94	1.40	2.08	2.07	4.15
TiO2	0.31	0.23	0.23	0.27	0.19	0.17
MnO	0.04	0.03	0.03	0.04	0.03	0.03
P2O5	0.09	0.07	0.07	0.08	0.05	0.05
TOTAL	100.00	100.00	100.00	100.00	100.00	100.00

ANNEXE 2

Analyses d'éléments trace et de terre rares des plutons de la Fosse de l'Ungava.

PLUTON NUMERO LITHO	D3 84015706 Monzdior.qtz	D3 84019556 Monzdior.qtz	CP5 89016071 Diorite	CP5 89016072 Diorite	CP5 89016058 Diorite	CP5 89016059 Monzdior.	CP5 89016060 Monzdior.	CP5 89016061 Monzdior.	CP5 89016073 Monzdior.	CP5 89016075 Monzdior.	D2 88019635 Diorite
Rb	79	25	6			31			5	11	19
Sr	871	1100	660	1000	1000	800	1000	980	1000	970	190
Ba	2700	806	304	403	464	581	481	443	600	422	386
Th					0.3	1.8	0.2	0.2			1.8
U						1.1		0.5			1.3
Sc					24	19	24	22			27
Ta											0.9
Nb						4					9
Hf					0.7	2.0	0.7	0.7			2.5
Y	8	13	20	15	18	19	16	16	18	23	19
Zr	110	190	54	42	44	72	45	45	64	150	94
Cr	2	1200	120	91	53	56	82	107	68	132	
Cu	2	24	126	93	184	100	167	153	104	134	27
Ni	2		47	25	38	28	30	35	22	37	58
Pb	19										
Zn	38	69	98	84	114	91	103	106	92	98	87
Co	5	9	35	23	30	22	24	29	19	26	30
Ga			21	23	25	21	22	20	23		17
Sc	3	8	34	18	23	13	24	22	19		31
La	15	28	9	13	15	15	18	15	16		11
Ce	27	52	17	25	30	31	30	28	33	34	19
Li	20	15	16	14	6	8	6	7	18	16	11
V	26	59	237	217	260	159	222	230	196	215	222
C					0.4	0.7					0.3
La					16	20	17	17			9.5
Ce					34	40	35	33			23
Nd					19	21	22	18			12
Sm					4.8	4.8	4.8	4.7			3.2
Eu					1.5	1.4	1.5	1.4			1.0
Tb					0.5	0.6	0.6	0.5			0.5
Ho					1.1	1.2	0.9	0.3			0.9
Tm											0.7
Yb					1.5	1.9	1.5	1.6			2.1
Lu					0.24	0.27	0.24	0.27			0.30
Rb/Sr	0.091	0.023	0.009			0.039			0.005	0.011	0.100
K/Rb	345	351	737			424			1567	719	476
(La/Yb)n					7.65	7.55	8.13	7.62			3.24

PLUTON NUMERO LITHO	D2 88019641 Diorite	D2 88019649 Diorite	D4 89016951 Diorite qtz	D4 89016952 Diorite qtz	D4 89016953 Diorite qtz	CP3 85017900 Tonalite	CP3 89016076 Ton.-Grano.	CP3 89016077 Monzdior.	CP3 89016078 Tonalite	D5 88019643 Diorite
Rb	10	11	29	11	19	77	50	73	43	15
Sr	470	400	900	1300	940	320	550	570	730	220
Ba	116	187	633	264	343	632	851	399	669	252
Th	0.7	1.2	2.9	0.6	1.1			1.3	3.4	2.4
U	<.5	1.1	2.1	<.5	0.9			0.8	1.2	<.5
Sc	29	30	15	16	5.1			19	5.3	34
Ta	1.0	3.5	2.7	1.6	3.6					0.5
Nb	10	10	6				6	6		6
Hf	2.9	1.4	2.5	2.0	2.7			2.5	3.0	1.3
Y	30	28	11	8	6	15	8	24	6	18
Zr	110	53	110	93	140	120	100	120	130	47
Cr						2	36	40	39	
Cu	80	20	39	63	8	7	6	36	11	70
Ni	57	4	38	39	9		6	11	12	204
Pb						32	24		14	
Zn	88	146	82	77	47	64	53	105	99	80
Co	28	18	18	25	5	7		23	6	40
Ga	18	20	20	25	20	15	20	21	21	13
Sc	30	30	19	21	4	12	2	11	4	38
La	12	10	31	17	7	35	8	8	10	9
Ce	20	16	46	25	5	54	17	20	20	12
Li	10	6	32	9	3	16	11	20	13	15
V	296	76	141	168	49	45	25	203	40	199
C	0.3	<.2	0.7	0.3	0.3			1.4	0.5	<.2
La	12	11	24	12	10			14	18	8.3
Ce	29	24	48	29	19			30	31	15
Nd	19	18	28	20	9			20	14	6.3
Sm	5.1	5.0	5.4	4.4	2.1			4.7	2.6	2.0
Eu	1.6	1.9	1.4	1.4	0.7			1.4	0.8	0.7
Tb	1.0	0.8	0.5	0.5	0.2			0.6	0.2	0.4
Ho	1.1	1.2	0.7	0.9	0.3			1.7	0.2	<.5
Tm	1.2	1.2	<.2	<.2	<.2			<.2	<.2	0.6
Yb	3.4	3.1	1.2	1.0	0.5			1.8	0.4	1.7
Lu	0.48	0.45	0.15	0.14	0.12			0.30	0.07	0.26
Rb/Sr	0.021	0.028	0.032	0.008	0.020	0.241	0.091	0.128	0.059	0.068
K/Rb	453	360	478	665	438	246	359	272	382	291
(La/Yb)n	2.53	2.55	14.35	8.61	14.35			5.58	32.28	3.50

PLUTON	D5	D8	D8	CP2	CP2	CP2	CP2	CP2	CP2	CP2	DL1
NUMERO	89016992	89016972	89016974	85017883	86024021	86024022	89016068	89016069	89016070	89016067	87022920
LITHO	Diorite qtz	Diorite	Diorite	Diorite	Amphibolite	Diorite	Diorite	Diorite	Dior.-Monzdior.	Diorite	Diorite
Rb	42	25	23	14	5	49	12	14	35	6	21
Sr	460	460	430	390	280	270	270	680	290	490	330
Ba	322	292	262	166	100	313	161	334	263	155	225
Th	3.2	2.8	2.4							0.7	
U	0.0	<.5	1.1							<.5	
Sc	25	24	17							28	
Ta											
Nb	9	11	9		8	14	5	9	12	10	10
Hf	5.1	4.3	3.5							1.8	
Y	33	35	22	33	20	39	22	38	32	28	27
Zr	210	200	150	75	67	100	89	130	150	95	120
Cr	0		0	2			205	143	56	44	
Cu	11	133	496	38	185	41	50	33	54	53	6
Ni	0	21	19		317		62	40	19	5	34
Pb	0		0								1
Zn	127	82	43	86	127	113	101	108	123	143	38
Co	0	30	27	45	68	34	39	29	37	26	17
Ga	24	23	22	28	12	18	20	22	23	25	16
Sc		27	20	33	51	41	14	22	18	24	30
La	16	32	17	19	11	23	5	27	13	13	15
Ce	29	62	33	43	23	53	12	67	28	28	33
Li	6	14	11	8	2	37	7	11	18	6	8
V	0	252	159	239	260	293	207	207	257	223	174
C	0.8	0.8	0.3							<.2	
La	17	27	16							14	
Ce	35	66	31							30	
Nd	18	34	15							18	
Sm	4.7	7.2	4.3							5.4	
Eu	2.4	2.2	1.0							1.9	
Tb	0.7	1.1	0.6							0.7	
Ho	1.0	1.5	1.0							1.6	
Tm	0.0	<.2	0.0							<.2	
Yb	3.1	2.7	1.4							2.0	
Lu	0.52	0.46	0.25							0.36	
Rb/Sr	0.091	0.054	0.053	0.036	0.018	0.181	0.044	0.021	0.121	0.012	0.064
K/Rb	294	349	336	303	446	262	434	482	283	476	329
(La/Yb)n	3.93	7.17	8.20							5.02	

PLUTON NUMERO LITHO	DL1 87022927 Diorite	DL1 87022928 Diorite	DL1 87024405 Granodior.	DL1 87024416 Granodior.	DL1 87024419 Granodior.	DL1 87024420 Granodior.	DL1 87024425 Granodior.	DL1 87024426 Granodior.	DL1 87024430 Granodior.	DL1 87024434 Granodior.	DL1 87024443 Granodior.
Rb	3	3	49	39	40	27	23	31	47	59	84
Sr	200	150	620	580	750	470	560	950	560	570	650
Ba	71	49	936	671	725	494	960	838	914	681	1300
Th											
U											
Sc											
Ta											
Nb	6	5	9	8	9	11	9	9	10	11	7
Hf											
Y	9	11					4			4	
Zr	39	37	120	130	130	120	120	150	120	110	91
Cr			150								
Cu	117	131	2	4	2	4	2				5
Ni	220	167	4	8	4	9	5		5		7
Pb	1	1	20	18	24	17	19	24	21	19	30
Zn	69	91	46	54	45	53	49	37	51	45	40
Co	54	51	5	6		5	7		5		
Ga	14	9	16	18	17	12	17	19	15	17	16
Sc	18	45	3	6	2	3	5	2	4	2	3
La	14	5	20	15	7	26	24	17	17	16	9
Ce	7	10	29	23	13	40	37	31	27	21	16
Li	15	11	38	40	29	28	6	27	33	34	38
V	76	132	22	31	14	26	28	8	27	11	14
C											
La											
Ce											
Nd											
Sm											
Eu											
Tb											
Ho											
Tm											
Yb											
Lu											
Rb/Sr	0.015	0.020	0.079	0.067	0.053	0.057	0.041	0.033	0.084	0.104	0.129
K/Rb	584	635	358	355	320	372	700	375	367	291	410
(La/Yb)n											

GENERAL DISCUSSION AND CONCLUSIONS

Evaluation of the geochemical evolution of the Earth involves the study and examination of various crustal compositions and mantle reservoirs throughout geological time. Such studies often concentrate on the geological evolution of a suite of rocks of a particular age and compare and contrast their characteristics with rocks of other ages. In this present study we have focused on rocks of Early Proterozoic age as the interval 1.8-2.0 Ga represented an important period of global continental growth and accretion. The papers presented in this thesis have documented the magmatic evolution and crustal accretion within the Early Proterozoic Ungava Orogen of northern Quebec.

5.1 Magmatic evolution and crustal composition of the Narsajuaq terrane

Characterization of the composition of the continental crust is one important means of contributing to the study of the geochemical evolution of the Earth and this has been addressed in the first paper of the thesis (Chapter 1) where the geology and geochemistry of the Narsajuaq terrane is described. The Narsajuaq terrane represents the roots of an Early Proterozoic magmatic arc complex that was accreted to the margin of the Superior Province ca. 1.82 Ga. The terrane consists of four different plutonic suites spanning more than 150 Ma of magmatic activity. The Older suite consists of a well layered sequence of diorite-tonalite, intruded and interlayered with variably deformed and metamorphosed granitic veins and ranges in age from 1863 to 1844 Ma. Intrusive into the Older suite is the Younger suite, consisting of discrete, km-size bodies of metamorphosed plutons (diorite - monzodiorite - tonalite - granite) that range in age from 1836-1800 Ma. The Late suite contains undeformed and unmetamorphosed granitic plutons and syenogranitic pegmatites that cross-cut all elements of the Ungava Orogen and have been dated at 1758-1742 Ma. Finally, a series of deformed and undeformed plutons that are intrusive into the Cape Smith Belt lithologies and range in age from 1898-1839 Ma, make up the Cape Smith suite. In addition, a series of metamorphosed sedimentary rocks (predominantly semi-pelites) of the Sugluk Group are also found in the Narsajuaq terrane, and contain zircons of both Archean and Early Proterozoic age.

The various plutonic suites of the Narsajuaq terrane are predominately calc-alkaline and metaluminous to mildly peraluminous in terms of major element geochemistry and contain variable but generally moderate to high concentrations of the LILE's and light REE's, pronounced negative Nb and Ti anomalies, and fractionated REE's. A range in isotopic compositions is noted for the suites, with the Older, Younger and Late suites having initial ε_{Nd} and $^{87}Sr/^{86}Sr_{(i)}$ values ranging from juvenile (+4.0 and 0.7020, respectively) to very enriched (-18.5 and 0.7064), indicating that at least three sources were involved in their petrogenesis: depleted mantle, average 2.7 Ga Superior Province crust and old (3.1 Ga) Superior Province crust. The geochemical and isotopic signature of most of the Older, Younger and Cape Smith suite plutons can be derived by fractionation and variable contamination of a slightly LILE- and light REE-enriched mantle source; such enrichment was produced by subduction-related processes involving slab dehydration and the release of the LILE's and light REE's into the overlying mantle wedge. A distinct series of monzodiorite plutons from the Younger suite contain very high concentrations of the LILE's and light REE's and were derived by partial melting of a highly enriched mantle source. The large granitic sheets of the Younger suite and the Late suite granites and syenogranites have geochemical and isotopic compositions indicative of significant crustal input and were most likely generated by partial melting of the crust (both Archean and Proterozoic) due to tectonic thickening and metamorphic hydration of the crust at this time. A detailed geochronological study of one Late suite pluton (the Lac Duquet pluton) using a laser ablation ICP-MS technique (Chapter 3) identified inherited zircons in the pluton which have a range in ages from 1.7 to 3.2 Ga, with more than 80% of the ages corresponding to the ages of the surrounding host rocks for the granite.

5.1.1 *Crustal section*

A composite crustal section through the Narsajuaq terrane has been constructed (presented in Chapter 1), representing approximately 15-25 km of section. The plutonic and sedimentary rocks of the Narsajuaq terrane represent a mid-crustal section through an Early Proterozoic magmatic arc. The uppermost part of the crustal section is missing although the volcanic rocks of the Parent Group, which have previously been interpreted as arc deposits (Picard et al. 1990) may represent such crust. Metasediments of the Sugluk Group are more prevalent near the top of the crustal section, as are large granitic sheets, whereas mafic (diorite) plutons are more common deeper in the section.

Significant decreases in the $(La/Yb)_n$ ratios and overall increase in the heavy REE contents have been noted in post-Archean rocks (e.g., Taylor and McLennan 1985; Martin 1986) which are borne out by our own data. These changes are believed to have occurred in response to the change in magma-generating processes from slab melting to wedge melting in subduction zones due to the secular cooling of the Earth. The relative amount of fractionation of the REE's for intermediate compositions was noted to be similar for the Narsajuaq terrane samples and samples from Phanerozoic magmatic arc suites (see earlier section on comparison with magmatic arcs and Fig. 33c) which suggests that the conditions for melt generation in the Proterozoic were the same as for the younger Phanerozoic samples. The notable geochemical differences (particularly for Th and U) between the suites must therefore be related to another process and it is suggested that variations in source composition is the most viable means of accounting for these differences.

5.2 Mantle reservoirs

An evaluation of the various mantle reservoirs that were present during the formation of the rocks of the Ungava Orogen at approximately 2.0 Ga was the focus of another paper (Chapter 2). Voluminous magmatism associated with rifting of the Superior Province basement at ≈ 2.04 Ga resulted in the development of a volcanic rift margin sequence and an ocean basin. Four distinct mafic magma suites were erupted over the course of approximately 100 Ma, related to the presence of various mantle and crustal reservoirs in the region at this time, namely depleted mantle, enriched mantle and continental crust. Suites associated with continental rifting and the development of a volcanic margin include: (1) the crustally contaminated continental basalts of the Eskimo Formation and the western and central Povungnituk Group; (2) the mafic lavas of the

Flaherty Formation, eastern Povungnituk Group and some Watts Group samples which have slightly enriched geochemical and isotopic signatures; and (3) the highly enriched alkaline lavas of the Povungnituk Group. During the development of the oceanic basin a fourth suite was erupted consisting of Mg-rich basalts and komatiitic basalts of the Chukotat Group, Ottawa Islands and some Watts Group samples, which have geochemical and isotopic compositions similar to N-MORB.

The spectrum of magma compositions erupted during the rifting event and generation of the oceanic basin have provided us with a window into the mantle at 2.0 Ga and permitted us to define mantle reservoir compositions for depleted mantle ($\varepsilon_{Nd} = +4.5$ to $+5.5$) and enriched mantle ($\varepsilon_{Nd} = +2.5$ to $+3.5$). Continental rifting may have been associated with, or contemporaneous with, the arrival of a mantle plume which generated the Povungnituk Group, as well as the Eskimo and Flaherty Formations. Melting within the plume could have produced the enriched mantle signature of these rocks with crustal contamination superimposed upon the Eskimo Formation and western Povungnituk Group samples. Complete rupture of the crust resulted in the generation of an oceanic basin into which the Watts Group, Chukotat Group and Ottawa Islands lavas were deposited. Most of these samples were derived from the depleted mantle reservoir although the Watts Group contains lavas having both depleted and enriched characteristics indicating the synchronous existence and tapping of these reservoirs.

An additional reservoir that served as the source for the subsequent arc magmatism was also defined based on the geochemical and isotopic composition of the most primitive arc magmas from the Narsajuaq terrane and has a similar Nd isotopic composition as the depleted mantle reservoir. The more detailed studies of the Narsajuaq terrane magmas presented in Chapter 1 suggest that an enriched mantle reservoir was also sampled by arc magmas, particularly for the Younger suite monzodiorites. However it is postulated that this enrichment was of a more local nature, directly related to the subduction processes active at that time.

The chemical and isotopic similarity of these Proterozoic magmas to modern-day magmas provide strong evidence that the interplay between depleted mantle, OIB mantle and sub-continental mantle during the Proterozoic was comparable to that of the modern Earth, indicating a continuity in magmatic processes through time. The presence of these mantle reservoirs at 2.0 Ga also indicates that the geochemical and tectonic processes that generate such reservoirs were already in existence at this time, which suggests that crust-mantle recycling was active prior to 2.0 Ga.

5.3 Crustal accretion and regional geotectonic setting

The origin of the Narsajuaq terrane is related to the overall tectonic evolution of the northern Ungava Peninsula during the Early Proterozoic. This region was involved in the development of the supercontinent of Laurentia (Fig. 1; Hoffman 1988; Van Kranendonk et al. 1993) which consisted of several Archean cratons (i.e., Superior, Nain and Rae provinces) welded together by Early Proterozoic orogenic belts such as the Ungava, New Quebec and Torngat orogens. A recent synthesis of all geological and geochronological data for the northeastern region of Laurentia by Van Kranendonk et al. (1993) presents a superb overview of the regional-scale tectonics.

Continental rifting along the present day margins of the Superior Province occurred from approximately 2.17-2.03 Ga as evidenced by unconformable sedimentary passive margin sequences overlying the Superior Province basement. In the Ungava Orogen the rifting produced the Povungnituk Group and Eskimo and Flaherty Formations and eventually resulted in the development of an oceanic basin into which the Chukotat and Watts Groups were deposited. Other sequences related to Superior Province rifting were deposited in the New Quebec Orogen (Seward Group) and along the east coast of Hudson Bay (Richmond Gulf Group) while sedimentary deposits interpreted as related to rifting of the Rae Province are found in the Lake Harbour Group on Baffin Island (Jackson and Taylor 1972; Chandler and Parrish 1989; Wardle et al. 1990; Skulski et al. 1993; Van Kranendonk et al. 1993).

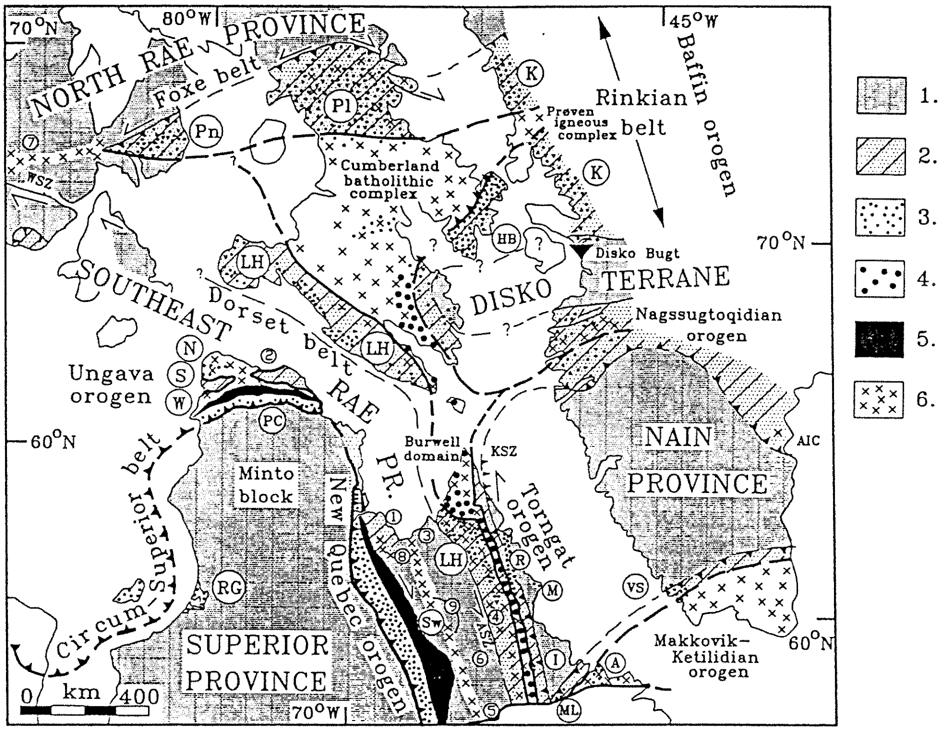


Figure. 5.1. Geological compilation map of northeastern Laurentia showing distribution of Archean cratons (1) and Proterozoic orogens (2-6). From Van Kranendonk et al. (1993).

The oldest sequences recording the onset of crustal compression and destruction in northeastern Laurentia are the apparently magmatic arc rocks of the Cumberland batholith, dated at ≈ 1950-1870 Ma, which are postulated to have formed by westward-directed subduction beneath the thinned eastern margin of the southeast Rae Province (Hoffman 1990; Jackson et al. 1990; Van Kranendonk et al. 1993). Destruction of the Ungava oceanic basin began ca. 1.90 Ga with the intrusion of arc-related plutons into the Cape Smith Belt (1898-1836 Ma Cape Smith suite) and was followed by (and partly synchronous with) the Older suite of plutonic activity. The development of the magmatic arc was related to northward subduction of the oceanic crust under a crustal fragment, possibly a part of the previously rifted Superior Province margin. A flip in the subduction direction between 1844-1836 Ma due to the attempted subduction of the more buoyant continental crustal margin attached to the subducting oceanic plate resulted in the generation of the Younger suite.

Voluminous mafic magmatism in the New Quebec Orogen due to the development of foredeep successions or pull-apart basins (Hoffman 1987; Skulski et al. 1993) was synchronous with the subduction magmatism in the Ungava Orogen. Younger suite magmatism in the Ungava Orogen was contemporaneous with accretion of Proterozoic terranes in the northern part of the New Quebec Orogen, emplacement of the Kuujjuaq batholith and collision of the southeast Rae and Superior Provinces (Machado et al. 1989; Perrault and Hynes 1990) while arc magmatism at 1840-1811 Ma related to east-directed subduction under the Rae Province generated the De Pas batholith (Krogh 1986; Martelain 1989; van der Leeden et al. 1990). Subsequent west-directed thrusting and retrograde metamorphism terminated at 1.79-1.77 Ga indicating final collision of the southeast Rae and Superior Provinces (Machado et al. 1989).

Collision and accretion of the Narsajuaq terrane to the Superior Province craton occurred between 1821 Ma (the age of the youngest arc-related pluton) and 1803 Ma (the age of the syn-collisional granitic sheets derived by crustal melting), although early deformation related to the accretion is recorded at 1870 Ma in the Cape Smith Belt.

Continent-continent collision between the Superior Province (and the accreted Cape Smith Belt and Narsajuaq terrane) and the southeast Rae Province occurred after 1803 Ma but before 1758 Ma, the age of the oldest post-tectonic Late suite pegmatite dykes that cross-cut all elements of the Ungava Orogen.

5.4 Future work and outstanding problems

While this present study has answered many questions and shed much light on the origin and evolution of the rocks of the northern Ungava Peninsula, several problems still remain to be solved. The geological and geochemical study of the Narsajuaq terrane has documented that some important differences exist between the composition of these Early Proterozoic rocks and younger, Phanerozoic magmatic arc rocks (e.g., lower Th and U) although these differences are not yet understood. Future and ongoing work consists of determining a geochemical composite composition for the Narsajuaq terrane and comparing it with estimates of the bulk composition of crust from other sections of Archean and Phanerozoic age in order to evaluate the geochemical changes in crustal composition, particularly during the Precambrian, at the Archean-Proterozoic boundary.

In terms of regional geology, one important unknown is the significance, age, origin and composition of the volcanic and volcaniclastic units of the Parent Group. Preliminary studies (Picard et al. 1990; St-Onge and Lucas 1992; Barrette 1994) have documented the general arc-affinity of these rocks but few geochemical analyses or petrological studies are available. The Parent Group may represent the extrusive equivalent of the plutonic units of the Narsajuaq terrane, and the plutons of Cape Smith suite that are found within the Parent Group may represent the synvolcanic reservoirs for the overlying Parent Group lavas. However, these correlations need to be tested through detailed study of the Parent Group.

Another important area of investigation is related to the relationship of the Narsajuaq terrane units to rocks of southeastern Baffin Island. Investigation of these rocks could further enhance our understanding of the origin and evolution of the

Narsajuaq terrane rocks and may define where the northern margin of the terrane lies, whether it was built on a crustal fragment of a relatively restricted size and perhaps identify intrusive contacts between the terrane and its basement. Preliminary work by Scott and Godin (1995) in the Lake Harbour region of southeast Baffin Island has identified a series of supracrustal rocks interpreted as reworked Archean units (Jackson and Taylor 1972; Hoffman 1988) that were intruded by migmatitic gneiss at approximately 1840-1830 Ma, synchronous with the Younger suite plutonism in the Narsajuaq terrane, suggesting a relationship between the two events (Scott and Godin 1995). In addition, preliminary geochronology indicates that most magmatic and tectonic activity in this region terminated at ≈ 1760 Ma (Scott and Godin 1995), similar to the 1758 Ma age for terminal deformation in the Ungava Orogen. This region of Baffin Island is now the focus of a new Geological Survey of Canada project which will further investigate the overall geologic and tectonic relationship of this region to the Narsajuaq terrane and the Ungava Orogen.

REFERENCES

Barrette, P.D. 1994. Lithostratigraphy and map-scale structure in the western Cape Smith Belt, northern Quebec: a tentative correlation between two tectonic domains. Canadian Journal of Earth Sciences, **31**:986-994.

Chandler, F.W., and Parrish, R.R. 1989. Age of the Richmond Gulf Group and implications for rifting in the Trans-Hudson Orogen, Canada. Precambrian Research, **44**:277-288.

Hoffman, P.F. 1987. Early proterozoic foredeeps, foredeep magmatism, and Superior-type iron-formations of the Canadian Shield. *In*: Proterozoic Lithospheric Evolution. *Edited by*: A. Kroner. American Geophysical Union, Geodynamic Series 17, pp. 85-98.

Hoffman, P.F. 1988. United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annual Reviews of Earth and Planetary Science, **16**:543-603.

Hoffman, P.F. 1990. Dynamics of the tectonic assembly of northeast Laurentia in geon 18 (1.9-1.8 Ga). Geoscience Canada, **17**:222-226.

Jackson, G., and Taylor, F.C. 1972. Correlation of major Aphebian rock units in the northeastern Canadian Shield. Canadian Journal of Earth Sciences, **9**:1650-1669.

Jackson, G.D., Hunt, P.A., Loveridge, W.D., and Parrish, R.R. 1990. Reconnaissance geochronology of Baffin Island, N.W.T. *In*: Radiogenic Age and Isotopic Studies, Geological Survey of Canada Paper 89-2, pp. 123-148.

Krogh, T.E. 1986. Report to the Newfoundland Department of Mines and Energy on isotopic dating results from the 1985-1986 geological research agreement. Newfoundland Department of Mines and Energy, Open File Report, LAB 707, pp. 15.

Machado, N., Goulet, N., and Gariépy, C. 1989. U-Pb geochronology of reactivated Archean basement and of Hudsonian metamorphism in the northern Labrador Trough. Canadian Journal of Earth Sciences, **26**:1-15.

Martelain, J. 1989. Le batholith de De Pas dans le segment oriental de l'orogene Trans-Hudsonien: Typologie et implications geodynamiques. PhD thesis, Université Claude Bernard, Lyon, France.

Martin, H. 1986. Effect of steeper Archean geothermal gradient on geochemistry of subduction-zone magmas. Geology, **14**:753-756.

Perreault, S., and Hynes, A. 1990. Tectonic Evolution of the Kuujjuaq Terane, New Quebec Orogen. Geoscience Canada, **17**:238-241.

Picard, C., Lamothe, D., Piboule, M., and Oliver, R. 1990. Magmatic and geotectonic evolution of a Proterozoic oceanic basin system: the Cape Smith Thrust-Fold Belt (New-Quebec). Precambrian Research, **47**:223-249.

Scott, D.J., and Godin, L. 1995. Preliminary geological investigation of the Lake Harbour Group and surrounding gneissic rocks near Lake Harbour and Markham Bay, souther Baffin Island, Northwest Territories. In Geological Survey of Canada, Current Research, Part C pp. 67-76.

Skulski, T., Wares, R.P., and Smith, A.D. 1993. Early Proterozoic (1.88-1.87 Ga) tholeiitic magmatism in the New Quebec orogen. Canadian Journal of Earth

Sciences, **30**:1505-1520.

St-Onge, M.R., and Lucas, S.B. 1992. New insight on the crustal structure and tectonic history of the Ungava orogen, Kovik Bay and Cap Wolstenholme, Quebec. In Current Research, Part C. Geological Survey of Canada, Paper 92-1C, pp. 31-41.

St-Onge, M.R., Lucas, S.B., and Parrish, R.R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Quebec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. Canadian Journal of Earth Science, **29**:746-764.

Taylor, S.R., and McLennan, S.M. 1985. The continental crust: its composition and evolution. Blackwell Scientific Publications, Boston, pp. 312.

van der Leeden, J., Belanger, M., Danis, D., Girard, R., and Martelain, J. 1990. Lithotectonic domains in the high-grade terrain east of the Labrador Trough (Quebec). *In*: The Early Proterozoic Trans-Hudson orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Survey of Canada, Special Paper 37, pp. 371-386.

Van Kranendonk, M.J., St-Onge, M.R., and Henderson, J.R. 1993. Paleoproterozoic tectonic assembly of Northeast Laurentia through multiple indentations. Precambrian Research, **63**:325-347.

Wardle, R.J., Ryan, B., Nunn, G.A.G., and Mengel, F.C. 1990. Labrador segment of the Trans-Hudson orogen: Crustal development through oblique convergence and collision. *In*: The Early Proterozoic Trans-Hudson Orogen of North America. *Edited by* J.F. Lewry and M.R. Stauffer. Geological Association of Canada, Special Paper 37, pp. 353-369.

APPENDIX 1

Summary descriptions of Narsajuaq terrane thin sections.

Sample	t.s. name	qtz	plag	kspar	opx	cpx	hb	epid	trem	chl	bio	mu	ap	titan	opq	gt	other/comments
Older Suite																	
D-41c	diorite	tr	40				59			tr	tr			1			
D-64	tonalite	20	20				35				23				1	1	gt porphyroblast
L-20b	qtz diorite	10	40				30	7			10		tr	3			ep often with allan or titan cores
L-90a	diorite		50				25				25		tr	tr			
L-90b	diorite	tr	45				40				15		tr	tr			
L-174	qtz mzdiorite	10	30	10			50			tr		tr				tr	plag lots ser; gt as one huge porphyroblast
L-211	diorite		10			20	70		tr								
S-18a	tonalite	10	35			15	15				24		tr		1		
M-34	qtz diorite	5	70				9				15			tr		1	
M-40	tonalite	20	28				35		tr		10				2	5	
M-103	diorite	1	48	tr		tr	25				25		1	tr	tr		
D-72	tonalite	40	35								25	tr		tr			
D-163	tonalite	20	25				25	tr			30		tr	tr	tr		
D-211	granodiorite	15	20	4			20	tr			40		1		tr		
L-1	syenogranite	15	10	30			5	2			35		1	2	tr		
L-175a	tonalite	25	60				1				14						thin section at contact between felsic-mafic
L-175a	tonalite	20	20				29				30					1	with geochemistry on felsic portion
M-89	tonalite	35	30				tr			tr	34	tr				1	few large gt porphyroblasts
D-14a	granodiorite	50	32	15						tr	3				tr		tr cc
D-23	granite	20	40	30						tr	10	tr					
D-29	tonalite	50	49							tr	1	tr					
D-41a	syenogranite	48	10	40						tr	1				1		
L-27	granodiorite	50	30	5				1			14	tr	tr		tr		
K-67	qtz granitoid	55	35								10		tr	tr	tr		
Younger Suite																	
D-26	diorite		30					69			tr			1			plag lots ser; hb to ep
D-28	diorite		20					79						1			hb to ep
D-165	diorite		30				53	tr			15		tr		2	tr	
L-18	diorite	tr	20			1	45	tr			31		1	2			
L-225	diorite		20			2	68				10		tr				hb with lots op (oxyhb?)
S-34	qtz diorite	5	45				39	tr			10			tr	1		hb to variable ep; lots distinct ep, mu xtals on plag
S-39a	diorite		25				70				5				tr		
S-106	diorite		20			5	65	tr			10				tr		hb to variable ep
K-69	qtz diorite	1	20				75				2			tr	2		
D-20	mzdiorite	tr	45	10			30	1			10		1	3			3 allan with ep rims
L-43	mzdiorite		40	20			30	tr			9		tr	1	tr		
S-12	qtz syenite	5	10	25			50				10		tr	tr	tr		

Sample	t.s. name	qtz	plag	kspar	opx	cpx	hb	epid	trem	chl	bio	mu	ap	titan	opq	gt	other/comments
E-4	monzonite		45	25		5	23				1		tr	1	tr		cpx with exsolved Fe-Ti oxides
D-24	qtz mzdiorite	10	35	10			40				4		tr	1			
D-43	monzonite	tr	25	25		25	10				15		tr	tr	tr		cpx with exsolved Fe-Ti oxides
L-38a	syenite		15	35		32	2				15		tr	tr	1		cpx with exsolved Fe-Ti oxides
L-39	k syenite		5	50		39	tr	1			5		tr				cpx with exsolved Fe-Ti oxides
S-19	qtz diorite	5	50	5		1	23	5			10			1			
L-118	qtz syenite	15	15	60				1			8				1		
D-220	syenogranite	40	5	50							tr			tr	5		
D-222	qtz granitoid	60	25	5							9	tr			tr	1	
D-283	granite	20	30	40							5		tr		5		
L-205	qtz k syenite	10	1	85				tr			3	tr		tr	1		1 allan
S-217	syenogranite	40	19	40							1	tr			tr		
D-168	syenogranite	40	16	40							3	tr		1	tr		
L-181	qtz granitoid	70	10	20							tr				tr		
L-187	k granite	49		50			0.5				0.5	tr		tr			
S-153	syenogranite	30	15	45			tr				8	tr	tr	2	tr		
S-154a	granite	40	23	27			1				9				tr		
Late Suite																	
D-19	granite	40	27	30							3	tr			tr		tr cc
L-67b	granite	30	25	40						tr	5	tr	tr	tr			
L-68	granite	25	25	45						tr	5	tr					
L-95	granite	24	35	40							1	tr	tr				plag lots ser
L-96	granite	35	25	39							0.5	0.5	tr		tr		
S-44	syenogranite	40	15	40							2	3					plag lots ser
S-33	k granite	45	3	50							1	tr		tr	1		
D-259	qtz syenite	10	10	78						tr	1				1		
L-33	qtz syenite	10	25	63						tr	2	tr					tr cc
S-218	qtz k syenite	19		80							1	tr					
E-41	syenogranite	20	20	59							1	tr			tr		
M-73	granite	25	35	35						tr	tr	5					
Cape Smith Suite																	
D-268a	diorite		30					tr	68	tr			tr		2		plag lots ser; hb to ep,trem,chl
D-272	diorite	2	48				40	tr		10					tr		hb varibly to ep; all bio to chl
D-273	tonalite	20	35			10	30				2				3		cpx with exsolved Fe-Ti oxides
D-274	tonalite	10	30				50	tr		9					1		plag lots ser; hb varibly to ep; all bio to chl
D4-2	diorite	2	40				45	1		10			tr		2		plag lots ser; bio to chl
D5-2	tonalite	20	50					15		14					1	tr	plag lots ser; hb to ep; bio to chl

Sample	t.s. name	qtz	plag	kspar	opx	cpx	hb	epid	trem	chl	bio	mu	ap	titan	opq	gt	other/comments
D8-1	qtz diorite	2	15					60		20			tr		3		plag lots ser; hb to ep; bio to chl
D8-2	qtz diorite	5	55					35		2					3		plag lots ser; hb to ep
CP2-7	diorite		40		10		40				tr		tr		10		
CP3-3	qtz diorite	5	45					45							5		plag lots ser; hb to ep
CP5-3	qtz diorite	5	45		20	25					tr		tr		5		
D-271	qtz diorite	10	70					5		15							plag lots ser
D-275	tonalite	30	60								10	tr					plag lots ser; tr cc
D-277	tonalite	25	50					15		tr	10			tr	tr		plag lots ser; bio varibly to chl
D-284	qtz diorite	10	55				15	tr		tr	19			tr	1		plag lots ser; bio variably to chl; hb varibly to ep
D3-2	tonalite	20	40	5			tr	20		tr	15	tr	tr	tr			many ep with allan cores; lots mu xtals on plag
D4-3	tonalite	20	55					15		10					tr		tr cc; plag with variable ser
CP3-4	tonalite	20	60					10		10			tr	tr	tr		plag lots ser; hb to ep; bio to chl
D-42	granite	35	29	35		0.5	0.5	tr						tr			1 huge euhedral cpx (10mm)
D-239	qtz granitoid	50	30	tr				4			10	6		tr			
D-240	granite	50	20	15				3			10	2		tr			
D-270	tonalite	40	50							tr	10	tr					bio variably to chl
D-278	qtz diorite	10	80	5		5		tr						tr			1 large euhedral cpx (7mm)
DL1-4	qtz granitoid	45	20	10				5			14	5	1	tr			

Abreviations: t.s. name=thin section name; qtz=quartz; plag=plagioclase feldspar; k-spar=k-feldspar; opx=orthopyroxene; cpx=clinopyroxene; hb=hornblende; ep=epidote; trem=tremolite-actinolite; chl=chlorite; bio=biotite; mu=muscovite; ap=apatite; titan=titanite; opq=opaque oxides; gt=garnet; allan=allanite; ser=sericite; cc=calcite; tr=trace amount

APPENDIX 2

Sample preparation and analytical techniques for geochemical analyses.

Sample preparation

Samples for geochemical analysis were scrubbed and cleaned prior to crushing. Samples were sliced into thin (≈1 cm thick) slabs with an electric water-cooled rock saw, and these slabs were subsequently broken into smaller (cm^3) pieces with the aid of a hammer. Approximately 150-200 g of sample were then ground to a fine powder in a shatterbox. A tungsten-carbide shatterbox was used for a small subset of samples while most others were powdered in an agate shatterbox.

Major element analysis

Major element analysis was performed on glass pellets that were prepared in the following manner (described in detail in Norrish and Hutton 1969). Samples were dried at 1050°C for 12 hours, mixed with a lithium tetraborate flux, then fused in a platinum crucible and rapidly quenched to form a glass pellet. Samples were analyzed by X-ray fluorescence (XRF) at the Université de Montréal and at McGill University. Analytical precision for the major element analyses is better than 1%.

Trace element analysis

Samples for trace element analysis (Nb, Y, Rb, Sr, Cr, Ni, Zr, Cu, Zn) were performed on powder pellets. These pellets were made from the sample powder mixed with a polyvinyl glue to hold the sample together, which was then surrounded by boric acid powder and compressed to 12 tonnes pressure. Trace element concentrations were determined by XRF, with typical analytical precision for these elements of approximately 5%. REE and other trace elements (Th, U, Ba, Hf, Ta, Sc, Cs, W) were analyzed by instrumental neutron activation analysis (INAA) at the Université de Montréal. Sample powders were placed into sealed plastic capsules which were irradiated in the SLOWPOKE II nuclear reactor located at the Ecole Polytechnique. Most of these analyses gave analytical precision of better than 5%, although the low abundances of the heavy REE resulted in precisions of 5-10%. Irregular concentrations for Gd, Ho and Tm were often obtained due to technical difficulties in analyzing for these elements (e.g., the half-life of the Ho isotope (^{166}Ho) is so short that it is almost completely disintegrated after 5 days but the samples are normally too "hot" (radioactive) at this point to handle safely. Hence such elements generally were not analyzed and are not included in the geochemical tables.

Sr- and Sm-Nd Isotopic analysis

Analytical techniques for the preparation of samples for isotopic analyses were modified from Hart and Brooks (1977) and Zindler et al. (1979). All reagents and materials were ultra-clean and all chemistry was carried out in laminar flow cabinets and fumehoods. Approximately 40-250 mg of sample powder was dissolved in an $HF-HClO_4$ solution in teflon beakers. A Sm-Nd spike ($^{149}Sm/^{150}Nd$) was added to the samples for isotope dilution analyses prior to dissolution. Rb and Sr concentrations were determined by XRF. Samples were heated for 24 h at 50°C in closed beakers, then dried at approximately 75-125°C. The precipitate was converted into soluble chlorides by adding HCl (6.2N) and dried once again. This step was repeated to ensure complete conversion. Finally the samples was redissolved in 2.5N HCl and loaded onto cation exchange columns. The cation columns were calibrated with standard solutions and samples were then extracted at the appropriate intervals. Sr was eluted with 2.5N HCl and the REE's were then extracted with 6.2N HCl. The Sr separate was dried down and was then ready for IC (isotope composition) analysis. The REE's needed further purification to separate Sm and Nd, and the REE solution was loaded onto anion exchange columns with the Nd collected by elution with 0.25N HCl while the Sm was removed in 0.6N HCl. Samples were dried and were then ready to load for IC/ID (isotopic composition and isotope dilution) analysis.

Mass spectrometry for the isotopic analysis was performed at different laboratories, with repeat samples performed for inter-correlation purposes. No discrepancies were found. Sr (IC) was analyzed at the Université de Montréal on a National Bureau of Standards (NBS) single collector mass spectrometer and on a VG Sector 54 mass spectrometer at the Université du Québec à Montréal. Analysis of the Sr standard E&A yielded $^{87}Sr/^{86}Sr = 0.70804 +/- 0.00003$ (2σ), while NBS 987 gave 0.710270 +/- 0.000018 (2σ), normalized to $^{86}Sr/^{88}Sr = 0.1194$. 1.42×10^{-11} is the decay constant used for ^{87}Rb. Sm-Nd (IC/ID) analyses were performed on a Finnigan MAT

261 mass spectrometer at the Geological Survey of Canada (Ottawa). All Sm-Nd data are corrected to give La Jolla $^{143}Nd/^{144}Nd = 0.511860$, normalized to $^{146}Nd/^{144}Nd = 0.7219$. Analysis of BCR-1 gave $^{143}Nd/^{144}Nd = 0.512652$ +/- 0.000011 (2σ), in agreement with the accepted value of 0.512643 +/- 0.000019 (Carlson 1984). Replicate reproducibility is approximately 0.006% for $^{143}Nd/^{144}Nd$ and 0.4% for $^{147}Sm/^{144}Nd$ and is generally better than 0.5 epsilon units at the crystallization age. Present day values of the chondritic uniform reservoir (CHUR) used for epsilon Nd calculations are $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$. The decay constant for ^{147}Sm is 6.54×10^{-12}.

U-Pb TIMS and Pb-Pb laser ablation ICP-MS geochronology

Mineral separates for geochronology were extracted from approximately 30 kg of sample by standard techniques of crushing and grinding, followed by Wilfley table concentration, heavy liquid and magnetic separation. Hand picking of grains, using criteria such as morphology, clarity and absence of cracks, was used to select the highest quality zircon grains for isotopic analysis.

Analytical methods for conventional thermal ionization mass spectrometry (TIMS) analysis are summarized by Parrish et al. (1987) and include air abrasion (Krogh 1982), chemical dissolution and separation of U and Pb, followed by isotopic analysis on a Finnigan Mat 261 variable multicollector mass spectrometer at the Geological Survey of Canada, Ottawa.

The analytical method and operating parameters used for the laser ablation technique are presented in detail in Feng et al. (1993). For laser ablation ICP-MS analysis (inductively coupled plasma mass spectrometry), zircons were selected from the same concentrate and using the same criteria that was used for the TIMS analysis. Zircons were mounted in epoxy and slightly polished to expose the grains, which resulted in a reduction of their size (thickness). The zircon samples were ablated with a Nd:YAG laser which produced cylindrical pits of ≈ 40-60 µm in diameter for these samples. For each sample pit, 10-20 repeats were performed, depending on the size of the zircon sample. Given the drilling rate of the laser estimated by Feng et al. (1993) of 0.1-0.3 μm/s and an analytical time of measurement of 100-200 s (depending on the number of repeats), a minimum grain thickness of 60 μm was necessary to ensure a complete run. The vaporized material was transported to the ICP-MS by an argon carrier gas for analysis. A Fisons VG PQII+ICP mass spectrometer and a Fisons VG Laser-Probe located at the Université de Montréal were used to measure the isotopic ratios of the samples. Correction for mass discrimination and instrument drift was made by using an external standard, a large zircon (CN-92-1) from a skarn in the Grenville Province which has an average $^{207}Pb/^{206}Pb$ ratio of 0.07784 ± 0.00006 (2σ) (TIMS analysis, Feng et al. 1993). The $^{207}Pb/^{206}Pb$ ages and their uncertainties were calculated using the ISOPLOT program of Ludwig (1990).

APPENDIX 3

Complete geochemical data for all Narsajuaq terrane samples.

Older Suite

Sample	D-41c	D-64	L-20b	L-90a	L-90b	L-174	L-211	S-18a	M-34	M-40	M-103	D-72	D-163	D-211	L-1	L-175a	M-89	D-14a	D-23	D-29	D-41a	L-27	K-67
Lithology	diorite	qtz dior	qtz dior	diorite	diorite	qtz dior	diorite	qtz dior	qtz dior	diorite	mzdior	tonalite	tonalite	grano	grano	tonalite	tonalite	grano	granite	tonalite	granite	tonalite	tonalite
SiO2 (wt %)	52.45	57.06	54.22	53.83	49.58	53.89	47.34	54.30	58.21	46.28	56.34	67.67	62.66	65.19	65.50	65.19	64.24	73.81	71.15	74.85	75.44	79.56	81.13
TiO2	1.03	0.98	0.77	1.04	0.95	0.69	0.81	0.84	0.83	1.80	0.85	0.47	0.84	0.51	0.53	0.64	0.77	0.08	0.19	0.04	0.08	0.17	0.26
Al2O3	16.65	14.15	18.72	20.25	16.19	16.91	14.64	19.01	18.43	16.44	16.79	16.24	18.04	15.34	16.21	15.50	16.02	15.03	16.05	15.11	14.45	11.74	10.82
Fe2O3	10.13	13.14	6.95	7.85	10.67	9.92	12.48	8.15	8.90	15.80	6.95	3.71	5.54	5.38	3.83	6.89	5.90	0.97	1.35	0.56	1.09	1.35	1.43
MnO	0.16	0.44	0.10	0.09	0.18	0.21	0.23	0.16	0.15	0.19	0.09	0.06	0.05	0.08	0.09	0.13	0.07	0.01	0.01	0.01	0.01	0.02	0.01
MgO	4.52	3.83	4.21	2.96	7.33	4.38	8.97	3.00	1.89	6.06	3.90	1.71	1.76	2.18	1.29	1.92	2.38	0.01	0.38	0.02	0.04	0.42	0.26
CaO	8.55	7.54	7.13	6.01	9.31	8.34	11.19	6.49	7.58	8.90	5.06	3.23	5.07	3.75	2.72	5.19	3.68	1.81	2.44	1.29	1.67	2.20	2.01
Na2O	4.25	1.83	5.82	5.15	3.81	2.88	2.11	4.64	3.28	2.83	4.37	4.14	4.99	3.47	4.27	3.41	3.89	4.32	4.90	6.48	3.70	3.74	3.45
K2O	0.87	1.12	1.35	2.05	1.28	2.39	0.41	1.73	1.51	1.18	4.07	2.11	1.23	2.83	3.37	1.53	2.04	3.68	2.51	1.11	3.83	1.02	0.84
P2O5	0.28	0.20	0.29	0.33	0.24	0.16	0.10	0.28	0.24	0.51	0.65	0.15	0.29	0.12	0.22	0.15	0.20	0.02	0.04	0.01	0.04	0.03	0.05
LOI	0.28	1.49	0.65	0.28	0.40	1.49	1.68	0.85	0.50	1.42	0.86	0.64	0.47	0.39	0.44	0.24	1.16	0.33	0.48	0.27	0.37	0.28	0.21
TOTAL	99.17	101.78	100.21	99.84	99.94	101.26	99.96	99.45	101.52	101.41	99.93	100.13	100.94	99.24	98.46	100.79	100.35	100.07	99.50	99.75	100.72	100.53	100.47
Th (ppm)	1.7	4.16	5.37	0.9	0.7	0.82	1.4	1.5	1.26	nd	nd	6.3	4.49	7.67	4.7	nd	nd	8.68	1.57	0.16	12.0	4.75	nd
U	2.4	3.26	1.6	0.7	0.2	0.49	1.17	0.9	0.52	nd	nd	0.8	0.13	0.83	1.7	nd	nd	1.46	0.2	0.41	1.8	0.53	nd
Rb	6.7	32.5	34.5	37.0	15.2	61.8	5.5	40.0	73.3	26.6	84.7	61.2	13.6	90.1	76.0	nd	45.5	76.9	66.6	80.0	80.7	28.1	12.0
Sr	660	234	1524	1398	836	324	145	928	296	811	1652	641	899	478	615	nd	517	405	596	576	389	529	610
Y	22.5	29.5	25.9	14.8	27.9	23.0	16.5	18.2	28.5	19.8	15.6	10.0	9.2	8.1	16.8	nd	15.1	2.9	2.0	2.4	4.2	7.0	0.9
Zr	90	112	292	384	134	75	51	203	87	47	307	152	288	99	233	nd	163	180	77	53	187	120	260
Nb	8.2	9.6	14.3	5.6	8.0	4.0	4.7	10.0	5.2	7.6	14.7	4.5	7.7	10.1	20.8	nd	11.4	2.4	2.2	8.0	3.2	3.9	2.3
Ba	374	432	525	1489	542	910	124	1012	387	nd	nd	824	1223	1506	1258	nd	nd	951	1175	955	1106	310	nd
Hf	2.40	3.36	6.50	9.50	3.70	2.23	1.36	5.70	2.99	nd	nd	4.50	6.31	2.58	6.60	nd	nd	4.91	1.87	1.27	6.20	3.12	nd
Ta	2.1	nd	nd	1.3	1.0	nd	0.24	1.6	nd	nd	nd	1.5	nd	0.35	2.9	nd	nd	nd	nd	nd	2.5	nd	nd
Cr	115.8	182.0	172.7	70.1	230.6	91.6	nd	22.1	10.7	19.6	nd	33.6	10.2	nd	14.3	nd	nd	6.8	10.1	3.9	12.1	4.5	nd
Ni	45.1	63.9	56.2	29.7	145.8	22.5	nd	11.3	1.9	26.9	nd	10.3	10.2	nd	9.7	nd	nd	4.9	8.5	4.8	4.9	6.6	nd
Zn	98.0	176.7	82.1	99.7	114.4	93.4	nd	110.3	167.9	131.7	nd	62.8	97.2	nd	74.9	nd	nd	19.4	29.5	19.5	15.7	28.9	nd
Cu	14.6	175.4	25.6	6.7	105.1	193.6	nd	38.9	153.0	109.7	nd	4.8	24.2	nd	20.4	nd	nd	4.2	48.6	3.9	2.8	11.2	nd
Co	35.3	35.2	20.1	22.6	44.3	26.0	48.3	35.2	15.7	nd	nd	28.8	10.9	14.0	40.3	nd	nd	0.8	2.9	0.5	52.6	1.9	nd
Sc	27.21	nd	nd	16.44	27.41	nd	34.19	15.41	nd	nd	nd	7.60	nd	11.18	4.77	nd	nd	nd	nd	nd	0.90	nd	nd
Cs	0.2	nd	nd	0.6	0.5	nd	nd	1.0	nd	nd	nd	1.7	nd	0.9	1.9	nd	nd	nd	nd	nd	1.2	nd	nd
W	397	2	3	304	140	3	24	334	4	nd	nd	355	nd	39	535	nd	nd	1	1	nd	602	1	nd
Au	0.002	0.005	0.011	0.002	0.002	0.007	0.009	0.002	0.006	nd	nd	0.001	0.011	0.007	0.005	nd	nd	0.001	0.022	0.002	0.005	0.004	nd
La	8.90	20.06	28.71	12.70	15.77	6.97	13.36	20.16	9.36	nd	nd	31.08	83.84	18.93	42.72	nd	nd	15.03	8.53	0.50	9.50	19.53	nd
Ce	21.40	36.95	55.05	26.70	38.83	13.45	25.50	46.09	18.29	nd	nd	62.62	159.23	45.47	88.05	nd	nd	22.26	13.12	1.02	23.17	36.84	nd
Nd	15.00	22.77	29.11	17.60	24.48	9.94	9.66	23.70	9.96	nd	nd	26.87	56.17	10.93	38.24	nd	nd	7.83	3.87	0.41	6.14	12.65	nd
Sm	3.38	4.87	6.26	3.37	5.06	2.55	2.76	4.04	3.48	nd	nd	3.88	7.50	2.67	5.90	nd	nd	1.68	0.74	0.09	1.03	2.54	nd
Eu	1.07	1.39	1.40	1.20	1.32	0.75	0.63	1.30	1.09	nd	nd	1.02	1.86	1.04	1.27	nd	nd	0.58	0.48	0.10	0.56	0.74	nd
Tb	0.60	0.93	0.67	0.49	0.81	0.48	0.31	0.55	0.63	nd	nd	0.34	0.45	0.23	0.54	nd	nd	0.06	0.03	0.02	0.08	0.23	nd
Ho	0.84	0.87	1.17	0.99	1.22	0.63	nd	0.93	0.79	nd	nd	0.52	0.42	0.22	0.69	nd	nd	0.03	0.09	0.03	0.42	0.21	nd
Yb	2.30	2.52	2.66	1.50	2.83	2.84	1.44	1.86	3.05	nd	nd	0.97	0.82	1.17	1.68	nd	nd	0.38	0.61	0.30	0.62	0.43	nd
Lu	0.32	0.41	0.33	0.21	0.40	0.47	0.21	0.30	0.44	nd	nd	0.18	0.14	0.17	0.26	nd	nd	0.06	0.05	0.03	0.10	0.08	nd
A/CNK	0.71	0.79	0.78	0.94	0.66	0.75	0.60	0.89	0.89	0.74	0.81	1.08	0.96	0.98	1.04	0.93	1.05	1.05	1.05	1.06	1.09	1.04	1.06
(La/Yb)n	2.8	5.7	7.7	6.1	4.0	1.8	6.7	7.8	2.2	nd	nd	23.0	73.3	11.6	18.2	nd	nd	28.4	10.0	1.2	11.0	32.6	nd
(La/Sm)n	1.7	2.7	3.0	2.4	2.0	1.8	3.1	3.2	1.7	nd	nd	5.2	7.2	4.6	4.7	nd	nd	5.8	7.4	3.6	6.0	5.0	nd

Younger Suite

Sample	D-26	D-28	D-165	L-18	L-225	S-34	S-39a	S-106	K-69	D-20	L-43	S-12	E-4	D-24	D-43	L-38a	L-39	S-19	L-118	D-220	D-222	D-283
Lithology	diorite	diorite	diorite	diorite	diorite	qtz dior	diorite	diorite	diorite	mzdior	mzdior	mzdior	mzdior	qtz mzdior	mzdior	mzdior	mzdior	mzdior	tonalite	IVUJ gran	IVUJ gran	IVUJ gran
SiO2 (wt %)	52.15	52.63	53.14	46.20	47.86	54.56	52.44	45.38	50.83	55.85	57.92	59.50	59.20	60.38	57.73	57.02	57.85	57.41	66.10	74.33	76.07	69.98
TiO2	0.75	0.58	0.88	1.60	1.52	0.94	0.76	1.54	1.45	0.93	0.78	0.72	0.71	0.69	0.70	0.76	0.76	0.68	0.50	0.25	0.12	0.50
Al2O3	15.59	16.65	19.54	15.40	13.72	17.38	16.32	15.32	15.59	18.19	17.49	17.03	16.33	15.26	14.65	14.68	14.97	17.76	16.95	12.73	13.40	14.29
Fe2O3	8.73	8.27	8.72	12.81	11.54	8.76	9.07	14.86	12.59	6.80	6.12	5.53	4.97	5.56	5.73	6.57	6.03	6.22	3.77	2.85	1.69	3.32
MnO	0.18	0.15	0.13	0.19	0.18	0.14	0.15	0.24	0.18	0.10	0.07	0.08	0.08	0.09	0.09	0.09	0.10	0.11	0.03	0.04	0.06	0.02
MgO	6.68	6.57	4.43	6.88	8.10	4.69	6.49	7.83	5.68	3.80	3.09	2.86	3.07	3.47	4.50	4.42	4.34	2.93	1.21	0.01	0.32	0.41
CaO	8.96	9.18	8.49	8.96	11.40	7.56	9.22	10.33	8.46	6.09	5.37	4.68	4.78	4.98	6.41	6.24	6.15	5.17	3.27	0.69	2.13	1.26
Na2O	3.54	3.75	4.27	3.79	2.78	3.30	2.97	2.72	3.70	5.10	5.17	4.80	5.08	4.30	4.31	4.51	4.83	5.62	4.61	3.19	3.51	3.09
K2O	0.71	0.65	0.84	2.39	0.98	1.30	0.76	1.56	1.03	2.86	3.09	3.88	4.08	3.93	4.03	4.05	4.04	3.48	3.28	5.57	2.36	6.07
P2O5	0.32	0.15	0.22	0.79	0.76	0.32	0.29	0.23	0.23	0.58	0.47	0.44	0.60	0.49	0.85	0.82	0.84	0.42	0.16	0.02	0.04	0.14
LOI	1.13	0.97	0.62	0.68	0.65	0.75	0.71	0.52	0.88	0.60	0.32	0.54	0.32	0.33	0.36	0.40	0.59	0.47	0.64	0.14	0.27	0.37
TOTAL	98.74	99.55	101.28	99.69	99.49	99.69	99.18	100.53	100.62	100.90	99.89	100.06	99.21	99.48	99.36	99.56	100.50	100.27	100.52	99.82	99.97	99.45
Th (ppm)	0.8	0.4	0.28	nd	nd	1.0	0.7	1.31	0.42	7.1	4.5	7.59	8.4	14.0	5.6	6.8	4.89	6.92	nd	nd	nd	1.21
U	0.3	0.2	0.02	nd	nd	0.3	0.3	0.53	0.61	1.5	1.0	1.28	2.8	2.4	1.4	2.0	1.12	1.36	nd	nd	nd	0.36
Rb	5.6	8.8	11.5	43.9	15.3	27.0	7.4	18.5	10.0	34.6	40.6	51.0	91.2	93.2	88.9	101.7	85.6	69.1	60.3	231.3	55.9	117.8
Sr	1305	907	833	1199	1128	799	861	305	372	1910	1707	1701	2209	1217	2145	2101	2056	2313	568	53	172	402
Y	17.5	12.7	18.9	36.8	34.6	19.5	15.7	31.9	26.0	19.9	16.5	16.3	16.3	19.5	21.3	19.9	21.5	22.0	7.0	96.6	43.6	12.3
Zr	68	47	75	394	207	120	53	99	145	261	209	238	410	227	186	226	185	264	98	476	249	524
Nb	5.6	3.0	4.4	29.6	12.3	7.4	3.8	9.7	8.6	13.3	10.5	12.1	16.6	15.3	11.6	11.0	11.5	14.2	5.8	20.6	6.1	10.1
Ba	1045	445	556	nd	nd	667	494	543	249	2240	2119	2598	2519	1549	2662	2564	2240	2867	nd	nd	nd	2507
Hf	1.50	1.30	1.73	nd	nd	3.70	1.70	2.43	3.25	5.90	5.40	5.06	9.70	5.40	3.80	5.30	3.68	4.04	nd	nd	nd	12.74
Ta	0.7	0.8	nd	nd	nd	1.2	0.9	0.36	0.27	2.1	1.1	nd	1.8	1.5	1.7	0.9	nd	nd	nd	nd	nd	0.25
Cr	358.1	210.7	30.8	nd	nd	81.6	175.2	nd	nd	77.6	72.1	71.0	58.5	90.3	113.4	137.6	118.5	61.0	nd	nd	nd	nd
Ni	83.9	59.3	28.6	nd	nd	33.0	53.0	84.9	nd	47.0	37.9	42.1	43.1	43.3	70.1	71.4	66.8	34.0	21.0	4.8	4.5	6.4
Zn	102.4	79.4	95.6	nd	nd	99.2	89.3	150.0	nd	112.0	90.2	95.1	86.0	84.5	92.0	102.2	98.1	110.5	54.3	70.8	23.2	53.1
Cu	13.5	8.1	42.0	nd	nd	123.0	71.5	16.9	nd	8.7	8.7	25.6	18.5	52.8	89.3	115.7	69.1	61.3	6.1	5.2	2.6	5.1
Co	34.3	38.8	24.4	nd	nd	36.5	48.2	43.1	41.0	28.4	27.4	15.1	33.5	33.0	27.1	28.7	18.0	13.7	nd	nd	nd	3.3
Sc	22.69	23.80	nd	nd	nd	20.51	24.71	39.81	25.63	11.08	9.78	nd	7.83	10.82	11.05	10.94	nd	nd	nd	nd	nd	1.50
Cs	0.2	0.5	nd	nd	nd	0.6	0.6	0.35	0.47	0.5	0.8	nd	2.0	1.7	1.3	2.0	nd	nd	nd	nd	nd	0.27
W	141	174	2	nd	nd	212	242	13	21	477	218	2	327	257	355	154	1	0	nd	nd	nd	2
Au	0.004	0.001	0.009	nd	nd	0.005	0.006	nd	nd	0.007	0.001	0.003	0.003	0.003	0.003	0.002	0.003	0.012	nd	nd	nd	nd
La	32.85	10.41	21.86	nd	nd	19.42	13.62	24.94	14.84	77.00	64.09	71.13	94.90	65.23	105.10	108.52	103.28	80.97	nd	nd	nd	32.49
Ce	66.43	23.10	51.20	nd	nd	41.50	30.92	50.14	32.98	160.80	126.78	137.57	183.91	132.96	209.70	210.96	146.89	147.76	nd	nd	nd	52.85
Nd	33.04	14.71	33.52	nd	nd	22.81	17.96	29.17	16.20	78.20	63.29	56.95	82.94	61.40	107.10	101.47	88.90	60.83	nd	nd	nd	17.87
Sm	6.47	2.93	6.91	nd	nd	4.73	3.90	6.56	5.09	11.76	9.99	10.45	14.08	10.86	17.11	17.10	17.45	12.46	nd	nd	nd	3.98
Eu	1.92	1.07	1.79	nd	nd	1.28	1.16	1.93	1.51	2.75	2.31	2.51	3.48	2.34	3.74	3.79	4.16	3.33	nd	nd	nd	1.68
Tb	0.68	0.37	0.55	nd	nd	0.66	0.48	0.87	0.75	0.85	0.76	0.68	0.92	0.91	1.09	1.08	1.09	1.88	nd	nd	nd	0.37
Ho	0.84	0.37	0.31	nd	nd	0.84	0.65	1.60	1.39	1.08	0.98	0.66	1.20	0.87	1.05	0.74	0.57	0.62	nd	nd	nd	nd
Yb	1.75	1.34	1.78	nd	nd	2.06	1.63	2.99	2.43	1.40	1.46	1.29	1.38	1.81	1.70	1.37	1.45	1.21	nd	nd	nd	1.07
Lu	0.27	0.19	0.21	nd	nd	0.41	0.28	0.43	0.39	0.16	0.19	0.14	0.23	0.23	0.20	0.19	0.17	0.16	nd	nd	nd	0.18
A/CNK	0.68	0.71	0.84	0.61	0.52	0.84	0.73	0.61	0.69	0.81	0.81	0.83	0.76	0.75	0.63	0.63	0.64	0.79	0.99	1.02	1.10	1.02
(La/Yb)n	13.5	5.6	8.8	nd	nd	6.8	6.0	6.0	4.4	39.5	31.5	39.6	49.3	25.9	44.3	56.8	51.1	48.0	nd	nd	nd	21.8
(La/Sm)n	3.3	2.3	2.0	nd	nd	2.7	2.3	2.5	1.9	4.2	4.1	4.4	4.4	3.9	4.0	4.1	3.8	4.2	nd	nd	nd	5.3

Late Suite

Sample	L-205	S-217	D-168	L-181	L-187	S-153	S-154a	L-67b	L-68	D-19	L-95	L-96	S-44	S-33	D-259	L-33	E-41	S-218	M-73
Lithology	IVUJ gran	IVUJ gran	CNF syen	CNF gran	CNF gran	CNF gran	CNF gran	granite	granite	granite	granite	granite	granite	granite	syeno	syeno	syeno	syeno	syeno
SiO2 (wt %)	70.27	74.14	75.17	72.31	70.08	70.64	77.05	74.01	74.60	74.26	72.47	75.03	75.99	75.07	72.36	71.25	72.14	74.25	72.67
TiO2	0.32	0.13	0.34	0.04	0.60	0.46	0.18	0.14	0.09	0.10	0.07	0.08	0.07	0.14	0.11	0.10	0.06	0.08	0.07
Al2O3	15.27	13.95	12.71	16.63	14.25	14.72	12.89	14.37	14.06	14.77	15.86	14.16	14.21	12.70	14.65	16.44	16.09	14.64	15.56
Fe2O3	2.69	1.25	1.70	0.29	3.58	3.08	1.09	1.10	0.80	0.85	0.59	0.78	0.59	1.29	1.97	0.85	0.73	0.61	0.88
MnO	0.02	0.01	0.04	0.01	0.10	0.05	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01
MgO	0.51	0.12	0.24	0.09	0.45	0.67	0.43	0.24	0.11	0.19	0.08	0.18	0.09	0.14	0.09	0.12	0.04	0.08	0.13
CaO	2.02	1.06	1.30	2.07	2.61	1.93	1.86	1.20	1.12	1.22	1.33	0.84	1.09	0.54	1.07	1.84	1.56	1.22	1.29
Na2O	3.55	3.61	3.13	4.94	3.38	3.46	3.47	3.45	3.37	3.83	4.49	3.36	3.98	2.55	3.47	4.36	4.02	3.42	4.41
K2O	4.71	4.87	4.75	3.44	4.39	4.95	3.16	4.61	4.82	4.52	4.66	4.94	3.67	5.75	5.32	3.98	5.44	5.35	3.39
P2O5	0.14	0.04	0.05	0.01	0.25	0.14	0.04	0.05	0.05	0.04	0.03	0.04	0.03	0.04	0.03	0.04	0.02	0.04	0.06
LOI	0.25	0.27	0.33	0.32	0.27	0.29	0.46	0.40	0.20	0.20	0.83	0.40	0.28	0.20	0.33	0.24	0.42	0.32	0.69
TOTAL	99.75	99.45	99.76	100.15	99.96	100.39	100.65	99.58	99.23	99.99	100.42	99.82	100.01	98.44	99.41	99.23	100.53	100.02	99.16
Th (ppm)	24.02	8.23	15.95	0.24	4.62	13.4	2.52	9.6	9.7	12.3	nd	17.2	6.5	9.3	7.73	3.2	17.83	1.48	nd
U	0.75	0.67	0.62	0.14	0.53	0.57	0.23	2.1	1.6	1.5	nd	2.5	1.4	1.4	0.92	0.7	1.04	0.27	nd
Rb	142.4	98.2	64.6	34.1	92.5	115.2	72.3	98.3	99.9	97.2	71.6	106.5	110.6	172.3	94.1	62.4	118.4	112.3	103.7
Sr	462	274	271	650	530	305	312	310	250	259	311	195	272	110	777	765	299	245	355
Y	8.8	1.9	57.2	1.0	73.1	31.4	4.3	2.7	2.9	3.2	1.5	3.5	3.3	7.5	1.6	2.8	7.0	3.0	4.9
Zr	470	123	313	17	490	340	105	133	103	106	128	84	53	139	283	76	63	51	48
Nb	6.6	4.2	24.3	1.3	32.7	18.4	3.7	8.3	7.3	6.5	4	7.4	6.7	8.9	2.5	2.3	2.1	4.8	9
Ba	1649	773	3200	2200	2913	1753	1371	1615	974	1145	nd	1010	725	747	3408	2801	968	959	nd
Hf	13.44	3.18	7.71	0.21	13.53	9.46	2.70	4.10	3.20	3.40	nd	3.20	1.80	5.10	5.08	1.90	1.86	1.94	nd
Ta	nd	nd	nd	nd	nd	nd	nd	6.6	5.0	3.1	nd	6.1	4.8	3.2	0.06	2.0	nd	0.08	nd
Cr	nd	nd	nd	5.1		11.5	7.0	4.9	3.3	6.5	nd	4.5	4.3	6.6	nd	9.0	4.3	nd	nd
Ni	nd	3.7	3.8	5.8	4.4	6.4	5.1	6.9	4.7	5.4	nd	7.0	5.5	5.4	6.3	4.8	5.9	4.6	2.6
Zn	nd	23.8	60.2	8.5	102.1	65.1	17.7	39.3	27.9	29.1	nd	24.0	15.0	13.9	19.0	13.9	12.9	16.2	16.0
Cu	nd	0.9	4.2	4.8	3.5	12.2	6.0	6.0	2.4	4.1	nd	3.9	4.6	7.4	4.2	5.5	3.3	4.7	18.3
Co	3.9	0.6	1.0	0.5	2.1	4.3	2.0	58.8	37.9	35.1	nd	74.7	69.1	47.3	1.0	28.1	0.6	0.9	nd
Sc	1.69	1.11	nd	nd	nd	nd	nd	1.23	0.99	1.07	nd	1.15	0.70	0.50	0.60	0.69	nd	0.87	nd
Cs	0.21	0.2	nd	nd	nd	nd	nd	0.8	0.7	0.7	nd	1.3	2.5	1.2	0.37	0.6	nd	0.27	nd
W	nd	nd	3	nd	5	1	2	1675	1259	751	nd	1448	1196	893	nd	547	1	nd	nd
Au	0.01	nd	0.002	0.001	0.002	0.005	0.011	0.001	0.004	0.001	nd	0.017	0.006	0.002	nd	0.001	0.003	nd	nd
La	75.80	23.75	412.01	17.30	109.64	135.79	24.53	11.20	21.20	23.90	nd	12.20	10.70	16.50	35.75	16.15	56.23	15.91	nd
Ce	117.17	45.08	774.36	20.44	210.04	207.00	31.10	41.20	35.10	41.10	nd	29.70	18.30	29.48	67.15	25.10	94.28	22.25	nd
Nd	32.87	10.35	244.13	4.22	102.88	80.80	9.68	10.20	12.90	13.90	nd	8.60	7.70	11.72	18.73	10.69	48.99	8.13	nd
Sm	4.49	1.62	35.18	0.39	19.70	12.36	1.39	1.65	1.68	1.76	nd	1.57	0.88	1.51	2.95	1.21	8.85	1.29	nd
Eu	1.13	0.66	5.30	0.83	4.44	1.88	0.97	0.57	0.55	0.55	nd	0.53	0.38	0.38	0.77	0.54	1.16	1.37	nd
Tb	0.39	0.06	1.99	0.02	2.16	0.97	0.10	0.17	0.09	0.11	nd	0.15	0.15	0.20	0.09	0.12	0.47	0.07	nd
Ho	nd	0.54	3.03	0.09	1.50	0.99	0.30	0.19	0.63	0.15	nd	0.76	0.08	0.55	nd	0.23	0.65	nd	nd
Yb	0.94	nd	5.11	0.20	7.01	3.25	0.35	0.40	0.50	0.50	nd	0.50	0.40	0.75	0.28	0.18	0.75	0.30	nd
Lu	0.16	0.03	0.58	0.03	0.98	0.44	0.05	0.13	0.06	0.08	nd	0.06	0.05	0.13	0.06	0.09	0.09	0.05	nd
A/CNK	1.05	1.06	1.00	1.07	0.95	1.01	1.03	1.12	1.10	1.10	1.07	1.14	1.14	1.11	1.09	1.11	1.05	1.07	1.17
(La/Yb)n	57.8	nd	57.8	62.0	11.2	30.0	50.3	20.1	30.4	34.3	nd	17.5	19.2	15.8	91.6	64.4	53.8	38.0	nd
(La/Sm)n	10.9	9.5	7.6	28.6	3.6	7.1	11.4	4.4	8.1	8.8	nd	5.0	7.8	7.1	7.8	8.6	4.1	8.0	nd

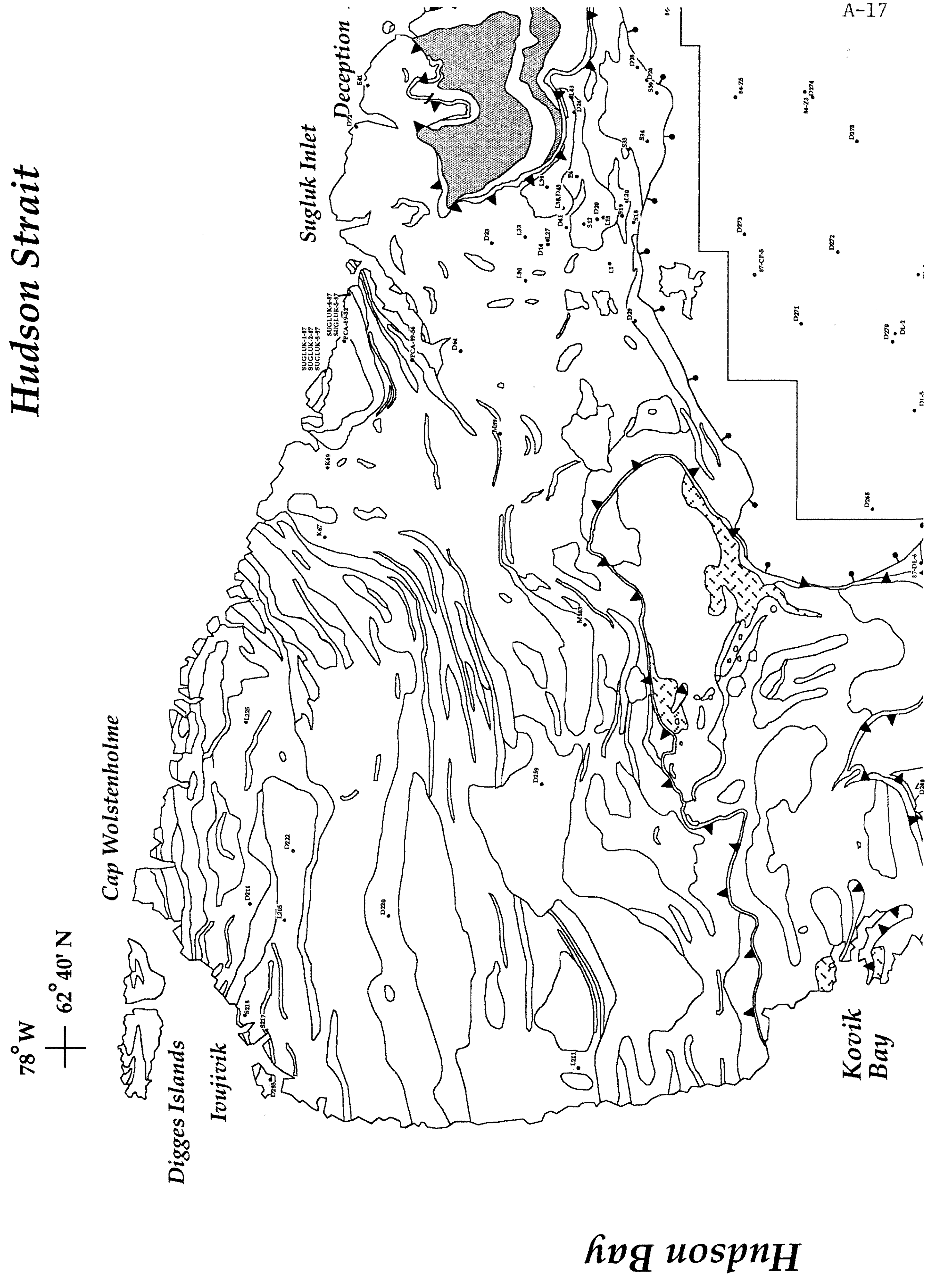
Cape Smith Suite

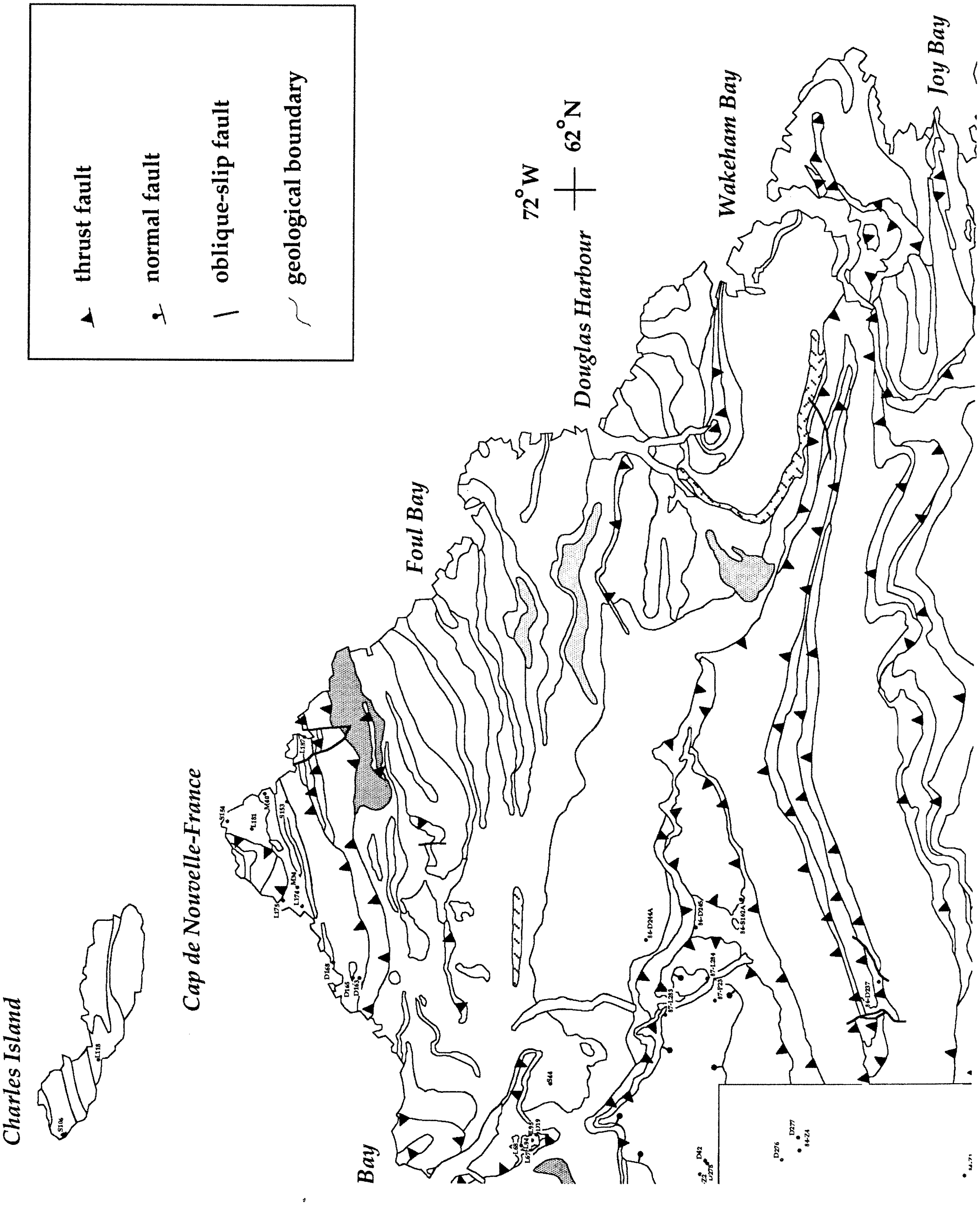
Sample	D-268a	D-272	D-273	D-274	S-195	D2-2	D2-3	D4-2	D5-2	D8-1	D8-2	CP2-7	CP3-3	CP5-3	CP5-4	D-271	D-275	D-277	D-284	D3-1	D3-2	D4-3	CP3-4
Lithology	diorite	diorite	diorite	qtz dior	qtz dior	diorite	qtz dior	diorite	qtz diorite	qtz dior	qtz dior	diorite	qtz dior	qtz dior	diorite	tonalite	tonalite	tonalite	tonalite	tonalite	tonalite	tonalite	tonalite
SiO2 (wt %)	47.07	50.86	50.82	55.80	51.99	48.50	53.40	52.30	59.30	49.30	56.40	48.10	51.60	51.10	56.00	66.51	68.41	59.89	65.29	68.00	62.00	65.40	66.60
TiO2	1.46	0.97	1.14	1.05	0.87	1.67	1.30	0.94	0.56	1.68	1.21	2.05	1.03	0.92	0.70	0.42	0.36	0.50	0.77	0.21	0.44	0.38	0.39
Al2O3	17.46	17.57	15.92	17.54	16.25	16.30	16.40	19.90	17.60	17.80	17.30	17.00	18.40	18.10	17.50	17.13	14.54	18.23	16.36	18.00	18.60	17.90	17.10
Fe2O3	12.90	9.76	11.80	10.76	8.49	13.10	12.90	7.72	9.37	11.50	7.77	14.10	9.52	9.88	7.57	3.60	4.56	5.69	4.94	2.07	6.01	3.21	3.23
MnO	0.19	0.16	0.20	0.29	0.12	0.25	0.31	0.08	0.27	0.15	0.11	0.22	0.15	0.16	0.13	0.05	0.10	0.14	0.07	0.06	0.09	0.04	0.05
MgO	6.14	5.43	6.42	1.78	5.60	4.20	2.43	3.42	0.45	4.29	3.42	4.12	3.97	4.86	3.54	1.20	0.69	1.66	1.61	0.81	1.60	1.12	1.10
CaO	8.73	9.02	9.51	6.81	5.84	8.05	7.60	8.22	5.20	8.43	6.27	8.81	7.41	9.70	7.33	3.29	2.91	6.04	4.67	2.84	5.70	4.15	3.60
Na2O	2.90	3.11	3.31	3.47	4.05	4.23	3.43	4.95	4.57	3.40	4.24	3.59	3.55	4.05	4.66	4.68	3.74	3.91	4.84	4.90	4.51	5.42	4.80
K2O	0.53	1.15	0.74	0.86	0.05	0.53	0.47	0.87	1.47	1.03	0.91	0.34	2.35	0.57	1.57	1.37	2.30	1.99	0.56	3.29	1.06	0.99	1.96
P2O5	0.20	0.25	0.39	0.26	0.15	0.23	0.34	0.33	0.10	0.32	0.21	0.40	0.26	0.34	0.25	0.14	0.10	0.23	0.23	0.06	0.13	0.13	0.10
LOI	2.64	1.98	0.02	1.61	6.43	1.07	1.01	1.88	1.74	1.95	2.09	0.17	2.16	0.03	0.19	1.59	2.02	1.47	0.78	0.56	1.39	1.56	1.70
TOTAL	100.22	100.26	100.27	100.23	99.84	98.13	99.59	100.61	100.63	99.85	99.93	98.9	100.4	99.71	99.44	99.98	99.73	99.75	100.12	100.8	101.53	100.3	100.63
Th (ppm)	nd	nd	nd	2.0	1.4	0.6	1.2	0.6	3.2	2.8	2.4	0.7	1.3	0.3	1.8	nd	nd	nd	3.2	nd	nd	1.1	3.4
U	nd	nd	nd	0.9	0.9	0.5	1.1	0.5	0.0	0.5	1.1	0.5	0.8	nd	1.1	nd	nd	nd	0.6	nd	nd	0.9	1.2
Rb	11.7	41.1	9.1	18.9	2.0	10.0	11.0	11.0	42.0	25.0	23.0	6.0	73.0	nd	31.0	40.7	46.9	56.1	10.4	79.0	25.0	19.0	43.0
Sr	330	683	881	407	167	470	400	1300	460	460	430	490	570	1000	800	776	291	1093	604	871	1100	940	730
Y	27.1	16.3	22.9	32.1	20.8	30.0	28.0	8.0	33.0	35.0	22.0	28.0	24.0	18.0	19.0	5.7	29.3	20.8	12.5	8.0	13.0	6.0	6.0
Zr	109	79	87	110	109	110	53	93	210	200	150	95	120	44	72	151	168	222	285	110	190	140	130
Nb	10.1	5.3	7.9	8.9	9.3	10.0	10.0	0.0	9.0	11.0	9.0	10.0	6.0	nd	4.0	6.6	11.3	13.5	16.1	0.0	0.0	0.0	0.0
Ba	nd	nd	nd	268	nd	116	187	264	322	292	262	155	399	464	581	nd	nd	nd	446	2700	806	343	669
Hf	nd	nd	nd	2.70	2.41	2.90	1.40	2.00	5.10	4.30	3.50	1.80	2.50	0.70	2.00	nd	nd	nd	6.43	nd	nd	2.70	3.00
Ta	nd	nd	nd	0.4	0.4	1.0	3.5	1.6	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.5	nd	nd	3.6	nd
Cr	0.0	0.0	nd	nd		0.0	0.0	nd	nd	nd	nd	44.0	40.0	53.0	56.0	nd	nd	nd	nd	2.0	1200.0	nd	39.0
Ni	44.3	40.8	56.5	4.9	45.9	57.0	4.0	39.0	nd	21.0	19.0	5.0	11.0	38.0	28.0	9.2	3.5	8.0	16.8	2.0	nd	9.0	12.0
Zn	109.4	103.1	119.3	116.5	82.6	88.0	146.0	77.0	127.0	82.0	43.0	143.0	105.0	114.0	91.0	56.3	60.0	87.9	68.1	38.0	69.0	47.0	99.0
Cu	55.9	47.8	155.0	4.8	25.6	80.0	20.0	63.0	11.0	133.0	496.0	53.0	36.0	184.0	100.0	7.8	2.4	13.7	17.0	2.0	24.0	8.0	11.0
Co	nd	nd	nd	9.7	23.4	28.0	18.0	25.0	nd	30.0	27.0	26.0	23.0	30.0	22.0	nd	nd	nd	9.7	5.0	9.0	5.0	6.0
Sc	nd	nd	nd	20.85	25.91	30.00	30.00	21.00	25.00	27.00	20.00	24.00	11.00	23.00	13.00	0.00	nd	nd	7.34	3.00	8.00	4.00	4.00
Cs	nd	nd	nd	0.4	nd	0.3	0.2	0.3	0.8	0.8	0.3	0.2	1.4	0.4	0.7	nd	nd	nd	0.3	nd	nd	0.3	0.5
W	nd	nd	nd	7	14	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	3	nd	nd	nd	nd
Au	nd	nd	nd	0.009	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.008	nd	nd	nd	nd
La	nd	nd	nd	12.64	11.63	12.00	10.00	17.00	16.00	32.00	17.00	13.00	8.00	15.00	15.00	nd	nd	nd	45.41	15.00	28.00	7.00	10.00
Ce	nd	nd	nd	27.76	24.62	29.00	24.00	29.00	35.00	66.00	31.00	30.00	30.00	34.00	40.00	nd	nd	nd	80.34	27.00	52.00	19.00	31.00
Nd	nd	nd	nd	19.48	14.19	19.00	18.00	20.00	18.00	34.00	15.00	18.00	20.00	19.00	21.00	nd	nd	nd	29.26	nd	nd	9.00	14.00
Sm	nd	nd	nd	4.99	2.93	5.10	5.00	4.40	4.70	7.20	4.30	5.40	4.70	4.80	4.80	nd	nd	nd	4.47	nd	nd	2.10	2.60
Eu	nd	nd	nd	1.75	1.03	1.60	1.90	1.40	2.40	2.20	1.00	1.90	1.40	1.50	1.40	nd	nd	nd	1.58	nd	nd	0.70	0.80
Tb	nd	nd	nd	0.82	0.49	1.00	0.80	0.50	0.70	1.10	0.60	0.70	0.60	0.50	0.60	nd	nd	nd	0.35	nd	nd	0.20	0.20
Ho	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Yb	nd	nd	nd	3.24	1.76	3.40	3.10	1.00	3.10	2.70	1.40	2.00	1.80	1.50	1.90	nd	nd	nd	1.30	nd	nd	0.50	0.40
Lu	nd	nd	nd	0.44	0.29	0.48	0.45	0.14	0.52	0.46	0.25	0.36	0.30	0.24	0.27	nd	nd	nd	0.20	nd	nd	0.12	0.07
A/CNK	0.82	0.77	0.68	0.92	0.94	0.74	0.82	0.83	0.95	0.81	0.89	0.76	0.84	0.73	0.77	1.13	1.04	0.93	0.96	1.07	0.98	1.02	1.03
(La/Yb)n	nd	nd	nd	2.8	4.7	2.5	2.3	12.2	3.7	8.5	8.7	4.7	3.2	7.2	5.7	nd	nd	nd	25.1	nd	nd	10.0	17.9
(La/Sm)n	nd	nd	nd	1.6	2.6	1.5	1.3	2.5	2.2	2.9	2.6	1.6	1.1	2.0	2.0	nd	nd	nd	6.6	nd	nd	2.2	2.5

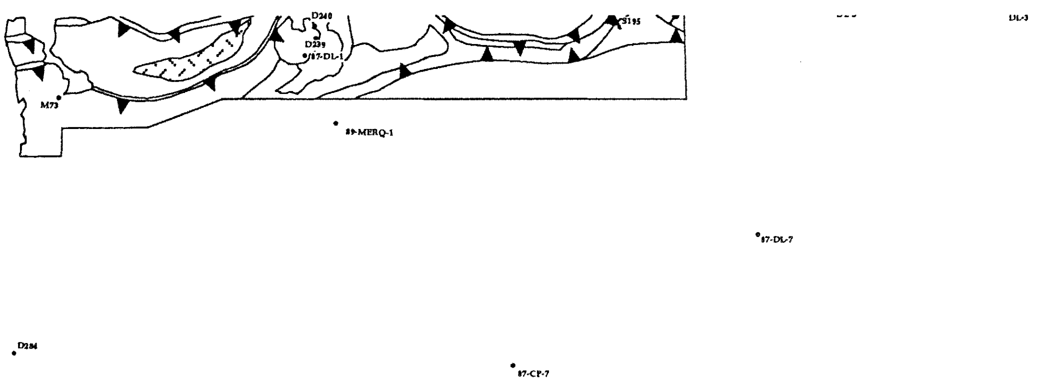
Sample	D-42	D-239	D-240	D-270	D-278	DL1-4
Lithology	granite	grano	grano	grano	grano	grano
SiO_2 (wt %)	72.14	70.60	71.41	71.04	71.27	70.30
TiO_2	0.09	0.28	0.20	0.23	0.10	0.28
Al_2O_3	15.33	16.17	15.89	15.86	16.40	15.70
Fe_2O_3	0.73	2.29	1.75	2.00	0.93	1.99
MnO	0.02	0.03	0.03	0.02	0.01	0.03
MgO	0.15	0.59	0.48	0.36	0.10	0.72
CaO	1.21	2.77	2.32	2.59	1.32	2.47
Na_2O	6.02	5.38	5.06	5.01	6.68	5.12
K_2O	3.05	1.40	2.28	1.51	2.74	2.09
P_2O_5	0.04	0.08	0.07	0.07	0.04	0.07
LOI	0.19	0.46	0.50	0.92	0.16	0.66
TOTAL	98.97	100.05	99.99	99.61	99.75	99.43
Th (ppm)	0.4	nd	2.5	2.0	nd	nd
U	0.3	nd	0.5	0.7	nd	nd
Rb	44.1	38.8	48.7	26.2	33.2	49.0
Sr	1006	842	687	476	1330	620
Y	2.6	6.1	4.1	6.6	2.2	nd
Zr	31	146	109	131	56	120
Nb	7.8	8.6	8.2	9.8	8.8	9.0
Ba	2073	nd	986	559	nd	936
Hf	0.70	nd	2.31	2.97	nd	nd
Ta	1.8	nd	0.4	0.4	nd	nd
Cr	15.8	nd	nd	0.0	0.0	150.0
Ni	6.2	5.8	7.1	3.8	5.6	4.0
Zn	20.2	49.0	44.4	23.0	17.3	46.0
Cu	2.0	2.3	0.9	1.5	2.4	2.0
Co	19.4	nd	2.4	1.9	nd	5.0
Sc	0.59	nd	2.11	1.79	nd	3.00
Cs	0.2	nd	0.5	0.3	nd	nd
W	384	nd	nd	nd	nd	nd
Au	0.002	nd	nd	nd	nd	nd
La	2.49	nd	12.46	10.90	nd	20.00
Ce	5.87	nd	23.53	21.42	nd	29.00
Nd	3.82	nd	4.62	9.16	nd	nd
Sm	0.56	nd	1.68	1.95	nd	nd
Eu	0.17	nd	0.49	0.64	nd	nd
Tb	0.10	nd	0.12	0.21	nd	nd
Ho	nd	nd	nd	nd	nd	nd
Yb	0.35	nd	0.55	0.43	nd	nd
Lu	0.09	nd	0.08	0.08	nd	nd
A/CNK	1.00	1.05	1.06	1.09	1.00	1.03
(La/Yb)n	5.1	nd	16.3	18.2	nd	nd
(La/Sm)n	2.9	nd	4.8	3.6	nd	nd

APPENDIX 4

Sample location map for Narsajuaq terrane samples.


Hudson Strait
Sugluk Inlet
Deception
Cap Wolstenholme
78° W
62° 40' N
Digges Islands
Ivujivik
Kovik Bay
Hudson Bay
SUGLUK-1-87
SUGLUK-2-87
SUGLUK-3-87
SUGLUK-4-87
SUGLUK-5-87
PCA-89-32
PCA-89-54
K69
K67
M69
M103
D64
D29
D23
D24
D26
D28
D72
D159
D211
D222
D220
D273
D271
D272
D274
D275
D268
D270
L225
L205
L211
S216
S217
S218
L33
L39
L90
L27
L43
L18
S12
S19
S33
S34
S39
E41
87-CP-5
84-Z3
84-Z5
87-DL-4
DL-2


thrust fault
normal fault
oblique-slip fault
geological boundary
Charles Island
Cap de Nouvelle-France
Bay
Foul Bay
Douglas Harbour
72° W
62° N
Wakeham Bay
Joy Bay
S106
L118
S154
L181
M40
S153
L187
M34
L174
L175
D168
D165
D163
86-D244A
86-D245
86-S162A
87-L284
87-P23
87-L283
S44
L68
L94
L67
L119
D42
D278
D276
D277
84-Z4
86-D237



Narsajuaq arc

Younger suite

☐ bt ± hbl monzogranite to syenogranite; cpx ± hbl ± bt monzonite

☐ hbl - bt tonalite to granodiorite; bt monzogranite veins

☐ hbl ± bt quartz diorite; hbl - bt diorite to monzodiorite; cpx - hbl - bt pyroxenite; bt monzogranite veins

Older suite

☐ layered bt ± hbl tonalite and hbl - bt quartz diorite; bt granodiorite; bt monzogranite veins

☐ layered hbl-bt quartz diorite and bt ± hbl tonalite; bt monzogranite veins

Late suite

☐ bt monzogranite

☐ syenogranite

Sugluk Group

☐ semipelite, quartzite, ironstone, pelite, marble, calc-silicate rock

Cape Smith Belt

Spartan Group

☐ pelite, semipelite, quartzite, ga

Parent Group

☐ andesite, rhyolite, rhyodacite, fe

☐ basalt, andesitic basalt, mafic tu

Chukotat Group

☐ basalt, gabbro, peridotite

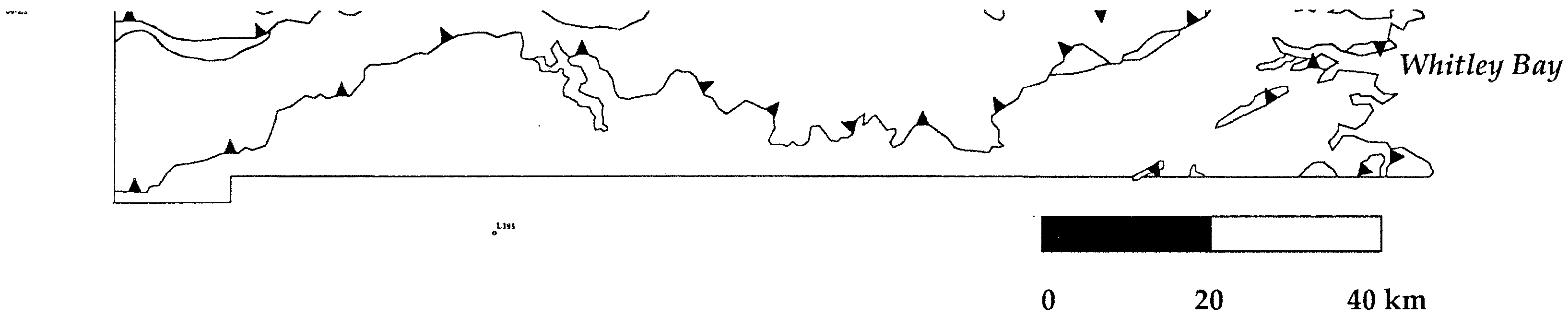
Povungnituk Group

☐ basalt, gabbro, peridotite

☐ semipelite, quartzite, ironstone,

Watts Group

☐ basalt, gabbro, peridotite, pyrox



bro

lsic and mafic tuff and agglomerate

ff

pelite, basalt, gabbro, peridotite

nite

Superior Province (Archean rocks in the Ungava orogen)

- ▢ bt ± hbl syenogranite
- ▢ bt ± hbl monzogranite
- ▨ bt ± hbl granodiorite
- ▢ hbl ± bt tonalite; bt monzogranite veins
- ▒ hbl ± bt tonalite and quartz diorite; bt monzogranite veins
- ▢ bt ± hbl tonalite (rare enclaves); bt monzogranite veins
- ▢ bt ± hbl tonalite (abundant enclaves); bt monzogranite veins
- ▢ semipelite, pelite, marble; bt monzogranite veins
- ▢ hbl - bt quartz diorite; hbl - bt amphibolite; peridotite; pyroxenite; bt monzogranite veins