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**TECTONIC EVOLUTION OF THE BARK LAKE AREA, EASTERN CENTRAL
GNEISS BELT, ONTARIO GRENVILLE: CONSTRAINTS FROM GEOLOGY,
GEOCHEMISTRY AND U-Pb GEOCHRONOLOGY**

by

Sarah Margaret McMullen, B.Sc.

**A thesis submitted to the Faculty of
Graduate Studies and Research in partial fulfillment
of the requirements for the degree of
Master of Science, Department of Earth Sciences**

**Ottawa-Carleton Geoscience Centre
Carleton University**

Ottawa, Ontario

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GNEISS BELT, ONTARIO GRENVILLE: CONSTRAINTS FROM GEOLOGY,
GEOCHEMISTRY AND U-Pb GEOCHRONOLOGY**

Submitted by

Sarah McMullen, B.Sc. (Hons.)

in partial fulfilment of the requirements
for the degree of Master of Science

Thesis Supervisor

Chair, Department of Earth Sciences

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ABSTRACT

Detailed field mapping of lithology and structure was completed at the 1:25 000 scale in the Barry's Bay-Bark Lake region of Ontario, in the SE margin of the Central Gneiss Belt, in the immediate footwall of a crustal-scale ductile shear zone. The area is characterized by upper amphibolite- to granulite-facies, granodioritic orthogneiss that exhibits extensive migmatitic textures and a penetrative transposition foliation. Geochemistry results demonstrate calc-alkaline trends, typical of arc-derived magmas. U-Pb geochronology of 6 samples from the study area yield protolith crystallization ages at ~1480 Ma. Penetrative ductile deformation was coeval with high-grade metamorphism and accompanying anatexis at ca. 1070-1040 Ma. Based on similar age and lithotectonic affinities of the Bark Lake area with the western Muskoka domain, this area is interpreted as representing the eastern extension of the Muskoka domain. These rocks are interpreted to have originated during the Mesoproterozoic in an Andean-type continental arc setting, on the southeast margin of pre-Grenvillian Laurentia

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Fieldwork and U-Pb laboratory procedures were supported by an NSERC operating grant to Dr. Sharon D. Carr, at Carleton University. Sharon is thanked for her support, both academically and financially, in the capacity of supervisor of this thesis. In addition to her patience, encouragement and unwavering enthusiasm, Sharon provided insight into the art of U-Pb geochronology, a dynamic work environment, and a cool office-mate. Mike Easton at the OGS is thanked for his visits in the field (you gotta love those 3 am arrivals!), and for his help with the geochemical aspect of this project, from sample-collection to short-course tutorials. My dear friend and trusty field assistant, Tanya Tettelaar, is thanked for her hard work, patience and humour in the field, and for keeping me "singing on fine-grained", even during the heat-wave. The maps for this thesis could not have been made without the patience and assistance of Chris ("Newfie") Rogers. Jim and Ruth of Sunny Hill Resort on Bark Lake are thanked for providing comfortable accommodations while in the field, for making sure we got home safely every day, and for entertaining us "Barry's Bay-style". Thank-you to all of my colleagues and friends within the department who have shared their time and knowledge with me, and for occasionally laughing at my jokes! In particular, Gibby, Tink and Inch are thanked for keeping me grounded during the "crunch-time" and for help with thesis-collating. I thank my roommate of almost three years, "the other Sarah", for her friendship, encouragement and for more laughs than I can remember. Last but not least, it is with unending love and gratitude that I acknowledge my parents, Sharon and Dave, who, even though they probably don't understand why *anyone* would want to look at rocks all day, are the reason I am who I am.

ORIGINAL CONTRIBUTIONS

During the summer of 1998, two months were spent in the Barry's Bay-Bark Lake region of the Muskoka Highlands, Ontario. The field area comprised the northern two-thirds of NTS map sheet 31 F/5, " Barry's Bay" and the southern margin of NTS map sheet 31F/12, "Round Lake" (North American datum 1983; UTM zone 18). I completed 1: 25 000 scale detailed mapping of lithology and structures, and collected samples for geochemical and U-Pb geochronological analyses. These data were compiled to create 2, 1: 50 000 geology and structures maps for the field area. All samples collected for petrography and analyses were collected by myself and my field assistant, Tanya Tettelaar. Additional help collecting samples for geochronological and geochemical analyses was provided by my supervisor, Dr. Sharon Carr, and by my external examiner, Dr. R. Mike Easton, respectively. Rock cutting and thin section preparation was completed by T. Tettelaar; subsequent grinding and mounting was completed by Mike Jackson. All petrography and selection of rocks for thin section was completed by myself. I completed U-Pb chronometry on four orthogneiss samples of varied granitic composition, and on two cross-cutting syenitic pegmatites. A total of XX fractions were analyzed. All stages of sample preparation for geochronological analyses, including crushing, grinding, mineral separation and picking of fractions was completed by myself at Carleton University. Dr. S. Carr performed the column chemistry and mass spectrometry of the fractions at Carleton University. Data reduction and interpretation was done by myself under the guidance and supervision of Dr. S. Carr. Geochemical analyses were completed at the Ontario Geological Survey (OGS), Sudbury, ON. Interpretation of geochemical data was completed by myself under the guidance of Dr. R.M. Easton (OGS).

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Chapter 1: Introduction

1.1 The Grenville Province

The southeast margin of the Canadian Shield represents the type area for the Grenville Orogen, a once-circumglobal, Meso- to Neoproterozoic collisional orogen that was active during Rodinian supercontinent-assembly (Figure 1.1a; Hoffman, 1991; Davidson, 1995). Wilson (1918, 1925) was the first to recognize the Grenville Province as a structural entity distinct from the rest of the Canadian Shield. This delineation was based on the abrupt transition from predominantly supracrustal rocks, in what are now assigned to the Paleoproterozoic Southern and Archean Superior Provinces, into the high-grade, regionally metamorphosed granitic orthogneiss of variable composition that characterize much of the Grenville Province. Wynne-Edwards (1972) established the Grenvillian Orogeny as a “major mountain building episode”. Further, he recognized that the Grenville Province constitutes a “coherent radiometric province” based on K/Ar ages distinctly younger than those of the Archean foreland (950 ± 150 Ma), and proposed that the present surface had originated from deep within the orogenic belt. The Grenville Orogenic cycle spanned almost 400 Ma, from ~1300-950 Ma. Today, it is generally accepted that the Grenville Province was formed as a result of NW-directed thrusting of a composite belt of continental and/or magmatic arcs onto the distal SE margin of the pre-Grenvillian Laurentian craton. A large part of the present-day erosional surface represents the deeply eroded metamorphic core of the orogen, corresponding to burial depths of ~25-35 km within tectonically thickened continental crust. As such, the Grenville Province provides a unique opportunity to study the tectonothermal

Figure 1.1 Map of the North America, illustrating the tectonic context and location of the Grenville Province (from Easton, 1992).

processes that operated at mid- to deep crustal levels within large collisional orogens with protracted crustal thickening histories.

1.2 Regional Tectonic Framework

Wynne-Edwards (1972), in a preliminary attempt to unravel the superimposed orogenic events in the Grenville, divided the province into seven subdivisions based on their contrasting geological histories and, in part, lithologies. These subdivisions are (i) the Grenvillian Foreland Belt; (ii) the Grenville Front Tectonic Zone; (iii) the Central Gneiss Belt; (iv) the Central Metasedimentary Belt; (v) the Central Granulite Terrain; (vi) the Baie Comeau Segment; and, (vii) the Eastern Grenville Province (Figure 1.2a). Although this terminology has been superseded by the tectonics-based classification of Rivers et al., (1989; Figure 1.2b) and more recently by that of Carr et al. (1999; Figure 1.2c), the classification of Wynne-Edwards (1972) is still used, particularly in Ontario, simply because it uses terms that can be applied in the field. Table 1.1 presents a comparison of terminologies.

The Ontario portion of the Grenville Province comprises two NE-trending, orogen-parallel belts, the Central Gneiss Belt (CGB), and the Central Metasedimentary Belt (CMB) (see Figure 1.2a). These two belts are separated from each other by the highly tectonized northwestern boundary of the CMB, the Central Metasedimentary Belt boundary thrust zone.

1.2.1 The Central Gneiss Belt

The Central Gneiss Belt (CGB) represents the reworked margin of the Pre-Grenvillian Laurentian craton. It consists of predominantly upper amphibolite- to

Figure 1.2. Evolution of Grenvillian terminology. (a) preliminary tectonic framework of Wynne-Edwards (1972); (b) first-order tectonic belts as proposed by Rivers et al. (1989).

Figure 1.2. Evolution of Grenvillian terminology (continued). (c) new tectonic subdivisions of Carr et al. (1999).

(c)

Table 1.1: Comparison of tectonic subdivision terminology for the Grenville Province (after Carr et al., 1999)

Tectonic Sub-divisions (Carr et al., 1999)	Domains or Terranes	Geologic Subdivisions (Wynne-Edwards, 1972)	First Order Tectonic Belts (Rivers et al., 1989)	Interpretations
Laurentia and its margin: L1	-Southern Province -Levack Gneiss Complex -Killamey Magmatic Belt (east of the GF)	Southern Province		-stable orogenic foreland; mainly pre-ca. 1450 Ma rocks that did not undergo Grenvillian penetrative deformation
Laurentia and its margin: L2	-Grenville Front Tectonic Zone -Nepewesset domain -Britt domain -Toniko domain	Central Gneiss Belt (lower structural levels)	Paraallochthonous Belt	-paraallochthonous Laurentian crust; mainly pre-ca. 1450 Ma rocks that were penetratively deformed and experienced NW-directed movement
Laurentia and its margin: L3	-Shawanaga domain -Parry Sound domain (PSD) -Powassan domain -Algonquin domain -Muskoka domain -CMABtz -Bancroft Terrane	Central Gneiss Belt (upper structural levels)	Allochthonous Polycyclic Belt	-(para)allochthonous* rocks including ca. 1450 Ma Laurentian margin and some Archean rocks; these rocks have experienced penetrative Grenvillian deformation and NW-directed transport; the PSD is interpreted as originating in the CAB.
Composite Arc Belt (CAB)	-Belmont domain (Elzevir Terrane) -Gimsthope (Elzevir Terrane) -Mazinaw Terrane -Sharbot Lake Terrane	Central Metasedimentary Belt	Allochthonous Monocyclic Belt	-volcanic arc and related basin rocks, formed and likely assembled at an unknown distance of the SE margin of pre-Grenvillian Laurentia; tectonically juxtaposed against Laurentia during NW-directed thrusting

* In keeping with Culshaw et al. (1997) and Kelchum et al. (1999) the term *allochthonous* refers to any tectonic entity that is interpreted to have been isolated from its source region; it does not imply an exotic origin (i.e., allochthonous terranes may have Laurentian affinities)

granulite-facies plutonic rocks and related supracrustal rocks that have been penetratively deformed, metamorphosed and tectonically telescoped toward the northwest during the Pinwarian (~1500-1450 Ma) and/or Grenvillian orogenies (1300-950 Ma).

Recent workers (e.g., Davidson and Morgan, 1981; Davidson et al., 1982; Culshaw et al., 1983; Hanmer and Ciesielski, 1984) subdivided the the Central Gneiss Belt into many domains based on their contrasting lithologies, metamorphic grade and deformation histories, and proposed that these domains represented imbricated panels of relatively low strain rocks bounded by anastomosing ductile shear zones. Data from geophysical land seismic surveys across the CGB (Abitibi-Grenville Lithoprobe lines 30 and 31, White et al., 1994) is consistent with the proposed crustal architecture.

The CGB contains terranes that are both parautochthonous and allochthonous with respect to the craton, the crustal architecture of which is interpreted have resulted from NW-directed thrusting (see Figure 1.2a). The parautochthonous terranes have lithologic affinities with the orogenic foreland (L1, Carr et al., in press), and are predominantly >1450 Ma, locally containing Archean-aged rocks. They constitute the lowest structural levels of the CGB (e.g., the Nepewassi, Tomiko, and Britt domains) and correspond to the Parautochthonous Belt (PB) of Rivers et al. (1989; L2, Carr et al., in press). The allochthonous terranes (e.g., the Shawanaga, Parry Sound, Powassan, Algonquin and Muskoka domains) were emplaced over the Archean Laurentian crust on the Allochthon Boundary Thrust (ABT; Rivers et al., 1989), a regionally extensive basal decollement whose surface location is poorly known (e.g., Ketchum and Davidson, in press). The allochthonous terranes make up part of the Allochthonous Polycyclic Belt (APB) of Rivers et al (1989) (L3, Carr et al., in press) and consist of voluminous ~1450-

1350 Ma plutonic rocks and related paragneiss, possibly with some minor pre-1450 Ma Laurentian crust. Emplacement of plutons and deposition of supracrustal rocks are interpreted both to have occurred on the distal margin of pre-Grenvillian Laurentia, in or near Andean-type continental arcs (Rivers et al., 1997).

The Central Gneiss Belt is bounded to the NW by the Grenville Front Tectonic Zone (GFTZ; Wynne-Edwards, 1972), a moderately dipping, tectonized zone of predominantly high-grade rocks. Ductile fabrics related to late Grenvillian deformation (ca. 1000 Ma) overprint pre-existing structures, such as the steep brittle faults that define its western-margin (Davidson, 1986; Easton, 1986; Bethune, 1997). The NW boundary of this composite tectonic zone is the Grenville Front (GF), which defines the exposed limit of penetrative Grenvillian deformation and metamorphism within the Precambrian foreland, truncating older structural trends. Geophysically, the GFTZ constitutes a 32 km wide zone of moderately steep, E-dipping reflectors that get progressively steeper with increasing distance from the GFTZ (Green et al., 1989).

1.2.2 The Central Metasedimentary Belt

The CMB is a collage of greenschist- to amphibolite-facies metasedimentary, metavolcanic and minor metaplutonic rocks (Easton, 1992). The CMB rocks are younger than those in the CGB, and they predominantly fall in the range of ~1300-1250 Ma, although ages as young as ~1060 Ma have been obtained for plutonic rocks (Corriveau et al., 1990; Lumbers et al., 1990; Easton, 1992; Davidson, 1995; Rivers, 1997). Supracrustal rocks are interpreted to have formed in volcanic arc and associated basin environments, at some unknown distance off of the pre-Grenvillian margin of Laurentia, and were subsequently amalgamated by ~1240 Ma, prior to their emplacement over the

CGB (Carr et al., in press). On the basis of contrasting lithology, metamorphic and deformation histories, the CMB has been subdivided into the Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes. Along with the CMBbtz, the CMB constitutes the Allochthonous Monocyclic Belt (AMB; Rivers, 1989), also termed the Composite Arc Belt (CAB; Carr et al., in press; see Figure 1.2c). Seismic reflection images in the CMB (Lithoprobe land seismic survey lines 32, 33; see White et al., 1994; Burr and Carr, 1992; 1994) are dominated by panels of moderately to shallowly dipping reflectors with S-SE dips suggesting interleaved crustal slices as a result of NW-directed thrusting (White et al., 1994).

1.2.3 The Central Metasedimentary Belt boundary thrust zone

The Central Metasedimentary Belt boundary thrust zone (CMBbtz; Culshaw et al., 1983) is a composite ductile shear zone that consists of gneissic tectonites derived from both the CGB and the CMB, gneissic tectonites derived solely from the CMB, and rocks from the Bancroft terrane (Easton, 1992; Hanmer, 1988; Hanmer and McEachern, 1992; Burr and Carr, 1992, 1994). Structurally, the CMBbtz is 8-10 km thick, >200 km long, and is characterized by alternating crystalline thrust sheets (horses) enveloped by anastomosing higher strain zones of porphyroclastic to migmatitic tectonites and marble tectonic melange, likely assembled at mid- to deep-crustal levels (Hanmer et al., 1985; Hanmer, 1988; Hanmer and McEachern, 1992; Burr and Carr, 1992; and references therein). This regionally extensive structure has long been recognized as the crustal-scale ductile shear zone along which the rocks of the CMB were tectonically juxtaposed with those of the CGB (e.g., Davidson and Morgan, 1981; Davidson et al., 1982; Culshaw et al., 1983; Davidson, 1984; Hanmer and Ciesielski, 1984; Hanmer et al., 1985). These

authors were the first to recognize the significance of penetrative SE-plunging extension lineations and regionally consistent NW-directed kinematic indicators, in addition to the anastomosing nature of the domain-bounding high-strain zones which accommodated major NW-directed overthrusting.

Lithology in the CMBbtz consists primarily of grey tonalite gneiss thrust sheets bound by quartzofeldspathic (syenitic) straight and porphyroclastic gneiss, interleaved with marble tectonic marble melange, aluminous paragneiss and rare meta-anorthosite (Hanmer et al., 1985). Rheological contrasts between the stiff metagabbroic bodies in the upper part of the CMBbtz and the weak ferro-aluminous pelites and ductile marbles occurring locally at the base of the CMBbtz are interpreted to have played an important mechanical role in the localization and transport of thrust slices in this zone (Hanmer and Ciesielski, 1984; Hanmer et al., 1985; Hanmer, 1988; Hanmer and McEachern, 1992 and references therein). Metamorphic grade in the boundary zone is upper amphibolite-facies ($\geq 650^{\circ}\text{C}$ and 6 kbar; Hanmer, 1988).

Although lithologically distinct from each other, the deformed marble and calc-silicate rocks of the Bancroft Terrane and the gneissic tectonites of the CMB are both characterized by a continuous panel of seismic reflectors, which flatten at mid-crustal levels, in support of field observations suggesting they both comprise the tectonized interface between the CGB and CMB (White et al., 1994; Burr and Carr, 1994). The panel of strong, discontinuous, gently to moderately dipping that characterize the CMBbtz are interpreted as representing ductile thrust ramp-and-cutoff geometries (Burr and Carr, 1994).

It is widely agreed upon that the CMBbtz represents a first-order tectonic entity along which the CMB was emplaced over the CGB, however, the question of the timing of activation and reactivation of this thrust remains under debate. Two models for the timing of thrust initiation predominate in the literature: (1) Marginal basin closure and resultant early thrusting at ca. 1190 Ma, followed by break-back thrusting in the tectonically thickened crust of the footwall, with subsequent reimbrication at ca. 1080-1050 Ma due to intraplate reactivation of the existing, mechanically weak zone (Hanmer and McEachern, 1992; McEachern et al., 1992; Hanmer et al., 1992; McEachern and van Breeman, 1993; Nadeau and Hanmer, 1993); and (2) Thrust initiation at ca. 1190-1080 Ma (Elzevirian phase) with ca. 1020 Ma extensional reactivation (Culshaw et al., 1997; Timmermann et al., 1997; Timmermann, 1998). The timing of thrust initiation is further addressed below.

1.3 This Study

During the summer of 1998, eight weeks were spent in the Barry's Bay region of the Muskoka Highlands of central Ontario. Fieldwork included systematic mapping at 1:25 000 scale, and the collection of fresh, unaltered samples for geochemical and U-Pb geochronological analyses. Mapping was focused on the lakeshores and highways, which constitute the best exposed outcrops in which to study the complex field relationships. Sample locations can be found on Map 1; results of mapping are presented on Maps 2 and 3, located in the back pocket of this thesis.

The transect was a roughly 10km wide x 20km long north-south transect, parallel to Bark and Papineau Lakes, ~10km W of Barry's Bay, Ontario, (Map 1, back pocket).

The transect runs perpendicular to, and crosses, the boundary between the southeastern CGB and the CMBbtz, and thus affords an ideal opportunity to: (1) characterize the lithologies in the immediate footwall; (2) to investigate the nature of the boundary between these two first-order belts; and, (3) to corroborate or amend previous attempts at locating the surface expression of the boundary in this area (e.g. Davidson and Easton, 1994).

1.4 Previous Work

Previous work in the vicinity of the Bark Lake map area, in what is now recognized as the interior of the orogen, extends back to the early mapping of Barlow (1899) and Adams and Barlow (1910), in the Haliburton and Bancroft areas. Renewed interest in the area in the mid-seventies and early eighties yielded reconnaissance-scale mapping and preliminary Rb/Sr geochronology and geochemistry (Appleyard, 1974; Appleyard and Stott, 1975; Bell and Blenkinsop, 1980; Lumbers and Vertolli, 1980; Breaks and Thivierge, 1982; Lumbers, 1982).

1.5 Geological Setting of the Field Area

1.5.1 The Muskoka Domain

The Muskoka domain, which is the highest structural level of the Central Gneiss Belt and constitutes the immediate footwall to the CMBbtz, is of paramount interest to this project. The Muskoka domain was originally defined by Davidson and Morgan (1981), as characterized by amphibolite- to granulite-facies orthogneiss "terrane" with irregular structural trends bounded by NW-trending belts of amphibolite-facies straight

gneiss (Figure 1.3a). Since then, the geographical area of the Muskoka domain has been redefined and expanded, and includes the contiguous Seguin and Moon River domains as its root zone (Nadeau, 1990; Ketchum and Davidson, 1999; Figure 1.3b). This domain is bounded by the underlying Algonquin domain to the north, and by the Parry Sound, Rosseau, Moon River and Go Home domains to the west. Its eastern extent, however, remains ambiguous. Recently, the western Muskoka domain was subject to a comprehensive field, metamorphic, structural and U-Pb geochronological study by Timmermann (1998) near Dorset, Ontario.

Timmermann (1998) investigated the nature of the boundary separating the western Muskoka domain and the underlying Algonquin domain, as well as the boundary separating it from the overthrust CMBtz, and characterized the lithotectonic and age affinities of the domains and their boundaries. The western Muskoka domain is characterized by ~1457-1363 Ma amphibolite-facies migmatitic orthogneiss, predominantly granodioritic in composition, associated with subordinate granulite-facies rocks. Associated with the orthogneiss are pink granitoid, megacrystic granitoid, rare anorthosite and paragneiss (marble tectonic melange), as well as numerous mafic enclaves of varied composition, ranging from amphibolite to coronitic metagabbro and ultramafic rocks. Orthogneiss mineral assemblages and compositions suggest calc-alkaline affinities, typical of collisional tectonic environments. These rocks were affected by coeval penetrative deformation and high-grade metamorphism at 1080-1075 Ma which resulted in widespread anatexis (Timmermann, 1998).

1.5.2 Metamorphic Studies

Annovitz and Essene (1989) completed geothermobarometry across the entire Grenville Orogen, and obtained pressures and temperatures that ranged from 4-6 kbar and 400-600°C in the CMB near Madoc, 8-9 kbar and 560-750°C close to the GFTZ near Sudbury to the NW and up to 10-11 kbar and 800°C in the Parry Sound region. They found that pressures were consistent across the CGB - CMBbtz boundary. In the Muskoka domain near the map area, Annovitz and Essene (1989) obtained pressures and temperatures 9-10.5 kbar and 730-790°C. More recently, Carr and Berman (1997) completed a ~30km transect in the Barry's Bay-Bancroft-Whitney region, which encompasses the current field area, and obtained near-peak pressures and temperatures of 8.9 ± 0.9 kbar and $800 \pm 40^\circ\text{C}$ for 10 samples. Their transect spanned the boundary between the CGB and CMBbtz and, consistent with the findings of Annovitz and Essene (1989), they found similar pressures and temperatures in both the CMBbtz and its footwall, suggesting that any component of vertical displacement on this shear zone occurred prior to thermal equilibration at ~ 30 km depth (Carr and Berman, 1997). The data of Carr and Berman (1997) are only slightly lower than the peak pressures and temperatures of 10-11 kbar and 750-850°C obtained in the western Muskoka domain (Timmermann, 1998). Table 1-2 presents a comparison of P-T-t data.

1.6 Scope of the Project

The current project evolved from the reconnaissance U-Pb geochronology and P-T-t study completed by Carr and Berman (1997). Detailed field mapping, U-Pb geochronology and preliminary geochemical analyses were completed in order to

Figure 1.3. Preliminary subdivisions within the Central Gneiss Belt illustrating early domains: B-Britt; P-Parry Sound; K-Kiosk; A-Algonquin; M-Muskoka (from Davidson and Morgan, 1981). Compare with Figure 1.2c

Table 1-2 : Summary of previous P-T data referred to in this thesis

Authors, year	Region (see Figure 1.3 a-c)	# of Samples	Pressure - Temperature data	Interpretation
Annovitz and Essene (1989)	entire Grenville Province	146 samples	P = 9-10.5 kbar T = 680-840°C	peak metamorphic conditions; P-T-t paths indicate cooling accompanied by decompression
Carr and Berman (1997)	Barry's Bay- Whitney-Bancroft, ON (CCB)	10 samples, across a ~ 30 km transect (eastern extension of the Muskoka domain)	P = 8.9 ± 0.9 kbar T = 800 ± 40°C	near-peak pressures and temperatures; thermal equilibration at ~ 30 km depth
Timmermann (1998)	western Muskoka domain in the vicinity of Dorset, ON	14 samples throughout the western Muskoka domain	P = 10-11 kbar T = 750-850°C	peak pressures and temperatures; P-T-t paths consistent with Annovitz and Essene (1989) showing decompression accompanied by slight cooling

characterize the rock types, and to determine the timing and nature of the high-grade metamorphism and deformation in these rocks. As the timing of the movement on the CMBbtz remains an outstanding question, it was expected that characterization of timing and conditions of high-grade metamorphism and deformation in the immediate footwall, interpreted as the response to crustal thickening due to overthrusting, would help elucidate the timing of thrust initiation.

The purpose of this project, therefore, is: (1) to characterize the geology and structures of the Bark Lake area and their relationship to metamorphism and migmatite-formation; (2) to determine the timing and conditions of high grade regional metamorphism and deformation in the immediate footwall of the CMBbtz in order to further constrain movement on this boundary; (3) to compare age and lithotectonic affinities of this area with others along strike; and, (4) to use lithology, age and geochemical information in order to place constraints on the primary tectonic setting of these rocks to gain understanding of the tectonic evolution of the outboard margin of pre-Grenvillian Proterozoic Laurentia.

Chapter 2. Geology of the SE Central Gneiss Belt, Bark Lake, ON.

2.1 Lithology and field relationships

The predominant lithology in the Bark Lake area is a granodioritic to monzonitic migmatitic orthogneiss, intruded by a variety of mafic and felsic rocks and containing a subordinate amount of quartzofeldspathic paragneiss (Map 2, back pocket). Previous reconnaissance-scale mapping of the area by Lumbers (1982) was based on the mineralogical classification of Buddington (1939); however, in this study lithologies are presented in terms of the classification of Streickeisen (1976). The following lithological descriptions are presented in order of decreasing abundance; mineralogy of units is also presented in order of decreasing abundance. Mineral abbreviations are from Kretz (1983).

2.1.1. Granodioritic to monzogranitic orthogneiss unit

The Bark Lake area is characterized by fine- to medium-grained, leucocratic to mesocratic migmatitic orthogneiss (Figure 2.1a-b). The rocks weather gray to pale brown and pink, depending on the amount of alkali feldspar and mafic minerals in the rocks, and have a distinctive mottled texture which is the result of differential weathering between the contrasting grain sizes of mesosome and leucosome. Colour of fresh surfaces range from dark gray through light gray and pink. Mineralogically, the orthogneiss is highly varied and ranges in composition from monzogranite to granodiorite with lesser amounts of granite, diorite and tonalite (pl-qtz-bt-hbl-kfs $\pm$ grt $\pm$ cpx $\pm$ opx) (Table 2-1; Map 2, back pocket). Mafic minerals comprise up to 35% of the total volume. Accessory minerals include magnetite, zircon, titanite, and apatite. Where biotite exceeds hornblende as the

Table 2-1. Rock types and mineralogy for the Bark Lake area

Rock Type ^{1,2}	Mineral Assemblages ^{3,4}
Paragneiss unit	qtz-pl-bt-kfs-grt ± hbl -+zm
Granodioritic-monzogranitic orthogneiss	
mesosome	pl-qtz-hbl-bt-hbl-kfs ± grt ± cpx ± opx ± chl -zm - int ± opaque (mag?)
leucosome	kfs-qtz-pl + hbl ± bt ± opx -zm -int ± opaque (mag?)
Mafic enclaves	hbl-pl-bt ± qtz ± grt ± cpx ± int
Megacrystic granodiorite	kfs-pl-qtz-hbl ± bt ± grt -zm -int
Biotite-magnetite quartz syenite intrusives	kfs-qtz-pl-bt -mag ± hbl ± chl
Coronitic metagabbro	hbl-cpx-grt-pl ± opx ± qtz
Gabbroic anorthosite	pl-hbl ± bt ± chl
Opx-bearing quartz syenite intrusives	kfs-qtz-pl-opx ± hbl ± chl

1. Rock types are listed in order of interpreted relative age, from oldest to youngest, corresponding to the legend found on Map 2 (back pocket).
2. Rock names are according to Streckeisen (1976).
3. Minerals listed in order of decreasing abundance.
4. Mineral abbreviations from Kretz (1983).

Figure 2.1. Migmatitic granodioritic orthogneiss typical of the Bark Lake area: (a) Strongly foliated mesocratic orthogneiss with cross-cutting pegmatites and concordant leucosome; (b) Mesocratic orthogneiss demonstrating typical field relationships: attenuated and folded mafic enclaves, cross-cutting pegmatites and leucosome concordant with the foliation.

(a)

(b)

Figure 2.1 continued. (c) and (d) illustrate migmatitic textures in this unit: (c) Subconcordant leucosome is tightly to isoclinally folded; (d) Stromatic migmatite (irregularly layered) with ~35% leucosome. Spine of notebook (for scale) is 16 cm long; pencil is 15 cm long.

(c)

(d)

dominant mafic phase, the rock is finer-grained and has a higher mafic content (up to 35%) than where hornblende exceeds biotite (up to 25-30% total mafics). The plutonic origin of the orthogneiss is indicated by the abundance of microcline as the primary albitic phase (Table 2-1) and the regional homogeneity (Map 2, back pocket), an observation further supported by geochemical analyses (see 2.3. *Geochemistry*, below).

Migmatitic textures are pervasive throughout the orthogneiss and paragneiss (Figure 2.1c-d). Leucosome, which can constitute up to 35% of the exposed rock, has the assemblage $kfs\text{-}qtz\text{-}pl + hbl \pm bt$. Hornblende is ubiquitous in leucosomes as large, euhedral porphyroblasts up to 2 cm in length. The occurrence of hornblende porphyroblasts in the leucosome imparts a distinct mottled or bird's-eye texture to these layers. Field evidence such as mafic selvages around leucosome (melanosomes) and preserved igneous plutonic textures within leucosomes suggested *in situ* partial melting for the generation of leucosome, an hypothesis further supported by petrography and geochemical analysis (see 2.3 *Geochemistry*, below; and 3. *Migmatites and Partial Melting*; this thesis).

2.1.2 Mafic enclaves

Mafic enclaves occur everywhere throughout the rocks of the transect. They are fine- to medium-grained and are typically amphibolites with the assemblage $hbl\text{-}plg \pm bt \pm grt \pm qtz \pm cpx$ (Table 2-1). They occur either as highly attenuated and laterally discontinuous cm- to dm- thick layers or lens- to lozenge-shaped boudins that range from cm- to a few meters in size (Figure 2.2a-c). The long axes of the mafic enclaves are generally parallel to the regional foliation (transposition foliation, S_T ; see 4. *Structural geology of the Bark Lake area*, this thesis), and commonly have a poorly to well-

Figure 2.2. Mafic enclaves of the Bark Lake area: (a) cm- to dm-thick attenuated and isoclinally folded amphibolite layers. Note disrupted and laterally discontinuous nature of the enclaves; (b) dm-thick attenuated mafic layer; note pooling of late pegmatitic material in the dilatational site; (c) 1-2 m long mafic boudins in leucocratic pink orthogneiss.

developed biotite and/or hornblende mineral lineation and an internal foliation discordant to that of the host rocks. Contacts between host rocks and the enclaves are sharp and rarely show chilled margins within the enclaves. Disrupted mafic layers and boudins are interpreted to represent deformed dykes which post-dated the intrusion of the host orthogneiss and pre-date deformation and transposition. Together, the mafic rocks and orthogneiss have undergone progressive deformation which resulted in tight to isoclinal folds and the development of the transposition foliation, and the obliteration of primary igneous contact relationships.

Subordinate cm- to $\geq$ dm-scale ultramafic pods and lenses are disseminated throughout the rocks. They are medium- to coarse-grained with 90-100% hornblende and/or biotite and rarely clinopyroxene, with up to 10% plagioclase and accessory minerals. These lozenge-shaped "inclusions" are foliation-parallel and are associated with amphibolite enclaves and selvages around leucosome.

2.1.3. Late Granitic Intrusive Rocks

Biotite – magnetite $\pm$ hornblende –bearing (quartz) syenite. Approximately 10% of the rocks along the transect are quartz syenite that occur as early and late syn-, and post-tectonic pegmatites (Figure 2.3a-c). The pegmatites are bright pink in colour, with dimensions measuring ~30-60 cm wide and extending up to >5m in length (true extent generally obscured by exposure). Quartz syenitic pegmatites are coarse-grained (crystals >2cm long) and biotite-, magnetite- and/or hornblende-bearing. They intrude all lithologies, cross-cutting the transposition foliation of the host migmatitic gneiss; as such, pre- and syn-tectonic pegmatites were extremely useful indicators for mapping deformation intensities across the transect (*see 4. Structural geology of the Bark Lake*

area, this thesis). Post-tectonic pegmatite was preferentially localized into late shear bands (Figure 2.3d).

Less common are medium- to coarse-grained quartz syenite sheets, which intruded the host orthogneiss. These typically occur on the scale of tens of meters in length and are weakly to non-foliated. Syenitic sheets are concordant with respect to the main foliation, and commonly contain biotite, magnetite $\pm$ hornblende, similar to the pegmatites.

Orthopyroxene-bearing pegmatites. Leucocratic orthopyroxene-bearing dykes occur only in the northern part of the transect area at Bell's Bay, Trout and Buch Lakes, and on Highway 60 (Map 2, back pocket). Pegmatites consist of kfs-qtz-opx-mag-bt. Orthopyroxene occurs as porphyroblasts up to 2 cm in length. The pegmatites are late syn-tectonic intrusions that are pale pink to pinkish-white in colour, on the order of 10-30 cm wide and decrease in width along their length. Contacts with the host rock are at a high angle, sharp and regular, and lack chilled margins (Figure 2.3e).

Most of the bt-mag $\pm$ hbl -bearing pegmatites have irregular contacts and are interpreted to be syn-tectonic. In contrast, the opx-bearing pegmatites have sharp, straight contacts and are thus interpreted to post-date both the main (transposition) foliation and the bt-mag $\pm$ hornblende -bearing pegmatites. However, rare, pristine igneous pegmatites of the latter composition (see, for example, pg. 136, this thesis; Sample 3, SM98-28) are the youngest pegmatites in the area.

2.1.4. Paragneiss Unit

Paragneiss is sparse; it comprises only ~5% of exposed rock. Unlike the rest of the rocks in the area, it is of sedimentary origin and interpreted to represent original

Figure 2.3. Pegmatites of the Bark Lake area: (a) Late syn-tectonic pegmatite from the north of Bark Lake, with sharp and regular contacts with granodioritic orthogneiss host. Pegmatite is weakly foliated and lacks chill margins; (b) Late syn-tectonic pegmatite with sharp contacts at a high angle to the foliation. Pegmatite is strongly foliated, with internal foliation parallel to the external foliation and apophyses attenuated along the foliation. Pegmatite from the CMBbtz on Papineau Lake (SM98-83); (c) Post-tectonic pegmatite, with graphitic texture. Note coarseness of crystals (SM98-28; Beavertail Island).

Figure 2.3. continued. (d) Pegmatite localized in late shear band truncating the foliation in a migmatitic straight gneiss; (e) Late, cross-cutting leucocratic, opx-bearing pegmatitic dyke. Spine of notebook is 16 cm long; shaft of rock hammer is ~30 cm long.

(d)

(e)

Figure 2.4. An example of the paragneiss unit: cm- dm-thick, ~100% biotite shear zone (Papineau Lake; person for scale).

country rock into which the protolith of the orthogneiss was intruded. In the field, metasedimentary rocks were distinguished from metaplutonic rocks based on their rusty brown- weathering colour and friable texture as well as an abundance of biotite and garnet compared to the more leucocratic orthogneiss. Garnet porphyroblasts commonly measure up to 3 cm in diameter and are dark red in colour. Except for thin, cm- to <dm-thick ~100% biotite shear zones (Figure 2.4) which occur throughout the orthogneiss, paragneiss enclaves are restricted in the southern part of the transect, and occur in close spatial proximity to the surface trace of the CMBbtz (Map 2, back pocket).

2.1.5. Distinctive Lithologies

The most distinctive lithologies observed in the map-area are the least abundant. These are: (i) megacrystic granodiorite; (ii) olivine coronitic metagabbro, and (iii) anorthositic gabbro. They are significant because they may be correlated with important regional markers, and are useful in discriminating between different paleotectonic environments (see 6. *Tectonic Synthesis*, this thesis).

(Meta-) megacrystic granodiorite (Figure 2.5a-b) comprises ~2% of exposed rock in the area. It is a medium-grained, gray and pink amphibolite-facies orthogneiss with the assemblage plg-kfs-qtz-hbl-bt-grt. It is a strongly foliated rock (S>L tectonite) characterized by distinctive eye-shaped aggregates of K-feldspar and plagioclase $\pm$ quartz, that formed as a result of recrystallization during deformation. Aggregates are up to 2-5 cm long and have a ~3:1 aspect ratio. They represent relicts of original K-feldspar megacrysts which are evidence in support of the plutonic origin of the gneiss.

(Olivine) coronitic metagabbro is extremely rare. A few (<5) potential candidates were observed (i.e., they are completely recrystallized to amphibolite with no

relict igneous texture or coronas visible), and only one 2m x 1m boudin can be definitively labeled "coronitic metagabbro" (Figure 2.5c-d; Map 2, back pocket). This boudin is weakly foliated and in tectonic contact with the host orthogneiss. It is characterized by the assemblage $\text{hbl-cpx-pl-grt} \pm \text{opx} \pm \text{bt}$ (note that no olivine remains), relict coronas with a fine-grained, sugary texture and patchy leucocratic net-veining, which is commonly present in these bodies (Davidson and Grant, 1986). In the field, the boudin is associated with other mafic boudins that are interpreted to represent the completely recrystallized amphibolite equivalent of the initial coronite. Within these related mafics, no vestiges of primary olivine or coronas remain, but the mineral assemblage of hbl-pl-bt-cpx as well as the close spatial proximity (<1m away) to the coronite suggest that these mafic boudins initially comprised a single, larger coherent mass. These bodies commonly occur in clusters, interpreted to represent larger bodies that were tectonically severed during deformation, resulting in numerous smaller fragments (Davidson and Grant 1986; Ketchum and Davidson, 1999). It is believed that further mapping along strike would reveal more of such bodies.

Gabbroic anorthosite also outcrops only once along the transect. It has a unique medium-grained, "tiger-striped" texture that distinguishes it from other rocks (Figure 2.5e). The gneissic layering is defined by thin, discontinuous bands of variable amounts of plagioclase and hornblende that become progressively wider, giving it a progressively striped appearance. The composition is $\text{plg} > \text{hbl-bt} \pm \text{qtz}$. In other parts of the Grenville, gabbroic anorthosite has been documented as being present within shear zones that represent lithotectonic domain-boundaries.

Figure 2.5. Distinctive lithologies in the Bark Lake area: (a) Megacrystic granodioritic orthogneiss (SM98-118); spine of notebook is 16 cm long; (b) close-up of megacrystic gneiss demonstrating detail of k-feldspar aggregates; width of hand-lens is approximately 3 cm. See text for details.

(a)

(b)

Figure 2.5. continued. Distinctive lithologies in the Bark Lake area: (c) Coronitic metagabbro boudin (SM98-166). Boudin is ~1.5m across; (d) Close-up of metagabbro demonstrating detail of net-veining; (e) Gabbroic anorthosite (SM98-121). Note “tiger-striped” texture. Spine of notebook is 16 cm long; hand lens is ~3 cm wide; pencil is 15 cm long.

2.2 Metamorphic Grade

The metamorphic grade is upper amphibolite- to lower granulite-facies. Previous determinations yielded 8.9 ± 0.9 kbar and $800 \pm 40^\circ\text{C}$, pressures and temperatures characteristic of lower granulite-facies (Carr and Berman, 1997; see Table 1-2). These values appear to contradict the field evidence (i.e., from mineral assemblages) which indicates dominantly amphibolite-grade rocks with minor, patchy preservation of higher grade assemblages in restricted areas (Figure 2.6; Map 2, back pocket). Ubiquitous hornblende and/or biotite in the absence of minerals indicative of higher or lower grades, for example orthopyroxene and actinolite, respectively, characterize the main granodioritic-monzodioritic orthogneiss unit, as well as the subordinate rock types. Plagioclase is oligoclase in composition, typical of amphibolite-facies (Carr and Berman, 1997). Garnet porphyroblasts are common in mafic enclaves, quartzofeldspathic paragneiss, and more rarely in the host orthogneiss. Hornblende porphyroblasts are characteristic of the voluminous leucosome that occurs virtually everywhere along the transect. Granulite-facies assemblages in the orthogneiss and pegmatites are characterized by large orthopyroxene porphyroblasts rimmed by hornblende, and are preserved in areally restricted regions: in the north along Highway 60, around Trout Lake, on the highway heading west from Barry's Bay, and in the southernmost part of the transect, on Papineau Island.

The apparent paradox between P-T estimates indicative of a higher grade of metamorphism than is supported by field evidence is interpreted to be a result of widespread retrograde metamorphism, due to the release of water from the crystallizing melts during the accompanying migmatization that produced the voluminous leucosome

characteristic of the transect. This is further discussed in *Chapter 3. Migmatites and Partial Melting* (this thesis).

It should be noted that although there is widespread documentation throughout the CGB for a ~1450Ma, pre-Grenvillian granulite-facies metamorphism (Culshaw et al., 1997; Timmermann, 1998; Ketchum et al., 1999), there is no definitive evidence for such an event in the map-area. The granulite-facies assemblages that are overprinted by amphibolite-facies assemblages are coeval with Grenvillian deformation, and this is interpreted to represent the main metamorphic event.

The existence of opx-bearing leucosomes and pegmatite dykes that cross-cut the foliation suggest a later, local granulite-facies event that post-dated the main (retrograde) amphibolite-facies metamorphic event. The leucocratic dykes and blebby leucosome containing hornblende-rimmed orthopyroxene porphyroblasts suggest a granulite-facies event that post-dates the main (retrograde) amphibolite-facies one. A similar case of late, cross-cutting leucocratic granulite veins and overprinting, opx-bearing leucosomes was documented by Timmermann (1998) in the western Muskoka domain. By analogy to amphibolite-facies granites with diffuse charnockitic alteration documented in southern India, Timmermann (1998) suggested that these late, granulite-facies events may have resulted from dehydration and metasomatism due to influx of CO₂-rich fluids. Alternatively, it was suggested that the granulite-overprint may be a result of limited infiltration of a low $-a_{\text{H}_2\text{O}}$ (water-undersaturated) fluid (Ashworth and Thompson, 1985, and references therein). Post-peak metamorphism influx of low $-a_{\text{H}_2\text{O}}$ fluids in the Bark Lake area may provide a reason for the limited, patchy distribution of granulite-facies

Figure 2.6. Metamorphic grade across the Bark Lake transect.

assemblages observed in post-tectonic dykes and some late leucosomes across the transect. This is obviously a question that demands further attention.

2.3 Geochemistry

2.3.1 Sampling and analytical procedure

Nine samples were collected in the Bark Lake area for geochemical analysis. Samples were analyzed for major, minor and trace elements at the Ontario Geological Survey laboratories in Sudbury, Ontario. Major elements were analyzed by XRF on fused pellet; trace elements by ICP-AES and XRF; and REEs by ICP-MS. A detailed account of the methods used at the OGS analytical laboratories is given in VanderVoet and Riddle (1993). The locations of the samples collected can be found in Figure 2.7; results are presented in Appendices A, B and C and Map 1 (back pocket).

The samples were chosen because they are representative of the rocks along the transect; seven of the nine are granitoids while the remaining two are a fine-grained, undeformed mesocratic dyke (SM98-88D), and an amphibolite (SM98-106). The dyke was chosen, although it is anomalous with respect to the rest of the rocks in the Bark Lake area, to test the hypothesis that it is correlative with a suite of young, post-Grenvillian trachyandesite dykes (M. Easton, pers. comm., 1998). Based on preliminary geochemistry and initial field observations, this appears to be the case. The amphibolite sample was chosen because it is representative of the abundant mafic rocks which are intercalated with the granodioritic orthogneiss throughout the transect. With both of these samples however, it is difficult to interpret their geochemical signatures (beyond a

Figure 2.7 Map of Barry's Bay and Bark Lake illustrating sample sites for geochemical analyses.

broad chemical classification) based on single samples; it is therefore recommended that further sampling be carried out.

The remaining seven samples are representative of the granodioritic-monzodioritic orthogneiss that characterize the Central Gneiss Belt in the Bark Lake area. They are fine- to medium-grained, mesocratic, and have experienced moderate to high strain. In the instance of migmatitic textures (e.g., SM98-23, SM98-69), both meso- and leucosome were collected and analyzed separately in order to compare the geochemical signatures of melt and protolith.

2.3.2. Results

Major Elements. Harker variation diagrams were constructed for K_2O , MgO , CaO , and Al_2O_3 (Figure 2.8a-d). The Bark Lake samples have variable K_2O (2-3.5%) and show no correlation with SiO_2 within the sample suite. MgO , CaO and Al_2O_3 all show negative correlations with SiO_2 ; that is, their concentrations decrease with increasing SiO_2 . On Shand's Index (Figure 2.9a) which constitutes the balance of alkalis, all samples plot in the peraluminous domain [$A/CNK > 1.0$ (molar), $A/NK > 1.0$ (molar)].

Trace Elements. Harker variation diagrams were constructed for the trace elements Ba, Rb and Sr (Figure 2.10a-c). While no correlation exists between Ba and SiO_2 , Rb and Sr show slight positive and strong negative correlations, respectively. On the trace element tectonic affinity diagram of Pearce et al. (1984; Figure 2.11), the Bark Lake granitoids plot in the VAG (volcanic arc granitoid) field. For the most part, the samples lie well within this field; however, some (SM98-118, SM98-23T) lie on the boundary with WPG (within plate granites).

Figure 2.8. Harker variation diagrams illustrating the variation in weight % of selected major element oxides (a) MgO ; (b) Al_2O_3 ; (c) CaO ; and, (d) K_2O , with increasing SiO_2 (weight %). See text for discussion.

Figure 2.9. (a) Shand's index for aluminous character. All samples from Bark Lake plot in the 'peraluminous' field; (b) Chondrite-normalized REE plots. See text for discussion.

Rare Earth Elements. Chondrite-normalized REE plots were constructed for the samples. Normalizing factors used are those of Nakamura (1974) with modifications by G.Jenner at the Memorial University of Newfoundland analytical laboratories (Appendix D). REE patterns for the Bark Lake samples (Figure 2.9b) are typical of granitic rocks and demonstrate LREE enrichment relative to the HREE. Two samples in the suite, SM98-69 (both mesosome and leucosome fractions) and SM98-112, show slightly lower-than-average values for HREE.

Normative Calculations. Both CIPW norms and modified Barth-Niggli mesonorms were calculated for the samples (Table 2-2). In light of the complicated metamorphic and deformational history of these rocks, Barth-Niggli mesonorms were calculated because they take water into consideration, thus allowing the calculations to accommodate metamorphic mineral assemblages. SM98-69 (meso- and leucosome) and SM98-112 have normative corundum values >1%, and 0% normative diopside for both CIPW and Barth-Niggli calculations; all other samples have 0-1% normative corundum and positive normative diopside.

2.3.3 Significance of Results

This preliminary account of the geochemistry constitutes the first published data for the rocks in this area. Although further sampling and analysis is required, some conclusions may be drawn:

The variation in the K_2O vs. SiO_2 Harker variation diagram is to expected from multiply deformed rocks, due to the extreme mobility of K_2O when in solution. The lack of a distinct signature for rocks containing 1-3% K_2O is a common problem (Ciesielski and Sharma, unpublished data, 1993), however this variation diagram is useful for

Table 2-2: CIPW and Barth-Niggli Norms (Selected Minerals; wt%)

CIPW Sample	Diopside		Barth-Niggli		Biotite	Hornblende
	Corundum	Corundum	Corundum	Corundum		
SM98-23T(m)	0.00	0.00	0.00	8.20	6.71	
SM98-23T(l)	0.00	0.04	0.12	12.35	0.00	
SM98-69(m)	0.00	1.79	1.86	8.78	0.00	
SM98-69(l)	0.00	1.68	1.76	6.92	0.00	
SM98-112	0.00	0.97	1.03	1.63	0.00	
SM98-118	0.65	0.00	0.00	8.65	1.90	
SM98-106	9.75	0.00	0.00	1.01	37.22	
SM98-88D	3.93	0.00	0.82	12.06	20.51	
SM98-129	0.65	0.00	0.00	17.88	2.59	

Figure 2.10. Harker variation diagrams illustrating the variation in concentration (ppm) of trace elements (a) Rb; (b) Ba; and, (c) Sr with increasing SiO₂ (weight %). See text for discussion.

(a)

(b)

(c)

Figure 2.11. Pearce et al. (1984) trace element tectonic affinity diagram for granitoid rocks. See text for discussion.

making a preliminary distinction between oceanic plagiogranites and granitoids from all other environments (Maniar and Piccoli, 1989). The diagram indicates that, as expected on the basis of petrography and field observations, the Bark Lake granitoids do *not* constitute oceanic plagiogranites.

The Al_2O_3 vs. SiO_2 variation diagram is useful in discriminating between Group I (IAG: island arc granitoids; CAG: continental arc granitoids; and CCG: continental collision granitoids) and Group II granitoids (RRG: rift-related granitoids; and CEUG: continental epierogenic uplift granitoids), as defined by Maniar and Piccoli (1989). The Bark Lake samples fall well within the domain occupied by Group I granitoids (>12 wt% Al_2O_3 , ≤ 75 wt% SiO_2). In addition, on $\text{FeO}^*/[\text{FeO}^* + \text{MgO}]$ Vs. SiO_2 , AFM (wt%) and ACF (wt%) diagrams of Maniar and Piccoli (1989; Figure 2.12a-c) all samples consistently fall within the domains typical of Group I granitoids (see Figure 2.12 for Group I granitoid ranges). Following the tectonic discrimination scheme for granitoids outlined by Maniar and Piccoli (1989), we can deduce that the granitoid gneisses that comprise the Bark Lake area originated in either an arc or a continental collision environment (Group I). Table 2-3 compares selected parameters defined for the Bark Lake area with those documented from granitoids of known tectonic environments. The Bark Lake values are very similar to those obtained for continental arc granites (CAG).

The granitic rocks of the Bark Lake transect are further classified using Irvine and Baragar (1971) ternary plots (Figure 2.13a-b). On both of these diagrams, all seven granitoid samples cluster within the calc-alkaline domains. Calc-alkaline signatures are expected for rocks originating in an arc environment. However, it should be noted that although the samples *cluster* in the calc-alkaline field, they do not represent a

Table 2-3 Comparison of selected parameters from Bark Lake, Ontario
with granitoids from known tectonic environments from Maniar & Piccoli (1989)

<i>Maniar & Piccoli (1989): Chemistry of granitoids by tectonic environment</i>									
<i>Orogenic</i>					<i>Anorogenic</i>				
Bark Lake	IAG	CAG	CCG	POG	RRG	CEUG	OP		
Silica Range (wt%)	56-76	60-68	62-72	70-76	70-78	72-78	71-77	61-78	
	unimodal	unimodal	unimodal	unimodal	unimodal	60-63	60-62	unimodal	
Alkali-lime Index	calc-alkaline	calcic to calc-alkaline	calc-alkaline	calc-alkaline to alkali-calcic	alkali-calcic	alkalic	alkalic	calcic	
Shand's Index	> 1.0	predominantly peraluminous	metaluminoous- peraluminous	peraluminous	peraluminous (minor)	peraluminous (minor)	peraluminous	peraluminous	
	peraluminous	peraluminous	peraluminous		metaluminoous peralkaline (minor)	metaluminoous peralkaline	metaluminoous	metaluminoous	
Na₂O / CaO (wt%)	< ~3.5	~1.0	~<4.0	~2.0 - 10.0	~2.0 - 18.0	~2.0 - 25.0	~1.0 - 12.0	~<4.0	
Na₂O / K₂O (wt%)	~0.4 - 3.0	~0.4 - 3.0	~4.0 - 2.0	~0.4 - 1.5	~0.6 - 1.2	~0.7 - 1.0	~0.6 - 1.0	0.0 - 50.0	
MgO / FeO* (wt%)	0.27 - 0.47	0.3 - 0.85	0.10 - 0.50	0.05 - 0.6	0.02 - 0.30	0.0 - 0.20	0.0 - 0.12	0.0 - 0.70	
MgO / MnO (wt)	0 - 27	12.0 - 28.0	2.0 - 38.0	2.0 - 45.0	2.0 - 18.0	0.0 - 7.5	0.0 - 7.5	0.0 - 50.0	
Al₂O₃ / Na₂O+K₂O (molar)	> 1.1	> 1.5	> 1.1	> 1.1	0.9 - 1.4	< 1.15	< 1.15	> 1.0	

compositional range so much as an end member composition. To confidently say that these rocks are part of a calc-alkaline suite, further sampling is required.

On the trace element tectonic affinity diagram of Pearce et al. (1984), Figure 2.11, the samples plot in the VAG field, suggesting calc-alkaline affinities. The proximity of certain samples to the boundary with WPG likely represents a component of crustal contamination at the source of these rocks. This is further implied when looking at the REE patterns for SM98-69 (both the mesosome and leucosome) where there is notable depletion of HREE in comparison to the LREE (see Figure 2.9b), as well as when the data is plotted on Shand's index. On this diagram (see Figure 2.9a) all samples are highly peraluminous, but predominantly plot outside the fields typical of both CAG and CCG. This is attributed to Al-enrichment as a result of crustal contamination.

Of note are the samples with >1% normative corundum and no normative diopside (SM98-69 and SM98-116). Typically, this is indicative of S-type granites. These samples also exhibit abundant biotite, contain garnet and contain relatively low Na₂O and high SiO₂ in conjunction with molar A/CNK values exceeding 1.1, again characteristics typical of S-type granites (Chappell and White, 1974). It must also be noted, however, that normative calculations are intended for use with fine-grained, unmetamorphosed volcanic rocks and their applicability to multiply metamorphosed and deformed plutonic rocks, such as those at Bark Lake, is ambiguous at best (M. Easton, pers. comm., 1999). The absence of sedimentary inclusions and minerals such as muscovite and aluminosilicates suggests the small amount of normative corundum may simply be a product of the mesonorm calculation. Nonetheless, the presence of S-type granites would by no means preclude an arc environment.

Figure 2.12. Maniar and Piccoli (1989) tectonic discrimination plots: (a) $\text{FeO}_{(\text{T})} / [\text{FeO}_{(\text{T})} + \text{MgO}]$ Vs. SiO_2 (wt%); (b) ACF (wt%). In this diagram, all samples from the Bark Lake area fall at the lower end of the Group I granitoid range (0-40 wt. % $\text{FeO}_{(\text{T})} + \text{MgO}$, and 0-28 wt.% CaO), well below the range of Group II granitoids; (c) AFM (wt%). In this diagram, all samples from the Bark Lake area fall at the lower end of the Group I granitoid range (0-30 wt. % MgO, and 0-35 wt.% $\text{FeO}_{(\text{T})}$), well below the range of Group II granitoids.

Legend: □ SM98-23T (m)	● SM98-69 (m)	■ SM98-112
○ SM98-23T (l)	X SM98-88	□ SM98-118
● SM98-69 (l)	◇ SM98-106	◆ SM98-129

Figure 2.13. Irvine and Baragar (1971) ternary plot. Samples from Bark Lake plot in the calc-alkaline field and define a trend of increasing alkali content. See text for details.

The similarity of REE patterns obtained for both meso- and leucosomes (see Figure 2.9b) supports the field-based hypothesis of *in situ* anatexis as the mechanism that generated the melt.

In conclusion, preliminary geochemical findings and interpretations are consistent with previous hypotheses based on mineral assemblages and compositions (e.g., Timmerman, 1998) that proposed the origin for these meta-plutonic rocks within an Andean-type continental arc-setting.

2.4 Discussion

The rocks in the Bark Lake area consist of granitic orthogneiss of variable composition with ubiquitous migmatite texture, subordinate amounts of garnetiferous paragneiss, as well as distinctive lithologies such as coronitic metagabbro, megacrystic granodiorite, and gabbroic anorthosite. Although locally heterogeneous, the lithology across the transect exhibits map-scale homogeneity with respect to both compositional and textural variation. The distinctive lithologies are volumetrically insignificant, yet their presence are important in the terms of the regional tectonic framework. The presence of coronitic metagabbroic bodies has been noted throughout the CGB of Ontario (e.g., Davidson and Grant, 1986; Ketchum and Davidson, 1999). Based on the similarities in mineralogy, textural and field relationships of this boudin and those reported by other workers throughout the Central Gneiss Belt in Ontario (Davidson and Grant, 1986; Davidson and van Breeman, 1986; Davidson, 1991; Timmerman, 1998; Ketchum and Davidson, 1999), this outcrop is interpreted as belonging to the same suite of coronites, and therefore of similar age. Coronites throughout Ontario have been

isotopically dated yielding igneous crystallization ages ca. 1170 Ma (U-Pb baddeleyite; Ontario CGB; Davidson and van Breeman, 1988, 1990) with Grenvillian metamorphic ages in the range of 1060-1045 Ma (U-Pb zircon; Davidson and van Breeman, 1988, 1990). The presence of these bodies in the Bark Lake area and their interpreted correlation are indicative of an allochthonous origin for these rocks.

The presence of meta-megacrystic granodiorite in the map-area is important for two reasons: (1) these bodies are important regional markers in that, although volumetrically significant, they are restricted in distribution to the upper structural levels of the CGB, and thus to rocks of craton-margin affinity (Carr et al., 1999). These rocks have emplacement ages of 1460-1430 Ma (van Breeman et al., 1986; Nadeau and van Breeman, 1998) in the western CGB and similar ages of 1444-1466 Ma and 1484^{+24}_{-13} Ma in the McClintock assemblage of the Algonquin domain (structural level 4 of the CGB; Timmermann, 1998); (2) emplacement of these volumetrically significant plutonic rocks into the pre-Grenvillian crust was both spatially and temporally associated with the 1450-1430 Ma granulite-facies metamorphism documented throughout most of the CGB, and it has been suggested that their emplacement may have been the driving mechanism (heat source) for this event (Culshaw et al., 1997; Timmermann, 1998; Ketchum et al., 1999).

Chapter 3. Migmatites and Partial Melting

3.1 Introduction

Migmatitic textures are pervasive throughout the Bark Lake area. Peak pressures and temperatures reached 8.9 ± 0.9 kbar and $800 \pm 40^\circ\text{C}$, respectively (Carr and Berman, 1997; see Table 1-2), characteristic of lower granulite-facies metamorphism. Temperatures in this range are typical for tectonically thickened crust ($T \leq 850^\circ\text{C}$; England and Thompson, 1984; Thompson and England, 1986), and exceed the fluid-saturated melting curves for granites (Timmermann, 1998, and primary references therein). The Bark Lake area makes up a only small part of the eastern Central Gneiss Belt, which has experienced varied degrees of migmatite development. The transect exhibits ubiquitous development of granitic to granodioritic leucosome, which may comprise up to 35% of the volume of outcrop.

3.2 Migmatite nomenclature

Migmatites are common in medium- to high-grade metamorphic terrains. “Migmatite” is a non-genetic term for composite rocks containing metamorphic and/or igneous components. Migmatites may be derived either by open-system processes such as metasomatism and magmatic injection, or by closed system processes such as in situ partial melting (anatexis). The terminology used throughout this paper to describe the various components of migmatites is that of Mehnert (1968), with modifications by Ashworth (1985) (Figure 3.1a).

Migmatites are made up of three distinct textural and mineralogical domains: the leuco-, meso-, and melanosomes. The leucosome comprises the palest coloured part, generally of granitic to granodioritic composition, that is the “melt” component of the mixed rock. Melanosomes may or may not be present; these are mafic-rich selvages at the contact between the leuco- and mesosomes, which represent the residuum of the melting of the parent rock. The presence of melanosomes is considered definitive evidence for *in situ* partial melting (anatexis; Sawyer, 1996). Collectively, the leuco- and melanosome make up the neosome, or the newly formed part of the migmatite. In contrast, the mesosome (or, more rarely the “paleosome”), represents the protolith or parent-rock. It is that part of the migmatite that has not undergone anatexis and retains the appearance of metamorphic rock. Typically it is intermediate in colour between the pale leucosome and dark melanosome, and finer-grained than the leucosome with which it coexists.

Two distinct phases exist within migmatite terranes: a higher grade *diatexite* part and a lower grade *metatexite* part (Sawyer, 1996). In metatexites, structure within the mesosome (“*paleosome*” of Sawyer, 1996) is undisturbed and it maintains a metamorphic texture; i.e., the bulk of the rock has not melted and it behaves rheologically as a solid rock. In contrast, the mesosomes of diatexites show widespread modification of its former metamorphic texture in addition to the development of flow structures. Diatexites develop a more pervasive melt fraction, thus having the rheology of a magma as opposed to solid rock. The migmatites in the Bark Lake area are metatexites.

Figure 3.1. (a) Schematic illustration of a migmatite, illustrating the leucosome, mesosome and melanosome. (b) Example of a late, overprinting leucosome. This photograph is particularly interesting because the porphyroblasts are orthopyroxene rimmed by hornblende. Shaft of hammer is 30 cm long.

3.3 Field Observations of Leucosomes

In the field, leucosome was distinguished from their associated mesosomes based on their coarse grain size, pale colour, ubiquitous euhedral porphyroblasts of hornblende and blebby to concordant contact relationships. These observations were generally reliable discriminators regardless of rock type, although relatively mafic “protoliths” (e.g., diorite) appear to have developed a smaller volume of leucosome. Migmatites are predominantly stromatic (layered) but irregular nebulous migmatite also occurs. Two generations of anatectic leucosome were distinguished in the field, **AL₁** and **AL₂** (see Figure 2.1c-d). Sub-concordant to concordant, attenuated to highly attenuated leucosome that was part of **S_T** (i.e., located parallel to pre-existing planes of anisotropy) is interpreted as the oldest generation of anatectic melt and designated **AL₁**. Blebby, discordant, cross-cutting leucosome that was localized (injected?) into late extensional shear bands and other dilatational structures is designated **AL₂**. Common features of both generations include their similar granitic to granodioritic mineralogy including hornblende porphyroblasts, paler colour and coarser grain size relative to coexisting mesosome, as well as an igneous plutonic texture. In **AL₁**, despite penetrative deformation, leucosome is still slightly coarser-grained than its coexisting mesosome. In concordant **AL₁** leucosomes, the porphyroblasts demonstrate only a weak alignment parallel to the mineral and stretching lineations.

Late, opx-bearing leucosome (**AL₂**) appears to overprint the amphibolite-facies orthogneiss at rare outcrops along Highway 60 in the northern part of the map-area, and west of Barry’s Bay (Figure 3.1b; see Figure 2.6). Porphyroblasts of opx are large (1-2

cm in length) and rimmed by hornblende. Overprinting relationships suggest that this is the result of a later granulite-facies event.

From microtextural relationships and field observations, it is apparent that metamorphism was synchronous with deformation; however, deformation outlasted the early, higher-grade stages of metamorphism and formation of AL_1 , resulting in early leucosomes that are flattened and concordant with respect to the transposition foliation. Post-deformation metamorphism resulted in a second, structurally controlled, generation of anatectic leucosome (AL_2) that cross-cuts S_T and AL_1 , and is localized in late extensional structures such as shear bands.

3.4 Petrography

Mesosomes. The mesosome component (Figure 3.2a-b) generally comprises around 65-70% of the outcrop volume, and rarely exceeds 85% mesosome (100% mesosome = “non-migmatitic precursor”). Compositions range from granitic and granodioritic through monzodioritic and rarely, dioritic. The typical mineral assemblage found in mesosomes is $pl\text{-}qtz\text{-}kfs\text{-}hbl\text{-}bt\text{-}zrn\text{-}tnt\text{-}apt \pm ms \pm grt \pm cpx \pm opx$. Plagioclase is the dominant feldspar; it occurs as (sub-)polygonal recrystallized grains and more rarely as larger laths. Pericline and albite twinning is ubiquitous and deformation twins are also common. There is no optically detectable compositional zoning in the plagioclase, which is oligoclase in composition. Sericite alteration is also common. Quartz generally occurs as fine to medium recrystallized (sub)polygonal grains, but larger relic grains that escaped recrystallization retain evidence for deformation and recovery processes, and have subgrains and undulatory extinction. K-feldspar is subordinate to plagioclase in the

mesosomes; it typically has microcline and perthitic exsolution and like plagioclase and quartz, has generally been recrystallized to fine polygonal grains. Hornblende is the dominant mafic mineral in the mesosome. It is pleochroic with colours ranging from bright green to straw-brown; crystals are eu- to subhedral and define the foliation. It rarely shows significant alteration, but is observed breaking down to biotite; it may also be the product of garnet- and opx-breakdown. Biotite is also common in mesosomes; it is dark brown and pleochroic to lighter brown. It too defines the foliation which commonly wraps around porphyroblasts of garnet and plagioclase. It is generally decussate with a bimodal distribution. There are two phases of biotite growth; older biotite is pre- to syn-tectonic, and rich in zircon inclusions with accompanying pleochroic haloes. Secondary biotite commonly formed as the product of garnet- and/or hornblende-breakdown reactions, and overgrew the pre-existing foliation, thus “crossed micas” is a common observation. Garnet occurs as relicts of large porphyroblasts that are fractured and often broken down to biotite or hornblende. Typically, they exhibit two phases of growth; the first is a (quartz $\pm$ opaque) inclusion-rich core, the second is an inclusion-poor to -absent rim. Rare snowball garnets indicate syn-kinematic growth. Garnets are commonly wrapped by biotite sheaths. Pyroxenes are rare; clinopyroxene is pleochroic green, poorly preserved and always associated with hornblende. Orthopyroxene occurs as relic porphyroblasts with strong green to pink pleochroism. Commonly it has been broken down to give hornblende and plagioclase, and disseminated along the foliation. Accessory minerals include ubiquitous zircon, titanite and apatite. Opaque minerals (magnetite?) occur as inclusions in garnet and hornblende and also as interstitial blebs.

Leucosome. Leucosome comprises up to 35% of the volume of the migmatites at Bark Lake (Figure 3.2c-d). They are granitic to granodioritic in composition, with the typical assemblage $kfs\text{-}qtz\text{-}pl\text{-}hbl \pm bt \pm opx \pm ms$. K-feldspar is the dominant feldspar, in contrast to the plagioclase-dominated mesosomes; it occurs as coarse relic porphyroclasts with fine recrystallized grain boundaries, yielding a mortar texture. Microcline exsolution is ubiquitous. Plagioclase is commonly interstitial to ksp-porphyroclasts, as polygonal recrystallized grains. Pericline and albite twins are present; deformation twins are common in unrecrystallized grains. Sericite alteration of is particularly common to plagioclase. Myrmekitic quartz+plagioclase exsolution is ubiquitous; this is a primary crystallization feature indicating crystallization from a melt (Ashworth and McLellan, 1985). As in the mesosome, oscillatory zoning is conspicuously absent. Quartz occurs as medium recrystallized grains interstitial to large K-feldspar crystals and as a constituent of myrmekite exsolution. Rarely, it is observed as quartz-films around minerals such as hornblende and biotite. Plagioclase, quartz and microcline typically comprise the matrix of the leucosome, interstitial to coarse microcline and mafic porphyroblasts. Hornblende is the most common mafic mineral, present as coarse, euhedral porphyroblasts. Biotite is often a minor component, and occurs as late porphyroblasts that overgrow the plagioclase. Orthopyroxene is a rare constituent; when present, it has pink-green pleochroism and comprises relic porphyroblasts with hornblende rims. Opx-bearing leucosomes are rare, and occur as discordant, overprinting bodies. Muscovite is present only as a late alteration product, and most commonly as fine-grained sericite alteration of plagioclase. Zircon and apatite are common accessory minerals.

Melanosome. The melanosome makes up a small percentage of the migmatites; in fact, they are poorly developed. Where observed, melanosome consists of dark brown biotite concentrated at the boundary between leucosome and mesosome, in a matrix of plagioclase and quartz. It exhibits a strong foliation, with bimodal *distribution* of biotite.

Summary of Petrography. Characteristics common to both meso- and leucosome include: (i) the absence of compositional zoning in the feldspars; (ii) substantial recrystallization of quartz, plagioclase and k-feldspar; (iii) sericitization of plagioclase.

Characteristics distinguishing mesosomes from leucosomes include (i) abundant myrmekite exsolution in leucosome; (ii) leucosomes are medium- to coarse-grained and always coarser grained than their coexisting mesosome; (iii) Mesosomes are richer in hornblende and biotite and thus more strongly foliated than the leucosomes; mesosomes are fine- to medium-grained.

3.5 Discussion

The Bark Lake area is migmatitic throughout; as stated above, leucosome comprises up to 35% of the volume of migmatitic rocks. A question to answer then, is what is the mode of formation of this voluminous leucosome? A rigorous study of migmatites and leucosome-generation is beyond the scope of this project, however, some preliminary inferences can be made based on field and microtextural observations.

Closed system, or *in situ*, partial melting (anatexis) is proposed for the origin of the Bark Lake migmatites. This conclusion is based on: (i) presence of melanosome (melt residuum); (ii) primary replacement textures, such as myrmekitic exsolution and other symplectite textures, as well as rare quartz films. Symplectic intergrowths such as

Figure 3.2 Photomicrographs of mesosome. (a) Hornblende is the dominant mafic mineral in this sample; note the perfect cleavage at 56° and 124° . The other minerals in the photo are plagioclase (dominant light coloured phase), quartz (top left corner), minor biotite (slightly altered, top right corner) and opaques (right-centre of photo). Width of view is 1mm; photo taken under crossed polars with 10x objective; (b) Biotite is the dominant mafic mineral in this sample; note its decussate texture. Other minerals in this photo are plagioclase (light), altered hornblende (bottom left corner). Width of view is 1 mm; photo taken under plane polarized light with 10x objective.

Figure 3.2. continued. Photomicrographs of leucosome. (c) Leucosome composed dominantly of microcline (tartan twinning); other minerals in this photo include plagioclase (note sericitization) and quartz (bottom right corner). Width of photo is 1 mm; photo taken under crossed polars with 10x objective; (b) Leucosome with coarse microcline (tartan twinning) and myrmekite intergrowths (plagioclase and quartz) developed at the boundary with melanosome (bottom right corner). Leucosome is finer-grained closer to the melanosome, which is composed predominantly of biotite. Width of photo is 1mm; photo taken under crossed polars with 10x objective.

myrmekite have been interpreted as the product of retrograde dewatering in response to cooling, when slightly cool melt reacts with crystals (Ashworth and McLellan, 1985). Symplectic textures are indicative of closed systems, because it is in closed systems that melts store fluids which are later released upon cooling (Ashworth and McLellan, 1985). Timmermann (1998) cites evidence from rocks in the Kawagama Zone, the highly tectonized boundary between the Muskoka and structurally underlying Algonquin domain, that contain myrmekitic intergrowths in the absence of leucosome, thus strongly suggesting that these structures may form in response to strain. At Bark Lake, myrmekite was observed in only one sample of a non-migmatic rock, the megacrystic granodiorite orthogneiss. Otherwise, myrmekitic textures only occur in leucosomes and in mesosome material in close spatial association with leucosome; as such, they are interpreted here as primary replacement textures and evidence for *in situ* partial melting. Quartz films, though rare, have been noted; these have been interpreted by previous workers to be analogous to melt pockets, and thus evidence of crystallization from a melt. (iii) Similar bulk compositions between meso- and leucosomes indicate closed system partial melting. This evidence is provided by geochemical analyses, which show similar chemical signatures for mesosomes and their coexisting leucosomes (see 2.4 *Geochemistry*, this thesis). (iv) Because similar bulk compositions are expected for meso- and leucosomes formed in a closed chemical system, this implies that the composition of the leucosomes should approximate that of the granite minimum for the system Qtz-Ab-Or (Ashworth, 1985; Johannes, 1984). Compositions of the leucosomes in the study area range from granitic to granodioritic, owing to the formation of new hornblende and biotite porphyroblasts. In conjunction with similar bulk chemistries, the approximation of

leucosome compositions to that of the granite minimum is further support for *in situ* partial melting. The minor deviation from ideal granitic composition in the leucosomes is attributed to advanced partial melting (Brown and Fyfe, 1972; Ashworth, 1985) which resulted in the evolution of the melt composition away from the granite minimum, becoming progressively granodioritic with increased partial melting. This hypothesis is consistent with the significant volume of melt that was produced in this area.

The amount of water present in any given system is the most important factor controlling both the amount of melt produced and the temperature at which melting begins; i.e., the more water present either locked in hydrous minerals or as a free phase, the more melt can be produced, and melting will take place at lower temperatures. From previous work (see Table 1-2) it is known that near-peak temperatures in the Bark Lake area only reached $800 \pm 40^\circ\text{C}$ (Carr and Berman, 1997), and that pore fluid pressure for water would have been limited by the ambient pressures experienced while these rocks were at mid-crustal depths (8.9 ± 0.9 kbar; Carr and Berman, 1997). Current research indicates that partial melting at mid- to deep-crustal levels occurs at low- $a_{\text{H}_2\text{O}}$, or water-undersaturated, conditions. The mechanism by which melting occurs is called *dehydration melting*, during which the water that is required for melting reactions to proceed is released from hydrous minerals; in this case, biotite and hornblende. By analogy to the migmatites of the Opatika Belt in the Archean Superior Province (Sawyer, 1996, 1998) the presence of a water-rich volatile phase during the initial phases of melting is suggested by the presence of biotite and extensive melting at relatively low temperatures ($\sim 800^\circ\text{C}$). Subsequent melting would have occurred by dehydration melting at progressively lower values of $a_{\text{H}_2\text{O}}$.

Thus, the progression is envisioned as follows: The initial presence of a water-rich volatile phase allowed the onset of melting to begin (by lowering the melting point) at P-T conditions characteristic of lower granulite-facies, comparatively lower than required to melt "dry" bt- and hbl-rich quartzofeldspathic (870-880°C) to intermediate rocks (900-920°C; England and Thompson, 1984). Further water was released during crystallization from melt (to form leucosome). As melting proceeded, any available water was consumed, creating an increasingly low- $a_{\text{H}_2\text{O}}$ environment; melting continued at lower temperatures, until the solidus was reached, by the dehydration of hornblende and biotite. The release of water during crystallization of the extensive leucosome also provides an explanation for the widespread amphibolite-facies overprint that characterizes the rocks of the transect.

Chapter 4: Structural geology of the Bark Lake area

4.1 Introduction

In light of the lack of detailed mapping and structural analysis in the Bark Lake area, mapping at 1:25 000 scale was completed in order to characterize the geological history of the area. The transect lies mainly within the uppermost structural level of the eastern Central Gneiss Belt and crosses into what is ostensibly the tectonized northwestern boundary of the Central Metasedimentary Belt (Figure 4.1; Map 3, back pocket). The north-south transect runs across strike of the proposed trace of the CMBbtz and as such, provided a good location for the investigation into the structural context of this boundary and the legitimacy of the location of its surficial trace in this area (e.g., Burr and Carr, 1994; Easton and Davidson, 1994).

Structures are described chronologically, in the order in which they are interpreted to have developed. Further, those structures having a northeast-orientation are described as being *orogen-parallel*.

4.2 Transposition and Nomenclature

Transposition is a feature of progressive ductile deformation involving the continual partitioning of strain and the successive development of cognate parallel structures (Hobbs et al., 1967; Williams, 1983). In areas that have undergone intense large-scale, homogeneous deformation where transposition has been pervasive, overprinting relationships are ambiguous, and outcrop is discontinuous, it is difficult to accurately discern the number of discrete pre- to syn-transposition folding events

Figure 4.1. Map illustrating the location of the CMBbtz in the Bark Lake area, as determined in this study.

(Scammell, 1993; Williams, 1983; Hobbs et al., 1967). The Bark Lake transect comprises such an area. In areas of lower strain (c.f. Myers, 1978), for instance the necks between amphibolite boudins, early folds are best preserved. Preservation in low strain areas makes it possible to discern that more than one phase of folding constitutes S_T . Due to the nature of transposition, however, it is unclear whether these represent discrete pre- and syn-transposition events or a progressive syn-transposition sequence. For these reasons, the pre- to syn-transposition folds are collectively termed F_{ST} , while younger ones are denoted F_{ST+1} (notation from Scammell, 1993). The main foliation in the map area is the transposition foliation, S_T , which corresponds to the regional foliation.

4.3 Tectonites and related structures

4.3.1 Foliations

The rocks in the northern two-thirds of the Bark Lake area (Map 3, back pocket) are primarily S>L tectonites and rarely S-L tectonites. Throughout most of the transect, the foliation dips moderately (20°-50°), predominantly to the southeast (Figure 4.2a-g). In the southern part of the transect, the main foliation shallows and dips consistently to the south, with an average orientation of 092°/20°S (Figure 4.2h). Deformation of the plutonic and rare supracrustal rocks has resulted in the development of moderately- to highly-strained (c.f. Myers, 1978) banded gneiss with a penetrative transposition foliation, S_T (Figure 4.3a-b). S_T may include regular, porphyroclastic and straight gneissic layering or local mylonites depending on the amount of strain and annealing. In areas of relatively low strain, the tectonites exhibit regular and irregular gneissic layering

Figure 4.2 Stereonet plots for poles to the transposition foliation, S_T .

(a)

Central Bark Lake
★ Mean = 094/19 S
Poles to S,
N=46

(b)

North Central Bark Lake
★ Mean = 107/23 SW
Poles to S,
N = 46

(c)

Poles to S,
Norlock Lake
and Littlebark Bay
★ Mean = 352/15 NE
N= 17

(d)

Poles to S., Depot Point
★ Mean = 152/08 SW
N= 42

(e)

HWY 60, NW of Barry's Bay
★ Mean = 088/21 S
Poles to S,
N= 33

(f)

Poles to S.: Town of Barry's Bay
★ Mean = 327/74 NE
N= 54

(g)

Poles to S.; Hams Bay
★ Mean = 141/13 SW
N= 46

(h)

Poles to S.: Papineau Lake
★ Mean = 092/20 S
N= 25

(*sensu* Hanmer, 1988), with low-angle discordant features such as apophyses from dykes, patchy leucosome and folds with tight to close interlimb angles (Figures 4.3c-d). The gneissic layering is defined by alternating bands of fine- to medium-grained felsic and mafic minerals layered on the scale mm's to dm's. Felsic layers are granitic with variable amounts of plagioclase, k-feldspar and quartz, as well as minor biotite and hornblende. Mafic layers comprise variable hornblende, biotite and plagioclase, with minor clinopyroxene and quartz.

Straight and porphyroclastic gneisses become locally dominant due to increased strain, which may occur over distances as small as a few mm (see 4.2.2 *Strain Gradients*, below), and are transitional into one another both across and along strike. Straight and porphyroclastic gneisses are abundant in the southern part of the of the transect. Straight gneiss is a pink, white and black striped rock; the layering is composed of unbroken, laterally extensive, thin (1-5 cm) bands of alternating granite and amphibolite with sharply defined boundaries (Figure 4.4a). Porphyroclastic gneiss predominantly exhibits a streaky appearance of pink porphyroclastic lenses and individual porphyroclasts within foliated grey host, which are transitional into continuous layers resembling those of the banded straight gneiss but in which small (<2 - 0.5 cm), pink and white porphyroclasts are visible (Figure 4.4b). Porphyroclasts of K-feldspar and plagioclase occur both as large (up to 10 cm), relict single crystals and recrystallized aggregates of smaller crystals. The crystals represent remnants of pre-existing coarse-grained igneous and metamorphic rocks such as granitic leucosome and syenitic pegmatites that, together with their host, were sheared out and mechanically disaggregated (and recrystallized) under high strain.

Figure 4.3 Migmatitic granodioritic orthogneiss with penetrative transposition foliation, S_T . (a) regular banded gneiss, moderate strain. Note lozenge of low-strain within the higher strain body. Pen magnet is ~10 cm long; (b) moderate strain irregular banded gneiss with voluminous leucosome. Spine of notebook is 16 cm long.

Figure 4.3. Migmatitic granodioritic orthogneiss with penetrative transposition foliation; continued. (c) regular banded gneiss illustrating low-angle discordant apophysis from dyke (immediately below notebook). Spine of notebook is 16 cm long; (d) low strain blebby leucosome; pen is 15 cm long.

(c)

(d)

S_T is defined by concordant leucosome lenses, intrafolial rootless fold closures, and attenuated isoclinal fold limbs coplanar with the gneissic layering (Figure 4.4c-d). There is evidence of progressive stretching and rotation of pre-existing structures, such as coarse-grained felsic and amphibolitic vein-networks that intruded into the orthogneiss protolith, and early folds within the gneiss. Rare sheath folds have also been observed.

Microstructures associated with S_T include mm- to cm-wide isoclinally folded quartz veins, relic S/C fabrics defined by biotite laths, and abundant subgrains. Quartz veins show evidence of recovery; they are no longer coherent single crystals but consist of multiple subgrains. Deformation twins, undulose extinction, bent crystals and decussate biotite textures are common.

4.3.2 Structures associated with S_T

Folds. Important in the evolution of S_T are the pre- to syn-transposition folds that formed and were progressively deformed during its development, as well as those that were potentially formed prior to transposition. Folds within F_{ST} are tight to isoclinal, and intrafolial with respect to S_T (Figure 4.5a-b). F_{ST} folds are of “similar” style (Class 2 ; Ramsay, 1967), with thick hinges and thin limbs. These are the oldest structures in the map area, and they fold a generation of anatectic melt, AL_1 , as well as the mafic layers. Fold axes within F_{ST} plunge shallowly northwest or southeast, and are generally collinear with the mineral and stretching lineations throughout the area (Figure 4.5c). The average plunge and trend for F_{ST} fold hinges is $16^\circ/146^\circ$.

Mineral and Extension Lineations. Collinear mineral and extension lineations are common throughout the area. The mineral lineations are made up of elongate crystals or crystal-aggregates of biotite and/or hornblende, and are best observed in layers

Figure 4.4. High strain orthogneiss of the Bark Lake area. (a) Straight gneiss illustrating alternating bands of pink and white granite and black amphibolite, with sharply defined boundaries. Spine of notebook is 16 cm long; (b) Porphyroclastic gneiss illustrating the common streaky appearance defined by pink felsic porphyroclasts in a grey foliated host. Pencil is 15 cm long.

Figure 4.4. continued. (c) and (d) Migmatitic granodioritic orthogneiss illustrating ST defined by concordant leucosome lenses, tight-isoclinal folds and attenuated mafic layering. Spine of notebook is 16 cm long.

(c)

(d)

containing significant proportions of these minerals. Mineral lineations plunge shallowly to the northwest or southeast, with an average orientation of $09^{\circ}/132^{\circ}$ (Figure 4.6a). Mineral lineations exhibit a variable degree of development from weak to strong, and are homogeneously distributed throughout the transect.

Extension lineations consist of elongate recrystallized aggregates of quartz, plagioclase and k-feldspar, and less commonly hornblende, and plunge shallowly to moderately to the southeast or northwest with an average orientation of $14^{\circ}/134^{\circ}$ (Figure 4.6b-c). They show a heterogeneous distribution across the transect: in the northern two-thirds, they are preferentially developed in felsic-rich layers and local mylonite zones. Toward the south, extension lineations become penetratively developed throughout the rocks irrespective of layer composition, and plunge more shallowly to the southeast.

Extension lineations represent the X-axis of the finite strain ellipsoid (where $X>Y>Z$; Flynn, 1962); as such, their orientation is parallel to the direction of maximum horizontal shortening. Extension lineations in the Bark Lake area are interpreted to have formed during NW-directed thrusting and reactivated during later extension.

4.4 Post- S_T Structures

4.4.1 F_{ST+1} folds

F_{ST+1} folds are interpreted not as representing a discrete folding event, but the latter stages of compression in a continuum of deformation. F_{ST+1} post-dates S_T (and thus F_{ST}) and AL_1 , but predates the development of cross-cutting shear bands and the formation of AL_2 (AL_1 , AL_2 ; see Chapter 3, this thesis). F_{ST+1} comprises asymmetric, upright to inclined folds with open to tight interlimb angles. They are generally Class 2

Figure 4.5 Folds within F_{ST} . (a) Tight, intrafolial Z-fold plunging southeast (into the page); (b) Attenuated isoclinal fold defined by thin amphibolite layer (below notebook); (c) Stereonet plot of fold axes within F_{ST} . Mean orientation is $16^{\circ}/146^{\circ}$.

(a)

(b)

(c)

F_{ST} Fold Axes, Bark Lake
☆ = Mean = 16/146
N = 49

folds (“similar folds”; Ramsey, 1967) with thick hinges and thin limbs, but hinge thickness appears to decrease when interlimb angles are larger (Figure 4.7a-b). Where overprinting relationships with F_{ST} folds are discernable, they are predominantly Type II interference patterns (“arrowhead” or “mushroom”; Twiss and Moores, 1992), and less commonly Type I patterns (“dome-and-basin”; Ramsey, 1967). Open, symmetric folds are most obvious in the straight gneiss in the northernmost part of the map-area, adjacent to Highway 60, where they gently warp the rocks. Tighter, asymmetric folds with dominantly Z-vergence are best preserved along the shores of Bark Lake in the central part of the transect (Map 3, back pocket), where they occur in well-exposed, weathered-out fold trains (Figure 4.7c-d). Fold axes of F_{ST+1} are orogen-parallel, and plunge shallowly to moderately (2° - 32°) to the southwest or northeast with an average orientation of $08^{\circ}/240^{\circ}$ (Figure 4.7e).

There is limited preservation of folding in the southernmost part of the transect. Rare isoclinal folds within S_T (F_{ST}) and tight F_{ST+1} folds have been observed, but these are the exception rather than the rule. This is interpreted to be a result of very high strain which obliterated evidence of folding.

4.4.2 *Shear bands*

Shear bands, both extensional and compressional, occur at the cm- to m-scale throughout the map area, where they cut across S_T and truncate F_{ST+1} folds. Extensional shear bands outnumber compressional ones, and shear bands dip predominantly shallowly to moderately SE, less commonly to the SW. Extensional shear bands acted as sites of localization of pegmatite and/or late leucosome (AL_2) (Figure 4.8a-b). Kinematics most often indicate top-side-down to the SE movement, parallel to the finite extension

lineation. The paucity of thrust-sense kinematic indicators suggests that many of the extensional shear bands represent reactivated thrusts, which resulted in the obliteration of the evidence for early thrust movement. This does not imply, however, that *all* extensional structures are reactivated earlier structures; many extensional shear bands are probably late, "extension only" structures.

4.4.3 Ductile Annealing

The rocks in the CGB show variable degrees of dynamic recrystallization. For the most part, there is a bimodal grain-size distribution of coarse relict deformed grains in a matrix of finer, strain-free recrystallized grains. Large crystals preserve abundant evidence for internal deformation, such as undulose extinction, deformation lamellae and sharp-edged deformation twins (Figure 4.9a). Subgrains and deformation bands, evidence of recovery processes, are ubiquitous in minerals that have not yet been fully recrystallized (Figure 4.9b). In these samples grains are generally anhedral to subhedral, display a weak shape preferred orientation that defines the foliation, and contain relic S/C fabrics.

Local very high strain zones (e.g., at Carson Lake, SM98-129; Trout Lake, SM98-159) that in the field displayed the most intense deformation have experienced the most complete recrystallization (Figure 4.9c-d). Grain boundaries are straight and polygonal, crystals are strain-free and there is no evidence of strain at the crystal scale despite the mm- to cm-scale straight gneissic layering, penetrative stretching lineation and shear-sense indicators observed in the field. Although completely recrystallized, there are relic S/C fabrics that indicate both thrust-sense (SM98-129) and normal (tops to the southwest, in this case; SM98-159) movement.

Figure 4.6. Mineral and extension lineations at Bark Lake. (a) Stereoplot illustrating orientation of mineral lineations. Average orientation is $09^{\circ}/132^{\circ}$; (b) Photograph illustrating a well-developed extension lineation in a felsic layer within banded orthogneiss. Pen is 15 cm long; (c) Stereoplot illustrating orientation of extension lineations. Average orientation is $14^{\circ}/134^{\circ}$.

(a)
Mineral Lineations
Bark Lake
N=46

(b)

(c)

Extension Lineations
Bark Lake
N= 85

Figure 4.7 F_{S_T+I} folds in the Bark Lake area. (a) Tight S-fold with thick hinges and thin limbs ("similar" fold). Fold folds S_T . Brunton compass for scale. (b) Open fold plunges shallowly southwest (toward the reader). Layer thickness maintained around throughout the fold ("parallel" fold with rounded hinge). This fold deforms straight gneiss adjacent to Highway 60. Person for scale.

(a)

(b)

Figure 4.7. F_{ST+1} folds in the Bark Lake area, continued. (c) and (d) are photos taken along the shoreline of Bark Lake, and illustrate the well-exposed weathered-out fold trains common along the shoreline in the central part of the transect. Width of view is ~2 m. (e) Stereoplot illustrating orientation of F_{ST+1} fold axes. Average orientation is $08^{\circ}/240^{\circ}$.

(c)

(d)

(e)

F_{ST-1} Fold Axes
☆ Mean = 08/240
N = 50

4.4.4 Brittle Faults

Late, steeply dipping, conjugate sets of brittle faults dissect the rocks throughout the area, but appear more numerous in the southern part of the transect along Papineau Lake. Brittle faulting resulted the generation of sparse cataclasite, which ranges from a biotite-rich orange-brown material to mottled mesocratic material; in both cases it lacks coherency and is easily broken up into finer pieces. Faults of this type generally occur on the m- to dm-scale and displacement of pre-existing structures has been less than a few meters.

4.5 Mapped Deformation Intensity

The rocks throughout the map-area have experienced variable amounts of strain, from moderate to very high. In the field, variation in strain was mapped by using criteria that included the concordance or discordance of syntectonic structures with respect to S_T , such as leucosome (AL_1) and cross-cutting structures like dykes, apophyses and veins; tightness of folds; and, type of gneissic layering, from regular, banded gneiss through to straight and porphyroclastic gneisses. Due to the heterogeneity of structural preservation across the transect, the most useful tools for mapping deformation intensity proved to be syntectonic pegmatite dykes, which are ubiquitous across the transect.

4.5.1 Syntectonic Pegmatites

Grenvillian pegmatites, generally of (quartz)syenitic composition, intruded the orthogneiss during the progressive development of S_T and as such, imply reactivation of this foliation. In areas of relatively low strain, coherent pegmatite-host cross-cutting relationships are observed. Pegmatites cross-cut S_T at moderate to high angles, have

sharp and regular contacts with their host, and are massive to weakly foliated, mainly at dyke-margins. With increased strain, pegmatites were progressively rotated toward parallelism with the flow plane, thus reducing the angle between foliation and pegmatite contact. Pegmatites that cross-cut the foliation at low angles, have sharp but irregular contacts and an internal foliation (sub)concordant with that of the host are interpreted as representing higher-strain equivalents of coherent, high-angle pegmatites. Very high strain obliterated coherent cross-cutting relationships and resulted in mechanical disaggregation and shearing-out of pegmatites which yielded lenses, stringers and single porphyroclasts that collectively constitute porphyroclastic gneiss; all intermediate stages from coherent, foliated pegmatite to tectonoclastic gneiss have been observed in the field.

4.4.2 Strain Gradients

Variations in strain, or strain gradients, are manifested at all scales; the amount of strain changes over distances as small as mm's and cm's and as large as km's. This is a function of the small-scale compositional heterogeneities that characterize the transect, as well as the local rate of strain and orientation of pre-existing structures, all of which affected strain partitioning. Despite local small-scale variations in strain, the general trend is one of increasing strain toward the south relative to the north; a prominent north-south regional strain gradient has been documented across the transect (Figure 4.10). For simplicity, the northern two-thirds of the transect, from Highway 60 to the southern end of Harris Bay, is referred to as "the northern domain", and the remaining third which encompasses Papineau Lake is referred to as "the southern domain". Strain in the northern domain was moderate to high, with local zones of both low and very high strain

Figure 4.8. Examples of extensional shear bands commonly found in the Bark Lake area. (a) Orthopyroxene-bearing syenitic pegmatite localized into dilatant zone; (b) Hornblende porphyroblast-bearing AL2 localized in shear band. Spine of notebook is 16 cm long.

(a)

(b)

Figure 4.9 Microstructures. (a) Photomicrograph illustrating evidence for internal deformation: sharp-edged deformation twins in plagioclase. Width of view is 1 mm; photo taken under crossed polars with 10x objective. (b) Evidence of recovery: note undulose extinction and subgrain boundary development in quartz (top right corner); note absence of polygonal grain boundaries. Deformation bands visible in plagioclase. Width of view is 1 mm; photo taken under crossed polars with 10x objective.

Figure 4.9. Microstructure, continued. (c) near complete recrystallization of quartz and plagioclase in straight gneiss. Note the polygonal grain boundaries and strain-free crystals. Width of view is 1 mm; photo taken under crossed polars with 10x objective. (d) Polygonal grain boundaries and nearly strain-free crystals, also from straight gneiss. There is minor undulose extinction in some of the quartz (bottom left); note recrystallized quartz vein that traverses the photo. Width of view is 1 mm; photo taken under crossed polars with 10x objective.

(Figure 4.10). In the northern domain, the contacts between pegmatites and their hosts are sharp and straight, and the pegmatitic bodies are coherent. Pegmatite-host contacts become increasingly irregular and distorted toward the southern end near Harris Bay and Papineau Lake.

There is a discrete high strain zone in the northernmost part of the map area adjacent to Highway 60, where it occurs as an elongate tract of straight and porphyroclastic gneiss with a penetrative extension lineation (Map 3, back pocket). The transition from moderate to very high strain is abrupt and it is likely that this zone represents a lithotectonic domain-bounding ductile shear zone. While the regional significance of this zone is poorly constrained, it is possible that it is contiguous with the Kawagama Zone, a highly tectonized zone recognized further to the west, that separates the Muskoka domain from the underlying Algonquin domain (Timmermann, 1998).

4.6 Discussion

The Bark Lake area (Maps 1, 2 and 3) is characterized by a penetrative transposition foliation, S_T , defined by attenuated layering and rootless isoclinal folds. S_T was intruded by syenitic pegmatites and overprinted by folding, shear bands and brittle faulting. The state of deformation of the rocks in the map area ranges from moderate to very high strain (c.f. Myers, 1978).

F_{ST} and F_{ST+1} are interpreted to represent progressive ductile deformation in response to burial by overthrusting of the Central Metasedimentary Belt onto the pre-Grenvillian Laurentian margin. Early folds within F_{ST} comprise those that were formed synchronous with, and/or those that predated, the transposition foliation, S_T ; their fold

Figure 4.10. Stylized form-surface map illustrating transect-wide strain gradient in the Bark Lake area. Note the progression from coherent pegmatites with straight, sharp boundaries in the north, to the disaggregated, sheared-out pegmatites in the south (now porphyroclasts within foliated host).

axes plunge southeast, transverse to the orientation of the orogen and parallel to the extension lineation. Fold axes within F_{ST} are interpreted to have formed in different orientations and were progressively reoriented toward parallelism with the extension lineation as a result of differential flow due to heterogeneous deformation. This geometry, consisting of fold axes and extension lineations that are oriented at a high angle (transverse) to the orogenic belt, is consistent with crustal shortening perpendicular to the strike of the orogen (Escher and Watterson, 1974), in this case to the northwest. That F_{ST+1} fold axes plunge southwest and northeast, parallel to the orogenic front, is attributed to their formation late in the progressive deformation. Reorientation of structures is a function of both initial orientation and the time of their formation, thus later-formed structures are more apt to remain oriented at high angles to the extension lineation (Escher and Watterson, 1974).

The distribution of structures suggests that the northwestern limit of deformation attributed to movement on the CMBbtz does in fact pass through the southern part of the map area, close to where early workers have placed it. This interpretation is based on the distribution of extension lineations which, although of restricted occurrence in the northern map-area, are penetratively developed in the southern part. This observation, in conjunction with both the abundance of straight and porphyroclastic gneiss, and the culmination of the transect-wide strain gradient in the south, provides further support for the presence of the shear zone. It should be noted, however, that the edge of the shear zone in the Bark Lake area does not correspond to a lithologic boundary, rather it is a subtle boundary defined on the basis of structures as the edge of a km-scale strain gradient, which overprints earlier structures, including the lithologic contact between the

CGB and the CMB. This lithologic contact between the predominantly quartzofeldspathic orthogneiss of the CGB and the mixed rocks of the NW boundary of the CMB, was obscured and reworked by multiple thrusting events within the shear zone. This has been found to be the case along much of the basal zone of the CMBbtz in Ontario, where previous workers report an interdigitation of carbonate and silicate gneiss, in contrast to the Quebec extension of the boundary thrust zone which is lithologically abrupt (Hanmer and Ciesielski, 1984). In addition, in contrast to mapped parts of this boundary in the Killaloe, Centreview and Haliburton areas, no meta-pelites or marble tectonic melange are associated with it in the Bark Lake area (e.g., Davidson et al., 1982; Culshaw et al., 1983; Hanmer and Ciesielski, 1984; Hanmer et al., 1985; Hanmer, 1988; Hanmer and McEachern, 1992; McEachern and van Breeman, 1993); these authors proposed that the rheologically weak marble and paragneiss played an important role in thrust localization at the base of the boundary. The Papineau Lake area may simply represent a siliceous part of the large-scale “interdigitation” (Hanmer and Ciesielski, 1984), as just 4.5 km east of Papineau Lake in Centreview, garnet-sillimanite gneiss (meta-pelite) has been documented.

Geophysical data from the northern end of Lithoprobe seismic reflection profile Line 32 (see Figure 1.3) identifies a panel of strong, discontinuous seismic reflectors with gentle, southeasterly apparent dips interpreted to define ductile thrust ramp-and-cutoff geometries (Burr and Carr, 1994). When projected to the surface, this panel closely coincides with the mapped location of the northwestern limit of shear zone (CMBbtz) in this area (Maps 2, 3, back pocket).

Chapter 5: U-Pb geochronology of the Bark Lake area

5.1 Introduction and sample selection

Samples were collected throughout the Bark Lake area for U-Pb geochronological analysis (Figure 5.1; Map 1, back pocket). The objectives were to determine the protolith ages of prominent lithologies as well as the timing of regional high-grade metamorphism and deformation, in order to place constraints on the evolution of CGB rocks and the timing of movement on the Central Metasedimentary Belt boundary thrust zone (CMBbtz). These data are critical in terms of: (1) characterization of the tectonic evolution of the Bark Lake area; (2) making comparisons with other domains within the CGB; (3) understanding this transect in the context of the regional tectonic frameworks, for example, constraining the timing of initial thrusting on the CMBbtz.

5.2 Analytical procedures

Results and analytical data are presented in Table 5.1; age uncertainties are reported at the 2σ (95% confidence) level. Six fresh samples were collected, each weighing ~10 kg. Heavy minerals were liberated from the samples by standard crushing, grinding, Rogers Gold™ table, and heavy liquid techniques. Magnetic separation of the heavy mineral concentrate was accomplished using the Frantz™ isodynamic separator. The clearest, most inclusion- and crack-free zircons were selected from the 1.8 A non-magnetic (10° side-slope, 15° front-slope) fraction and air abraded using the methods of Krogh (1982). Air abrasion time for the samples ranged from 4 to 23 hours, in order to ensure removal of potential metamorphic overgrowths. U- and Pb-extraction from zircon

Figure 5.1. U-Pb geochronology sample location map.

Table 5.1. U-Pb geochronology data, eastern Muskoka domain, Barry's Bay - Barf Lake, ON

Fraction ^a	Wt. (µg) ^b	U (ppm)	Pb ^c (ppm) ^d	206Pb/238U ^e	Pb (ppm)	Th (ppm)	206Pb/232Th ^f	206Pb/238U ^g	Age (Ma) ^h	206Pb/238U ⁱ	Age (Ma) ^j	Corr. Coeff. ^k	206Pb/238U ^l	% Disc. ^m
SM98-221														
A	8	60	19	2766	3	118	0.40	0.2243 ± .47	2.806 ± .38	0.9070 ± 0.3	1304.8 ± 11	0.77	1440.4 ± 11.3	10.4
B	17	164	36	5999	3	10.1	0.41	0.1657 ± .16	2.240 ± .15	0.00751 ± 0.4	1097.8 ± 31.3	0.96	1371.7 ± 16.1	21.7
C	10	50	19	4073	3	6.3612	0.24	0.2056 ± .17	2.3510 ± .17	0.00277 ± 0.1	1205.2 ± 3.7	0.82	1263.8 ± 4.1	5.1
D	8	142	33	2355	7	0.6	0.50	0.2241 ± .19	2.689 ± 0.20	0.00702 ± 0.7	1303.4 ± 4.5	0.50	1361.0 ± 2.8	4.7
SM98-232														
A	43	136	27	20090	3	20.9	0.93	0.1711 ± .24	1.7535 ± .22	0.0743 ± .11	1018.2 ± 4.6	0.89	1050.4 ± 4.6	3.3
C	38	19	4	1034	5	19.9	0.88	0.1746 ± .73	1.766 ± .69	0.0742 ± .57	1037.3 ± 14.1	0.68	1040.4 ± 8.9	1.0
D	26	80	16	3535	5	21.0	0.94	0.1731 ± .14	1.766 ± .16	0.0740 ± .10	1029.3 ± 2.6	0.75	1040.7 ± 4.2	1.2
E	22	49	10	2520	5	19.7	0.87	0.1741 ± .25	1.796 ± .24	0.0748 ± .15	1034.6 ± 4.7	0.81	1063.3 ± 6.1	2.9
SM98-238														
A	22	430	53	21450	3	3.0	0.11	0.1280 ± .53	1.315 ± .49	0.0741 ± .22	781.1 ± 7.7	.91	1042.9 ± 8.6	26.6
C	25	287	49	5249	14	8.8	0.34	0.1634 ± .65	1.6445 ± .76	0.0730 ± .46	975.5 ± 15.5	.84	1014.2 ± 18.5	4.1
D	26	659	111	13230	13	10.9	0.43	0.1919 ± .13	1.616 ± .84	0.0724 ± .94	967.4 ± 22.5	.66	996.7 ± 37.9	3.2
E	24	471	79	22790	5	8.8	0.34	0.1653 ± .13	1.6450 ± .13	0.0722 ± .06	986.3 ± 2.3	.90	994.2 ± 2.3	0.9
SM98-118														
A	24	114	27	7231	3	10.7	.43	0.2301 ± .21	2.805 ± .21	0.0084 ± .08	1335.2 ± 5.1	.93	1391.4 ± 2.9	4.5
B	14	101	25	5688	4	11.6	.48	0.2378 ± .15	2.8394 ± .15	0.0086 ± .06	1375.3 ± 3.6	.92	1418.0 ± 2.3	3.3
C	14	168	42	9632	4	10.9	.44	0.2375 ± .10	2.8311 ± .10	0.0085 ± .05	1373.7 ± 2.4	.89	1415.0 ± 1.8	3.2
D	14	171	43	4192	8	13.3	.54	0.2333 ± .18	2.8612 ± .18	0.0080 ± .06	1351.7 ± 4.3	.95	1403.1 ± 2.2	4.1
SM98-120														
B	2	447	99	787	14	10.2	.41	0.2131 ± .35	2.5469 ± .46	0.0067 ± .28	1245.6 ± 7.9	.79	1352.9 ± 10.8	8.7
C	4	160	34	521	15	14.2	.46	0.2012 ± .15	2.3040 ± .29	0.0030 ± .23	1182.1 ± 3.2	.61	1269.9 ± 9.1	7.6
D	2	390	87	146	60	11.3	.46	0.2119 ± .66	2.4915 ± .19	0.0052 ± .17	1238.8 ± 14.9	.52	1322.0 ± 64.4	6.9
E	1	361	87	949	7	9.2	.37	0.2041 ± .23	2.878 ± .36	0.0069 ± .29	1358.9 ± 5.7	.59	1402.8 ± 11.1	3.5
G	3	207	46	1024	8	9.7	.38	0.2141 ± .39	2.4879 ± .42	0.0042 ± .23	1250.5 ± 8.9	.75	1299.2 ± 11.1	4.1
H	3	270	67	2854	4	10.2	.41	0.2369 ± .30	2.9132 ± .31	0.0092 ± .11	1370.6 ± 7.4	.91	1408.1 ± 4.0	3.0
SM98-149														
B	6	279	65	9322	3	6.8	.26	0.2311 ± .34	2.8463 ± .31	0.0094 ± .17	1340.2 ± 8.3	.88	1412.6 ± 6.3	5.7
C	8	195	46	1453	15	5.5	.21	0.2369 ± .22	2.9193 ± .28	0.0084 ± .22	1370.5 ± 5.4	.84	1412.2 ± 8.4	3.3
D	7	201	44	2578	10	5.9	.23	0.2222 ± .41	2.6393 ± .36	0.0061 ± .32	1293.7 ± 9.7	.86	1341.1 ± 12.4	3.9

^aUnless otherwise specified, all zircon are clear, transparent, crack- and inclusion-free grains, abraded (Krogh, 1982) and obtained from the 0.8µm non-magnetic fractions of Frantz[®] separates. Weighing error is 1 microgram. Radiogenic Pb, ²⁰⁶Pb, corrected for spike and Pb fractionation of 0.09%, 10.03%/amu; total common Pb in analysis corrected for spike and fractionation; corrected for Pb and U laboratory blank where 2004/201/204/206/204 are 19.02/15.65/38.23, errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma; ^c correlated for common Pb and laboratory blank; errors are 2 standard errors of the mean in Ma.

was completed in the Carleton University U-Pb geochronology laboratory following the isotope dilution procedures outlined in Parrish et al. (1987). Isotope compositions of U and Pb were measured using the Finnigan-MAT 261 variable multicollector mass spectrometer at Carleton University, following procedures outlined by Roddick et al. (1987). Pb blanks were all 3-5 pg. Regression and error propagation algorithms were those of Roddick et al. (1987) and York (1969). Most fractions analyzed were single crystals; multiple-crystal fractions consisted of a single morphology and were restricted in number ($N < 6$).

5.3 Sample Descriptions

5.3.1 Sample 1: Migmatitic monzogranitic orthogneiss (SM98-23T)

Sample 1 is representative of the ubiquitous pink-gray, fine- to medium-grained, dioritic migmatite that is the main map unit in the Bark Lake area. This amphibolite-facies quartz monzodiorite orthogneiss has experienced moderate strain (c.f., Myers, 1978). It is foliated, and contains mafic enclaves and cross-cutting syn-tectonic pegmatites, both of which were avoided during sampling. The sample was collected from the eastern side of the island in Littlebark Bay, in the northern part of the map-area.

Four zircon fractions were analyzed: A and B were single crystals, while fractions C and D each consisted of two zircons. Crystals were of two different morphologies (Figure 5.2a): sub-rounded, ovoid crystals (fractions A and D) and euhedral, elongate prisms (fractions B and C). All selected crystals were large (105-149 μm), colourless and inclusion-free. Faint oscillatory zoning and overgrowths were common

Figure 5.2. Photomicrographs illustrating zircon morphology for individual samples.
(a) SM98-23T; (B) SM98-23P; (c) SM98-28; (d) SM98-118; (d) SM98-118; (e) SM98-129; (f) SM98-149. See text for discussion.

features in fractions A and D, however, crystals were abraded for 3-4 hours to eliminate overgrowths.

Despite the colourless nature and excellent clarity of selected crystals, all fractions were 5-22% discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages that range from 1263-1440 Ma. 1440 ± 11 Ma is the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ age, and interpreted as the minimum age of protolith crystallization. Given that oscillatory zoning is a common feature of zircons interpreted as being igneous in origin (i.e., that have crystallized from a magma; Pidgeon, 1992), in conjunction with their prismatic morphology, these fractions are interpreted to constitute a single zircon population. Data do not fall on a line (Figure 5.3a); they are distributed about a discordia chord defined by the least discordant fractions (C and D), with an upper intercept of $1514 +40/-34$ Ma and a lower intercept of $1046 +34/-39$ Ma. Fraction A was excluded from the regression because its distribution relative to the other fractions implies a component of inheritance. The upper and lower intercepts are interpreted as ages of igneous crystallization and high-grade metamorphism with accompanying anatexis, respectively.

5.3.2 *Sample 2: Foliated quartz syenogranite pegmatite (SM98-23P)*

This sample is a moderately foliated, ~30 cm wide biotite-magnetite quartz syenite pegmatite that intruded host migmatitic orthogneiss, and was collected from the same location as sample SM98-23T (above). The foliation in the pegmatite is sub-parallel to that in the host, and strongest at the margins. The pegmatite has sharp, irregular contacts and cross-cuts its host at a moderate angle, and is thus interpreted as late-synectonic. Apophyses are parallel to the foliation in the orthogneiss.

Figure 5.3. Concordia diagrams for Samples 1 and 2. See text for discussion.

Four fractions were analyzed, consisting of both multiple and single-crystal fractions so that each fraction was approximately the same weight. Zircons were large ($>149\ \mu\text{m}$), colourless, and euhedral faceted prisms with prominent igneous zoning (Figure 5.2b). No cores or overgrowths were optically detectable. Based on the crystal morphology and prominent zoning, the zircons are interpreted as igneous.

Fractions are 1-3% discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ ages falling in the range of 1040-1050 Ma. Fraction E is an outlier with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1063 Ma. When plotted on a concordia diagram, the distribution of data suggests a component of inheritance in fraction E, despite optically undetectable cores. For these reasons, E was eliminated from the regression (York, 1969). Fractions A, C and D define a chord (Figure 5.3b; MSWD = 1.38) with an upper intercept of $1036 \pm 10/3$ Ma and a lower intercept that passes through the origin, interpreted as igneous crystallization and recent Pb-loss, respectively.

5.3.3 *Sample 3: Undeformed, cross-cutting syenogranite pegmatite (SM98-28)*

Sample 3 was collected from a small (20m x 30m) island ~50 m east of Beavertail Island, in the north of Bark Lake. It is a striking outcrop: the biotite-magnetite quartz syenite dyke is 8-10 m wide and 25-30 m long, undeformed, and exhibits pristine igneous graphic texture (Figures 5.4a-b). This sample was collected in order to place an upper limit on the age of deformation in the area.

Four fractions were analyzed. Crystals were large ($>149\ \mu\text{m}$), colourless and transparent, euhedral, faceted elongate prisms with strong oscillatory zoning, and a distinct lack of both cores and overgrowths (Figure 5.2c). Despite apparently pristine crystals, all fractions were discordant (1-27%) with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging from 994-1042 Ma. A York (1969) best-fit regression through all points yielded a discordia chord

Figure 5.4. Sample 3 (SM98-28) is an undeformed and unmetamorphosed quartz syenite pegmatite. (a) Note coarseness of grain size; (b) Photomicrograph illustrating striking graphic texture. Photo taken with a x10 objective lens under crossed polars; (c) concordia diagram for Sample 3. See text for details.

(a)

(b)

(c)

with an upper intercept of 993 ± 2 Ma and a lower intercept that passes through the origin (Figure 5.4c). The upper intercept is interpreted to reflect the time of igneous crystallization, and the lower intercept indicates recent Pb-loss. Chord rotation due to lead loss is interpreted to have had a negligible effect on the calculated upper intercept value due to the fact that one fraction is near concordant and the other fractions plot close to the upper intercept (e.g., Nadeau and van Breeman, 1998). Based on its age and field relationships, it is probable that this pegmatite belongs to a suite of ~990 Ma pegmatites, the occurrence of which is widely documented throughout the Grenville Province (van Breeman, 1986; Lumbers et al., 1990; Corfu and Easton, 1995; Ketchum and Davidson, 1999). This interpretation is consistent with the field observations that this pegmatite is both undeformed and unmetamorphosed, and with zircon morphologies typical of growth in igneous environments.

5.3.4 Sample 4: *Meta-megacrystic gneiss (SM98-118)*

Sample 4 is a pink and gray, metamorphosed megacrystic granitoid gneiss. It is an amphibolite-facies rock with a well-developed foliation ($S > L$). This sample was collected ~1 km southwest of Barry's Bay, and is described in section 2.1.5 (this thesis).

Four single-crystal fractions were analyzed. Zircons were large ($>149 \mu\text{m}$), elongate faceted prisms with overgrowths and faint igneous zoning (Figure 5.2d). They are colourless and transparent with minor inclusions but no visible cores. All fractions are 3-4% discordant, with 1391-1418 Ma $^{207}\text{Pb}/^{206}\text{Pb}$ ages. Regression through all fractions yields a discordia curve with an upper intercept of $1468 +26/-17$ Ma, and a lower intercept of $894 +110/-116$ Ma (Figure 5.5a; MSWD = 0.95). While the upper intercept is interpreted as protolith crystallization, the significance of the lower intercept

Figure 5.5. Concordia diagrams for (a) Sample 4; and (b) Sample 5. See text for discussion.

(a)

(b)

value is speculative at best, and not necessarily relevant. One possibility is that the lower intercept represents an episode of Pb-loss during exhumation.

5.3.5 Sample 5: *Retrogressed granulite-facies orthogneiss (SM98-129)*

Sample 5 was collected from Highway 60 immediately south of Carson Lake (Figure 5.1; see also *Stop 1-1* of Easton and Davidson, 1994; *sample BLB-95-13*, Carr and Berman, 1997). This sample is locally garnet- and orthopyroxene-bearing, strongly foliated gneiss collected from a rare outcrop in which granulite-facies assemblages are preserved. Previous interpretations of this outcrop suggested that the intrusion of syn-kinematic pegmatites (1029 $\pm$ 8/-4 Ma, U-Pb, zircon; Carr and Berman, 1997) was accompanied by retrogression of the granulite-facies assemblages. Sample 5 was collected in an effort to date protolith crystallization and the high-grade metamorphic event(s?).

Six fractions were analyzed, consisting of two different morphologies (Figure 5.2e): (1) equant, rounded crystals, which were interpreted to have grown, or been modified in the solid state, during metamorphism; and (2) tips from euhedral, elongate prisms which were interpreted as igneous. All crystals were colourless with excellent clarity, very few inclusions and the elongate crystals were faintly to well-zoned. $^{207}\text{Pb}/^{206}\text{Pb}$ ages range from 1270-1408 Ma, and all fractions are between 2-9% discordant; there is no correlation between age and morphology. Linear regression (York, 1969) through four of these fractions produced a discordia chord with an upper intercept of 1499 $\pm$ 51/-35 Ma and a lower intercept of 1073 $\pm$ 74/-86 Ma (Figure 5.5b; MSWD =

Figure 5.5. (continued) Concordia diagrams for (c) Sample 6; and (d) the granitoid compilation. See text for discussion.

0.39). These ages have been interpreted to reflect protolith crystallization and amphibolite-facies metamorphism, respectively.

5.3.6 Sample 6: *Non-migmatitic orthogneiss (Migmatite precursor; SM98-149)*

Sample 6 is a pink-brown, strongly foliated, granodioritic gneiss which was interpreted to be the non-migmatitic precursor to the migmatite that is ubiquitous across the transect. It was collected from the north shore of Bark Lake immediately east of Depot Point. At this site, leucosome is locally absent but increasingly developed in the area surrounding it.

Three zircon fractions were analyzed. Morphologically, crystals were sub-rounded and equant; colourless, relatively clear and contained few inclusions (Figure 5.2f). Faint igneous zoning and some cores were observed. All fractions are discordant (3-6%) and do not fall on a regression line (Figure 5.5c; York, 1969). $^{207}\text{Pb}/^{206}\text{Pb}$ ages range from 1341-1413 Ma; the fractions are interpreted to represent a single population of crystals with a minimum age of 1413 Ma. Based on conspicuous zoning and crystal morphology, this is interpreted to represent protolith crystallization; the age is consistent with those obtained for igneous crystallization in the western Muskoka domain (~1390-1460 Ma; U-Pb zircon, Timmermann, 1998) and underlying Algonquin domain (1410-1475 Ma; U-Pb zircon, Nadeau and van Breeman, 1998). The $^{207}\text{Pb}/^{206}\text{Pb}$ ages show no evidence of >1450 Ma inheritance, suggesting that these rocks represent juvenile Laurentian crust.

5.4 Discussion of U-Pb results

Table 5.2 presents a summary of the age determinations for the field area.

Table 5-2. Summary of age results, Barry's Bay, QN.

Sample	Lithology	Age (U-Pb, zircon)	Interpretation
1. SM98-23T	migmatitic monzogranitic orthogneiss	1. 1514 $^{+40}_{-30}$ Ma 2. 1046 $^{+34}_{-39}$ Ma	1. igneous crystallization; 2. high grade meta- morphism , anatexis and associated Pb-loss
2. SM98-23P	syntectonic magnetite, biotite quartz syenite pegmatite	1036 $^{+10}_{-2.5}$ Ma	igneous crystallization
3. SM98-28	post-tectonic magnetite, biotite quartz syenite pegmatite	993 $^{+/-}_{-2}$ Ma	igneous crystallization
4. SM98-118	meta-megacrystic granodioritic orthogneiss	1. 1468 $^{+26}_{-17}$ Ma 2. 894 $^{+110}_{-115}$ Ma	1. igneous crystallization; 2. high-grade meta- morphism and Pb-loss.
5. SM98-129	garnet, orthopyroxene intermediate orthogneiss	1. 1499 $^{+50}_{-35}$ Ma 2. 1073 $^{+74}_{-85}$ Ma	1. igneous crystallization; 2. retrogression to amphibolite-facies.
6. SM98-149	non-migmatitic granodioritic orthogneiss	1413 Ma	inferred age of crystallization.

5.4.1 Plutonism

Emplacement of protolith plutonic rocks occurred between >1413 Ma and <1515 Ma. When all fractions from the granitoid orthogneiss samples (1, 4, 5, 6) are regressed together, 10 of the 15 fractions are collinear (Figure 5.5d). They fall on a discordia that closely approximates that defined by Sample 4 (SM98-118), the most tightly constrained and least discordant analysis. The discordia for this compilation of granitoid samples has an upper intercept of $1488 \pm 34/-26$ Ma and a lower intercept of $979 \pm 72/-28$ Ma. The collinearity of samples suggests they constitute a single population of crystals derived from one plutonic suite, with crystallization ca. 1480 Ma.

5.4.2 Metamorphism and Deformation

Penetrative ductile deformation was synchronous with, but outlasted by, regional metamorphism. The formation of at least two generations of anatectic melt (AL_1 , and AL_2) accompanied the ~ 1070 -1030 Ma upper amphibolite-facies metamorphism that affected the rocks of the transect (see Table 5-2). AL_1 is concordant within S_T , the penetrative transposition foliation that characterizes the transect, and is defined by coplanar gneissosity, attenuated layering, rootless isoclinal folds, and concomitant SE-plunging stretching and mineral lineations. Field observations of the subparallel alignment of hornblende porphyroblasts (to each other) within AL_1 that grew during anatexis, the highly attenuated, concordant nature of AL_1 with respect to the foliation suggest that the development of S_T is synchronous with, but outlasted, the first anatectic event. Sample 2, a $1036 \pm 10/-2.5$ Ma syntectonic pegmatite that cross-cuts the foliation, indicates active strain at this time; further, S_T must be older than this pegmatite and is thus interpreted to have formed progressively over the interval ~ 1070 -1040 Ma.

Asymmetric tight to open folding (F_{ST+1}) post-dates S_T and its associated structures. Its southwest- and northeast-plunge, at high angles to the collinear mineral and extension lineations, as well as the early F_{ST} fold axes is interpreted to reflect their formation late in the progressive ductile deformation, and are therefore interpreted to have formed at or shortly before ca. 1040 Ma.

In contrast to AL_1 , the second generation of leucosome, AL_2 , is blebby and discordant with respect to S_T , and often localized into late dilatational sites that cross-cut and overprint the foliation. AL_1 and AL_2 have the same compositions and are characterized by hornblende porphyroblasts, indicating formation under the same P-T and a- H_2O conditions, thus suggesting that upper amphibolite-facies metamorphism in the area outlasted penetrative ductile deformation. Amphibolite-facies conditions prevailed at ~1029 Ma (Carson Lake outcrop, Sample 5; Carr and Berman, 1997), and there was subsequent minor ductile strain.

Extension affected the Bark Lake area, resulting in the reactivation of pre-existing shear zones and the development of new ones that cross-cut S_T and localized AL_2 and late pegmatites. Based on the upper limit of metamorphism and the ~990 Ma crystallization age of a pristine post-tectonic pegmatite in the Bark Lake area (U-Pb zircon, SM98-28, this study), extension is interpreted as having occurred between ~1030 and ~990 Ma. This range is slightly younger than extension in the western Muskoka domain, which occurred at 1040-1020 Ma (Timmermann, 1998). Because AL_2 is localized in these late extensional sites, it likely formed synchronously with, or immediately following extension. It is thus interpreted as having formed at ca. 1030 Ma, during the latter phases of amphibolite-facies metamorphism. Regional deformation in the area was over by 993

Ma. Subsequent high-level brittle faulting (<990 Ma) was the last event to affect the Bark Lake area, which resulted in minor offset of pre-existing ductile structures and the formation of cataclasite.

5.4.3 Regional significance of protolith ages

The range of ages obtained for plutonism in this part of the southeastern Central Gneiss Belt are consistent with those throughout the upper structural levels of this belt. For example, granitic plutonism has been dated at ~1440-1470 Ma in the western Muskoka domain (Timmermann et al., 1997; Timmermann, 1998), and at ~1475-1410 Ma in the immediately underlying Algonquin allochthon (Nadeau, 1990; Nadeau and van Breeman, 1998). The absence of older (>1450 Ma) memory in these rocks in conjunction with the similarity of protolith crystallization ages to the 1.41-1.62 Ga Nd-model ages obtained throughout the Algonquin and Muskoka allochthons (Dickin and McNutt, 1989; 1990) affirm the juvenile nature of these rocks.

Sample 4, the megacrystic granitoid (SM98-118), was selected for U-Pb geochronology for two reasons: (1) these metaplutonic rocks have been correlated with important regional markers confined to rocks of Laurentian-margin affinity (Carr et al., in press); and, (2) near Dorset, Ontario, Timmermann (1998) obtained ages of $1484 \pm 24/-13$ Ma and 1444-1466 Ma for megacrystic granitoid in the McClintock assemblage of the structurally underlying Algonquin domain, and we wanted to see if they belonged to the same suite of rocks. The $1468 \pm 26/-17$ Ma and $894 \pm 110/-116$ Ma upper and lower intercepts, respectively, are the same, within error, as those obtained Timmermann (1998; above). Based on similar field relationships, mineralogy and ages obtained for the two megacrystic bodies, they are interpreted to belong to the same suite of rocks.

5.4.4 *Regional significance of the timing of metamorphism and deformation as determined in the Bark Lake area*

Like the protolith ages, the timing of coeval amphibolite-facies metamorphism and ductile deformation in the footwall are in close agreement, like the protolith ages, with that in the western Muskoka domain. Timmermann (1998) constrained metamorphism and associated anatexis to ca. 1060 Ma, synchronous with deformation. Timmermann (1998) also documented a pre-Grenvillian granulite-facies event in the western Muskoka domain at ~1430-1480. Although not representative of the rest of the transect, at Carson Lake (Sample 6; Map1, back pocket), granulite-facies assemblages containing hypersthene and garnet are patchily preserved within its retrograde amphibolite-facies equivalent. Previous workers (Bell and Blenkinsop, 1980; Easton and Davidson, 1994; Carr and Berman, 1997) have interpreted the field evidence as suggesting that the retrogression was spatially associated with local small-scale shearing and intrusion of late-syntectonic ca. 1029 Ma granitoid dykes (U-Pb zircon, Carr and Berman, 1997). Early Rb-Sr whole rock analyses at Carson Lake yielded an age of 1430 ± 20 Ma with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.7034 (Bell and Blenkinsop, 1980), which was interpreted as the pre-Grenvillian granulite-facies metamorphic age. Current U-Pb geochronology at this outcrop yielded no evidence that would suggest metamorphism at ca. 1450 Ma, despite field evidence of a granulite-facies event that appears to predate the main (i.e., Grenvillian) event. It is suggested that the granulite-facies assemblages at Carson Lake represent those rocks that escaped widespread retrogression during the main metamorphic and accompanying migmatization event.

U-Pb geochronology provides no evidence for an earlier, ~1190-1160 Ma Grenvillian event in this area, as required by some models which propose that the tectonic

juxtaposition of the CMB over the Laurentian margin took place as early as ca. 1190 Ma (Hanmer, 1988; Hanmer and McEachern, 1992; McEachern et al., 1992; Nadeau and Hanmer, 1993; McEachern and van Breeman, 1993). Geochronological analyses of the Bark Lake area do, however, provide substantial evidence for Grenvillian strain (penetrative ductile deformation) that was distributed through a thick structural section in the immediate footwall of a crustal-scale ductile shear zone, and was coeval with upper amphibolite-facies regional metamorphism and associated anatexis. The timing of the lithological juxtaposition, however, due to burial, metamorphism and deformation which obscures this boundary cannot be dated. The pre-Grenvillian tectonic evolution of the Bark Lake area between the time of plutonism at ~1450 Ma and Grenvillian deformation and metamorphism at ~1070-1030 Ma remains unconstrained.

Chapter 6. Tectonic Evolution of the Bark Lake area

6.1 Introduction

Detailed 1:25 000 field mapping, in conjunction with geochemistry and U-Pb geochronology have helped characterize the geology and primary tectonic setting, as well as the spatial and temporal evolution, of the rocks in the Bark Lake area. Based on a comparison of ages and lithotectonic affinities of the Bark Lake rocks and those from the western Muskoka domain as described by Timmermann (1998), the area is interpreted as representing the eastern extension of the Muskoka domain. Structural and geochronological studies indicate that penetrative ductile deformation and upper amphibolite-facies metamorphism occurred over the interval ~1070-1040 Ma (see Table 5-2), and that metamorphism outlasted deformation. In the southern part of the map-area, deformation was localized in the CMBbtz, which gave rise to the kilometer-scale strain gradient that characterizes the transect.

6.1 Primary Tectonic Setting

The Bark Lake transect is characterized by granitic orthogneiss of variable composition, from monzodioritic to monzogranitic with minor tonalite, granite and syenogranite (see Figures 2.1a-d; Map 2, back pocket). Preliminary geochemistry indicating calc-alkaline trends supports previous suggestions, based on mineral assemblages and compositions from southwest of the Bark Lake area (Timmermann, 1998), that rocks of the Muskoka domain originated in a collisional tectonic setting, such as an Andean-type continental arc. Similarity of protolith crystallization ages to

documented Nd-model ages for the Muskoka and Algonquin domains of the Central Gneiss Belt (1.62-1.41 Ga; Dickin and McNutt, 1989; 1990) indicate a juvenile nature for these rocks. Although not voluminous, mafic enclaves are widespread; they comprise attenuated amphibolite layers within granitic orthogneiss and cm- to dm-scale boudins (see Figures 2.2a-c).

Mafic rocks are predominantly amphibolites associated with minor ultramafic bodies, and are interpreted as pre-Grenvillian dykes and sills that, together with their host, were penetratively deformed. There is a sparse paragneiss unit that outcrops predominantly in the southern part of the transect (see Figure 2.4a-b; Map 2, back pocket), interpreted as disrupted country rocks into which the protolith to the granitic orthogneiss was intruded.

Distinct lithologies, such as meta-megacrystic granitoids and coronitic metagabbro (see Figure 2.5a-d), are useful tectonic discriminators; the former are regional markers restricted to rocks of pre-Grenvillian Laurentian margin affinity (Carr et al., 1999, and references therein), while the latter are restricted to allochthonous terranes (Ketchum and Davidson, 1999). Based on similar protolith ages (see Table 5.2) and field relationships, deformed meta-megacrystic granitoid gneiss in the Bark Lake area is interpreted as belonging to the ~1450-1480 suite that has been documented throughout the parautochthonous Britt (Ketchum and Davidson, 1999), and underlying Algonquin domains (Timmermann, 1998; Ketchum and Davidson, 1999), thus suggesting that the Muskoka domain may have been spatially associated with these domains at ca. 1460 Ma. Alternatively, the similarity in ages may be fortuitous; the rocks could have been widely separated from each other, existing different parts of the suggested Andean arc, but

experienced the same phase of plutonism. Coronitic (olivine) metagabbro confirms the allochthonous nature of the rocks at Bark Lake, as metagabbro of this variety (*Algonquin metagabbro*; Rivers, 1997) is restricted to the hanging wall of the Allochthon Boundary Thrust (Ketchum and Davidson, 1999), a first-order tectonic boundary that separates parautochthonous rocks in the footwall from the allochthonous hanging wall units (ABT; Rivers et al., 1989).

6.2 Deformation history and its relationship to metamorphism

Granitic plutonism within the southeast margin of pre-Grenvillian Laurentia occurred >1413 Ma and < 1514 Ma, likely ca. 1480 Ma (see Figure 5.5d). Mafic dykes and sills were intruded prior to the onset of Grenvillian Orogenesis (~1300-950 Ma) however their precise ages are unknown. With the exception of Algonquin metagabbro, which was intruded into the upper structural levels of the Central Gneiss Belt at ca. 1170 Ma (Davidson and van Breeman, 1988; Ketchum and Davidson, 1999), no Elzevirian-aged (~1250-1190 Ma; Rivers, 1997) events are recorded in the area.

Regional metamorphism was coeval with, but outlasted, penetrative ductile deformation. The timing of upper amphibolite-facies metamorphism is based on lower intercept values for the discordia plots of the orthogneiss samples, and from crystallization ages of syntectonic pegmatites that, in the field, are interpreted as being spatially associated with amphibolite-facies metamorphism; metamorphism is therefore interpreted to have occurred over the interval 1073-1029 Ma, and resulted in the formation of at least two generations of anatectic melt (AL_1 , and AL_2). This interval is consistent with, but more protracted than, the 1079-1063 Ma age range obtained by

Timmermann (1998) for upper amphibolite-facies metamorphism in the western Muskoka domain. AL_1 is flattened and attenuated, and concordant within S_T , the penetrative transposition foliation that characterizes the transect. S_T is defined by coplanar gneissosity, attenuated layering, rootless isoclinal folds, and concomitant SE-plunging stretching and mineral lineations. In light of the diachronous nature of transposition, the subparallel alignment of hornblende porphyroblasts (to each other) within leucosome of AL_1 (syn-anatexis), and the highly attenuated, concordant nature of AL_1 with respect to the foliation, the development of S_T is interpreted as being synchronous with, and outlasting, the first anatectic event. Based on cross-cutting syntectonic pegmatites (1036 Ma, U-Pb zircon, SM98-23P, this study; 1029 Ma, U-Pb zircon, BLB-95-13, Carr and Berman, 1997), S_T , and therefore a significant interval of the progressive ductile deformation that affected the Bark Lake area, occurred ca. 1040-1030 Ma (see section 5.4.2 for discussion).

AL_2 is blebby and discordant with respect to S_T and often localized into late dilatational sites that cross-cut and overprint the foliation. AL_1 and AL_2 have the same compositions and are characterized by coarse hornblende porphyroblasts, indicating formation under the same P-T and a-H₂O conditions, thus indicating that amphibolite-facies metamorphism in the area outlasted penetrative ductile deformation.

Asymmetric tight to open folding (F_{ST+1}) post-dates S_T and associated structures but predates shear band development and formation of AL_2 . F_{ST+1} is interpreted as reflecting the latter stages of progressive ductile deformation and does not represent a discrete episode of folding. The orientation of SW- and NE-plunging, orogen-parallel F_{ST+1} fold axes is interpreted to reflect both their initial orientation and time of formation

at the latter stages of the progressive deformation sequence, thus remaining at high angles to the orogenic front; this is in contrast to the SE-plunging fold axes within F_{ST} that, as a function of their earlier formation and initial orientation, have had time to become progressively oriented into parallelism with the finite extension lineation.

Extension affected the Bark Lake area, resulting in the reactivation of pre-existing shear zones and the development of new ones that cross-cut S_T and localized AL_2 and late pegmatites. Based on the upper limit of metamorphism and the ~990 Ma crystallization age of a pristine post-tectonic pegmatite in the Bark Lake area (U-Pb zircon, SM98-28, this study), extension is interpreted as having occurred between ~1030 and ~990 Ma, at least 10 Ma younger than extension in the western Muskoka domain at 1040-1020 Ma (Timmermann, 1998). Brittle faulting was the last event to affect the Bark Lake area, which resulted in minor offset of pre-existing ductile structures and the formation of cataclasite.

6.3 Discussion of Grenvillian deformation

The Bark Lake area represents a little-studied part of the immediate footwall to the Central Metasedimentary Belt boundary thrust zone (CMBbtz), a crustal-scale ductile shear zone along which the Central Metasedimentary Belt was tectonically juxtaposed over the Laurentian margin, now represented by the Central Gneiss Belt. Although kinematics and structures of the shear zone are well documented (Davidson and Morgan, 1981; Culshaw et al., 1982; Davidson, 1984; Hanmer and Ciesielski, 1984; Hanmer et al., 1985) the regional significance and timing of initial movement across it is incompletely constrained. Two tectonic models predominate in the literature:

(1) ***The break-back thrusting model:*** this model was developed primarily on the basis of thematic studies in the Huntsville region of the Algonquin domain, as well as the Haliburton and Killaloe segments of the CMBbtz (Nadeau, 1990; Hanmer, 1988; Hanmer and McEachern, 1992; Nadeau and Hanmer, 1993); in it, break-back (out-of-sequence) thrusting is interpreted to have played a dominant role in creating the present-day crustal architecture of the Central Gneiss Belt. In this model, the above authors propose marginal basin closure in the interval 1230-1190 Ma which resulted in thrusting on the CMBbtz at 1190 Ma. The onset of break-back thrusting within tectonically overthickened crust at 1160 Ma culminated in the intra-plate reactivation of this boundary at 1080-1050 Ma. Slow orogenic cooling and exhumation of the CGB is attributed to protracted break-back thrusting in the upper structural levels.

(2) ***The foreland propagation model:*** this model is based on comprehensive studies near Georgian Bay; it proposes normal, northwestward propagation of the orogen into its foreland by “piggy-back” thrusting of lithotectonic entities on progressively younger structures (Jamieson et al., 1992; Culshaw et al., 1994; Wodicka et al., 1996; Culshaw et al., 1997; Timmermann, 1998). In contrast to the first model, early deformation and metamorphism at 1190-1160 Ma is restricted to the CMB and Parry Sound Domain which are thought to have been located offshore of Laurentia, with an initial encounter between these outboard terranes and the craton which did not occur until ca. 1120 Ma. Northwest-directed thrusting of allochthonous terranes ensued during the interval 1080-1040 Ma, with thrusting ages younging toward the foreland. This model locally invokes break-back thrusting as a component of early orogenic construction

important within certain domains (e.g., the Algonquin domain, and the Moon River lobe), but suggests it did not play an important role on the regional scale.

Penetrative ductile deformation and regional upper amphibolite-facies metamorphism in the Bark Lake area are interpreted as the tectonothermal responses to crustal thickening by ductile thrusting of the Central Metasedimentary Belt over the SE margin of the Laurentian craton. While the pre-Grenvillian evolution of the Bark Lake area between the time of plutonism at ca. 1480 Ma until onset of penetrative deformation and regional upper amphibolite-facies metamorphism at ca. 1079-1029 Ma remains poorly understood, constraints from geology, geochemistry and U-Pb geochronology transect area indicate that there is no evidence of Elzevirian deformation or metamorphic events, other than the intrusion of Algonquin metagabbro. This suggests that the eastern Muskoka domain was not affected by Grenvillian Orogenesis until the main, or *Ottawan*, phase of the orogenic cycle (1080-1020 Ma; Rivers, 1997), consistent with the findings of Timmermann (1998) in the western Muskoka domain.

6.4 Conclusions

Although the pre-Grenvillian tectonic evolution of the Bark Lake area remains poorly constrained, there are significant conclusions which can be drawn from the current study:

- (1) Based on similar age and lithotectonic affinities of the Bark Lake area with the western Muskoka domain, this area is interpreted to represent the easternmost extension of this domain. The Bark Lake area, like the western Muskoka domain, is characterized

by upper amphibolite- to granulite-facies migmatitic granodioritic orthogneiss with Mesoproterozoic protolith crystallization ages (>1413 and <1515 Ma).

(2) The granitoids at Bark Lake represent a diachronous yet genetically related suite of ca. 1480 Ma calc-alkaline plutonic rocks that are interpreted to have originated in a collisional tectonic setting, such as an Andean-type continental arc, on the southeastern margin of pre-Grenvillian Laurentia.

(3) The eastern Muskoka domain was largely unaffected by Grenvillian Orogenesis until the Ottawa phase of the orogenic cycle (1080 -1020 Ma; Rivers, 1997), which resulted in penetrative ductile deformation and upper amphibolite-facies and associated anatexis in the interval ~ 1070 -1030 Ma. Ages of syn-tectonic pegmatites indicate active strain at ~ 1036 Ma; metamorphism outlasted deformation, and amphibolite-facies conditions prevailed at 1029 Ma, with only minor subsequent ductile strain.

(4) Anatexis was widespread and episodic, resulting in the formation of at least two generations of anatectic melt. Results from geochemistry indicate that migmatitic textures formed as a result of *in situ* partial melting, in a closed geochemical system. P-T conditions in the Bark Lake area (see Table 1-2) were insufficient to produce extensive partial melt under water-undersaturated, or low $a\text{-H}_2\text{O}$, conditions, strongly suggesting that the voluminous melt that anatexis took place under water-saturated, or high $a\text{-H}_2\text{O}$, conditions in the mid- to deep crust.

(5) Extension affected the area at ca. 1030-990 Ma, which was the final stage of regional deformation in the area. Subsequent steeply dipping brittle faults dissected the rocks upon cooling and exhumation, which occurred <990 Ma.

(6) Due to burial, metamorphism and deformation, the lithological juxtaposition of the CMB over the CGB has been obscured and therefore cannot be dated directly. However, U-Pb geochronological analyses of the immediate footwall to the CMBbtz in the Bark Lake area provide no evidence for ~1190-1160 Ma (*Elzevirian*) regional metamorphism or deformation, as required by the break-back thrusting model.

REFERENCES

- Adams, F.D., and Barlow, A.E. 1910. Geology of the Haliburton and Bancroft Areas, Ontario. Geological Survey of Canada Memoirs 6.
- Anovitz, L.M., and Essene, E.J. 1990. Thermobarometry and pressure-temperature paths in the Grenville Province, Ontario. *Journal of Petrology*, 31: 197-241.
- Appleyard, E.C. 1974. Basement/cover relationships within the Grenville province in eastern Ontario. *Canadian Journal of Earth Sciences*, 11: 369-379.
- Appleyard, E.C., and Stott, G.M. 1975. Grenville gneisses in the Madawaska Highlands, eastern Ontario. Geological Association of Canada, Field Trip No. 2, 61pp.
- Ashworth, J.R. (ed.) 1985. Migmatites. Glasgow, Blackie.
- Ashworth, J. R., and McLellan, 1985. Textures. *In*: Migmatites, J.R. Ashworth (ed). 180-203. Glasgow; Blackie.
- Buddington, A.F. 1939. Adirondack igneous rocks and their metamorphism. *Geological Society of America Bulletin*, 7.
- Bell, K., and Blenkinsop, J. 1980. Whole rock Rb/Sr studies in the Grenville Province of southeastern Ontario and western Quebec – a summary report. *In* Current Research, Part C, Geological Survey of Canada, Paper 80-1C: 152-154.
- Bethune, K.M. 1997. The Sudbury dyke swarm and its bearing on the tectonic development of the Grenville Front, Ontario, Canada. *Precambrian Research*, 85: 117-146.
- Breaks, F.W., and Thivierge, R.H. 1985. Geology of the Wicklow area. Ontario Geological Survey, Open File Report 5536, 256pp.
- Brown, G.C. and Fyfe, W.S. 1972. The transformation from metamorphism to melting: status of the granulite and eclogite facies. *Proceedings of the 24th International Geologic Conference, Montreal, Section 2*: 27-34.
- Burr, J.L., and Carr, S.D. 1994. Structural geometry and U-Pb geochronology near Lithoprobe Seismic Line 32, Western Central Metasedimentary Belt, Grenville Province, Ontario. Lithoprobe Abitibi-Grenville Transect Report, 41: 59-62.
- Burr, J.L., and Carr, S.D. 1992. Structural analysis and U-Pb geochronology in the Bancroft area near Lithoprobe profiles 32 and 33: A progress report. Lithoprobe Abitibi-Grenville Transect Report 26: 141-144.

- Bussy, F., Krogh, T.E., Klemens, W.P., and Schwerdtner, W.M. 1995. Tectonic and metamorphic events in the westernmost Grenville Province, central Ontario: New results from high-precision U-Pb zircon geochronology. *Canadian Journal of Earth Sciences*, **32**: 660-671.
- Carr, S.D., Easton, R.M., Jamieson, R.A., and Culshaw, N.G. 1999. Geologic transect across the Grenville Orogen of Ontario and New York. *Canadian Journal of Earth Sciences*, **36**: In press.
- Carr, S.D., and Berman, R.G. 1997. Metamorphic history of the Central Metasedimentary Belt boundary thrust zone and adjacent footwall rocks in the Algonquin domain (Central Gneiss Belt): Bancroft- Whitney - Barry's Bay area, Ontario Grenville. Geological Association of Canada-Mineralogical Association of Canada, Annual Meeting. Program with abstracts.
- Chappell, B.W. and White, A.J.R. 1974. *Pacific Geology*, **8** 173~174.
- Ciesielski, A. and Sharma, K.N.M. 1993. Geochemistry and petrogenesis of 1280 Ma tonalitic orthogneiss of the Central Metasedimentary Belt of the Grenville Province, Quebec. (unpublished date for submission to the *Journal of Geology*).
- Corfu, F. and Easton, R.M. 1995. U-Pb geochronology of the Mazinaw terrane, an imbricate segment of the Central Metasedimentary Belt, Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **32**: 959-976.
- Corriveau, L., Heaman, L.M., Marcantonio, F., and Van Breemen, O. 1990. 1.1 Ka K-rich alkaline plutonism in the SW Grenville Province, U-Pb constraints for the timing of subduction-related magmatism. *Contributions to Mineralogy and Petrology*, **105**: 473-485.
- Culshaw, N.G., Davidson, A., and Nadeau, L. 1983. Structural subdivisions of the Grenville Province in the Parry Sound-Algonquin region, Ontario. In *Current Research, Part B*, Geological Survey of Canada. Paper 83-1B: 243- 252.
- Culshaw, N.G., Ketchum, J.W.F., Wodicka, N., and Wallace, P. 1993. Deep crustal ductile extension following thrusting in the southwestern Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **31**: 160-175.
- Culshaw, N.G., Jamieson, R.A., Ketchum, J., Wodicka, N., Corrigan, D., and Reynolds, P.H. 1997. Transect across the northwestern Grenville Orogen, Georgian Bay, Ontario: Polystage convergence and extension in the lower orogenic crust. *Tectonics*, **16**, 6: 966-982.
- Davidson, A. 1995. A review of the Grenville orogen in its North American type area. *AGSO Journal of Australian Geology & Geophysics*, **16**: 3-24.

- Davidson, A. 1986. A new look at the Grenville Front in Ontario. Geological Association of Canada, Field Trip Guidebook 15, 31pp.
- Davidson, A. 1984. Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield. in *Precambrian Tectonics Illustrated*. Edited by Kroner, A., and Greiling, R. Schweizerbart'sche Verlagsbuchhandlung (Nägele u. Obermeiller), Germany, Stuttgart.
- Davidson, A., and Morgan, W.C. 1981. Preliminary notes on the geology east of Georgian Bay, Grenville Structural Province, Ontario. *In Current Research, Geological Survey of Canada, Paper 81-1A*: 291-298.
- Davidson, A., and Grant, S.M. 1986. Reconnaissance geology of western and central Algonquin Park and detailed study of coronitic olivine metagabbro, Central Gneiss Belt, Grenville Province of Ontario; *In Current Research, Part B, Geological Survey of Canada, Paper 86-1B*: 837-848.
- Davidson, A., and van Breeman, O. 1988. Baddeleyite-zircon relationships in coronitic metagabbro, Grenville Province, Ontario: Implications for geochronology. *Contributions to Mineralogy and Petrology*, 100: 291-299.
- Davidson, A., Culshaw, N.G., and Nadeau, L. 1982. A tectono-metamorphic framework for part of the Grenville province, Parry Sound region, Ontario. *In Current Research, Part A, Geological Survey of Canada, Paper 82-1A*: 175-190.
- Davidson, A., Nadeau, L., Grant, S.M., and Pryer, L.L. 1985. Studies in the Grenville Province of Ontario. *In Current Research, Geological Survey of Canada, Paper 85-1A*: 463-483.
- Dickinson, A.P. and McNutt, R.H. 1989. Nd model age mapping of the southeast margin of the Archean foreland in the Grenville Province of Ontario. *Geology*, 7: 299-303.
- Dickinson, A.P. and McNutt, R.H. 1990. Nd-model age mapping of Grenville lithotectonic domains: mid-Proterozoic crustal evolution in Ontario. *In Mid-Proterozoic Laurentia-Baltica*. edited by C.F. Gower, T. Rivers, and A.B. Ryan. Geological Association of Canada, Special Paper 38; pp. 79-94.
- Easton, R.M. 1986. Geochronology of the Grenville Province. *In The Grenville Province. Edited by J.M. Moore, A. Davidson, and A.J. Baer. Geological Association of Canada, Special Paper 31*: 127-173.
- Easton, R.M. 1992. The Grenville Province and the Proterozoic history of central and southern Ontario. in *Geology of Ontario*. Thurston, P.C., Williams, H.R., Sutcliffe, R.H., and Stott, G.M. (eds.) Ontario Geological Survey, Special Volume 4, Part 2: 715-904.

- Easton, R.M., and Davidson, A. 1994. Terrane boundaries and lithotectonic assemblages within the Grenville Province, eastern Ontario. Geological Association of Canada, Field Trip A1 Guidebook, 89pp.
- England, P.C. and Thompson, A. B. 1984. Pressure-temperature-time paths of regional metamorphism, Part I: Heat transfer during the evolution of regions of thickened continental crust. *Journal of Petrology*, **25**: 894-928.
- Escher, A. and Watterson, J. 1974. Stretching fabrics, folds and crustal shortening. *Tectonophysics*, **22**: 223-231.
- Flinn, D. 1962. On folding during three-dimensional progressive deformation. *Geological Society of London, Quarterly Journal*. **118**: 385-433.
- Green, A.G., Milkereit, B., Davidson, A., Spencer, C., Hutchinson, D.R., Cannon, W.F., Lee, M.W., Ager, W.F., Behrendt, J.C., and Hinze, W.J. 1988. Crustal structure of the Grenville Front and adjacent terranes. *Geology*, **16**: 788-792.
- Hanmer, S. 1988. Ductile thrusting at mid-crustal levels, southwestern Grenville Province. *Canadian Journal of Earth Sciences*, **25**: 1049-1059.
- Hanmer, S., and Ciesielski, A. 1984. A structural reconnaissance of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario and Quebec. *In Current Research, Part B, Geological Survey of Canada, Paper 84-1B*: 121-131.
- Hanmer, S., and McEachern, S.J. 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville Orogen, Canada. *Canadian Journal of Earth Sciences*, **29**: 1779-1790.
- Hanmer, S., Thivierge, R.H., and Henderson, J.R. 1985. Anatomy of a ductile thrust zone: part of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario. *In Current Research, Part B, Geological Survey of Canada, Paper 85-1B*: 1-5.
- Heaman, L.M., and Parrish, R.R., 1991. U-Pb geochronology of accessory minerals. *In Applications of radiogenic isotope systems to problems in geology*. Heaman, L.M., and Ludden, J.J. (eds.) MAC Short Course Handbook, 19.
- Hoffman, P.F. 1991. Grenville Front (Canadian Shield) as a Windward Mountain Front. *Eos*, **72**: 497.
- Hobbs, B. E., Means, W. D., and Williams, P.F. 1976. *An Outline of Structural Geology*. New York, John Wiley & Sons.

- Irvine, 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.
- Jamieson, R.A., and Beaumont, C. 1989. Deformation and metamorphism in convergent orogens: a model for uplift and exhumation of metamorphic terrains. *In* *Evolution of Metamorphic Belts*. Daly, J.S., Cliff, R.A., and Yardley, B.W.D. (eds.), Geological Society Special Publication, **43**: 117-129.
- Jamieson, R.A., Culshaw, N.G., and Corrigan, D. 1995. North-west propagation of the Grenville orogen: Grenvillian structure and metamorphism near Key Harbour, Georgian Bay, Ontario, Canada. *Journal of Metamorphic Geology*, **13**: 185-207.
- Jamieson, R.A., Culshaw, N.G., Wodicka, N., and Corrigan, D., and Ketchum, J.W.F. 1992. Timing and tectonic setting of Grenvillian metamorphism -constraints from a transect along Georgian Bay, Ontario. *Journal of Metamorphic Geology*, **10**: 321-332.
- Johannes, W. 1984. Beginning of melting in the granite system Qz-Or-Ab-An-H₂O. *Contributions to Mineralogy and Petrology*, **86**: 264-73.
- Ketchum, J.W.F., and Davidson, A. 1999. Crustal architecture and tectonic assembly of the Central Gneiss Belt, southwestern Grenville Province, Canada – a new interpretation. *Canadian Journal of Earth Sciences*, **36**: In press.
- Ketchum, J.W.F., Wodicka, N., Culshaw, N.G., and Jamieson, R.A. 1999. Polystage convergence and extension in the mid- to lower orogenic crust: An examination of the Central Gneiss Belt, Grenville Province, along Georgian Bay, Geological Association of Canada. Field Trip B10 Guidebook, 65pp.
- Kretz, R. 1983. Symbols for rock-forming minerals. *American Mineralogist*, **68**: 277-279.
- Krogh, T.E. 1982. Improved accuracy of U-Pb ages by the creation of more concordant systems using an air abrasion technique. *Geochimica et Cosmochimica Acta*, **46**: 637-649.
- Lumbers, S. 1982. Summary of Metallogeny: Renfrew County Area. Ontario Geological Survey Report 212.
- Lumbers, S.B., and Vertolli, V.M. 1991. Proterozoic plutonism within the Central Gneiss Belt and Grenville Front Tectonic zone, Grenville Province, Ontario. *In* *Program with Abstracts*, Geological Association of Canada, **9**: A85.
- Lumbers, S.B., Heaman, L.M., Vertolli, V.M., and Wu, T.W. 1990. Nature and timing of Middle Proterozoic magmatism in the Central Metasedimentary Belt, Ontario.

In Mid-Proterozoic Laurentia-Baltica. Edited by C. Gower, T. Rivers, and A.B. Ryan. Geological Association of Canada, Special Paper 38: 243-276.

- Maniar, P.D. and Piccoli, P.M. 1989. Tectonic discrimination of granitoids. *Geological Society of America Bulletin*, **101**: 635-643.
- Mawer, C.K., and Williams, P.F. 1991. Progressive folding and foliation development in a sheared, coticule-bearing phyllite. *Journal of Structural Geology*, **13**: 539-555.
- McEachern, S.J., and van Breeman, O. 1993. Age of deformation within the Central Metasedimentary Belt boundary thrust zone, southwest Grenville Orogen: constraints on the collision of the Mid-Proterozoic Elzevir terrane. *Canadian Journal of Earth Sciences*, **30**: 1155-1165.
- McEachern, S.J., Hanmer, S., and van Breemen, O. 1992. The raglan metagabbro belt, Elzevir terrane: its age and implications for marginal basin crust within the southwest Grenville. 1992 Grenville Workshop, Abstracts: 49.
- Mehnert, K. R. 1968. *Migmatites and the Origin of Granitic Rocks*. Amsterdam: Elsevir.
- Mezger, K., van der Pluijm, B.A., Essene, E.J., and Halliday, A.N. 1991. Synorogenic collapse: A perspective from the Middle Crust, the Proterozoic Grenville Orogen. *Science*, **254**: 695-698.
- Myers, J.S. 1978. Formation of banded gneisses by deformation of igneous rocks. *Precambrian Research*, **6**: 43-64.
- Nadeau, L. 1990. Tectonic, thermal and magmatic evolution of the Central Gneiss Belt, Huntsville region, southwestern Grenville orogen. Unpublished Ph.D. thesis, Carleton University, Ottawa, Ontario, 269pp.
- Nadeau, L., and Hanmer, S. 1992. Deep-crustal break-back thrusting and slow exhumation of the continental footwall beneath a thrust marginal basin, Grenville Orogen, Canada. *Tectonophysics*, **210**: 215-233.
- Nadeau, L. and van Breeman, O. 1998. Plutonic ages and tectonic setting of the Algonquin and Muskoka allochthons, Central Gneiss Belt, Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **35**: 1423-1438.
- Nakamura, N., 1974. Determination of REE, Ba, Fe, Mg, Na and K in carbonaceous and ordinary chondrites. *Geochimica et Cosmochimica Acta*, **38**: 757-775.

- 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 isotopic studies: Report 1. Geological Survey of Canada, Paper 87-2: 3-7.
- Pearce, J.A., Harris, N.B.W., and Tindle, A.G. 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. 25, 4: 956-983.
- Pidgeon, R.T. 1992. Recrystallization of oscillatory zoned zircon: some geochronological and petrological implications. Contributions of Mineralogy and Petrology, 110: 463-472.
- Ramsay, J.G. 1967. Fracturing and folding of rocks. New York: MacGraw Hill.
- Rivers, T. 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications. Precambrian Research, 86: 117-154.
- Rivers, T., Martignole, J., Gower, C.F., and Davidson, A. 1989. New tectonic divisions of the Grenville Province, southeast Canadian Shield. Tectonics, 8: 63-84.
- Roddick, J. C., Loveridge, W.D., and Parrish, R.R. 1987. Precise U-Pb dating of zircon at the sub-nanogram level. Chemical Geology (Isotope Geoscience Section), 66: 111-121.
- Sawyer, E.W. 1996. Melt segregation and magma flow in migmatites: implications for the generation of granite magmas. Transactions of the Royal Society of Edinburgh: Earth Sciences, 87: 85-94.
- Sawyer, E.W. 1998. Formation and evolution of granite magmas during crustal reworking: the significance of diatexites. Journal of Petrology, 39: 1147-1167.
- Scammell, RJ 1993. Mid-Cretaceous to Tertiary thermotectonic history of former mid-crustal rocks, southern Ominica Belt, Canadian Cordillera. Ph.D. thesis, Queen's University, Kingston, ON.
- Streckeisen, A. 1976. To each plutonic rock its proper name. Earth-Science Reviews, 12: 1-33.
- Thompson, A.B. and England, P.C. 1984. Pressure-Temperature-Time paths of regional metamorphism, Part II: their inference and interpretation using mineral assemblages in metamorphic rocks. Journal of Petrology, 25: 929-955.
- Timmermann, H., Parrish, R.R., Jamieson, R.A., and Culshaw, N.G. 1997. Time of metamorphism beneath the Central Metasedimentary Belt boundary thrust zone, Grenville Orogen, Ontario: accretion at 1080 Ma? Canadian Journal of Earth Sciences, 34: 1023-1029.

- Timmermann, H. 1998. Geology, metamorphism, and U-Pb geochronology in the Central Gneiss Belt between Huntsville and Haliburton, southwestern Grenville Province, Ontario. Ph.D. thesis, Dalhousie University, Halifax, NS.
- Twiss, R.J. and Moores, E. M. Structural Geology. New York: W.H. Freeman and Company.
- van Breeman, O., and Hamner, S. 1986. Zircon morphology and U-Pb geochronology in active shear zones: studies on syntectonic intrusions along the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario. *In* Current Research. Geological Survey of Canada. Paper 86-1B: 775-784.
- van Breeman, O. and Davidson, A. 1988. U-Pb zircon ages of granites and syenites in the Central Metasedimentary Belt, Grenville Province, Ontario. *In*: Radiogenic age and Isotopic Studies: Report 2. Geological Survey of Canada, Paper 88-2: 45-50.
- van Breeman, O., and Davidson, A. 1990. U-Pb zircon and baddeleyite ages from the Central Gneiss Belt, Ontario. *In* Radiogenic Age and Isotopic Studies: Report 3, Geological Survey of Canada. Paper 89-2: 85-92.
- van Breeman, O., Davidson, A., Loveridge, W.D., and Sullivan, R.W. 1986. U-Pb geochronology of Grenville tectonites, granulites and igneous precursors, Parry Sound, Ontario. *In*: The Grenville Province. Edited by J.M. Moore, A. Davidson, and A.J. Baer. Geological Association of Canada Special Paper, 31: 191-207.
- Wayne, D.M., and Sinha, A.K. 1988. Physical and chemical response of zircons to deformation. Contributions to mineralogy and petrology. 98: 109-121.
- White, D.J., Easton, R.M., Culshaw, N.G., Milkereit, B., Forsythe, D.A., Carr, S.D., Green, A.G., and Davidson, A. 1993. Seismic images of the Grenville Orogen in Canada. Canadian Journal of Earth Sciences, 31: 293-307.
- Williams, P.F. 1985. Multiply deformed terrains-problems of correlation. Journal of Structural Geology, 7: 269-280.
- Williams, P.F. 1983. Large scale transposition by folding in northern Norway. Geologische Rundschau, 72: 589-604.
- Williams, P.F., and Zwart, H.J. 19--. A model for the development of the Seve-Koli Caledonian nappe complex.

- Wilson, M.E. 1918. The subprovincial limitations of Precambrian nomenclature in the St. Lawrence basin. *Journal of Geology*, **26**: 325-333.
- Wilson, M.E. 1925. The Grenville pre-Cambrian subprovince. *Journal of Geology*, **33**: 389-407.
- Wodicka, N., Parrish, R.R., and Jamieson, R.A. 1996. The Parry Sound domain: a far-traveled allochthon? New evidence from U-Pb zircon geochronology. *Canadian Journal of Earth Sciences*, **33**: 1087-1104.
- Wynne-Edwards, H.R. 1972. The Grenville Province. *In: Variations in Tectonic Styles in Canada. Edited by R.A. Price and R.J.W. Douglas. Geological Association of Canada, Special Paper 11*: 263-334.
- York, D. 1969. Least squares fitting of a straight line with correlated errors. *Earth and Planetary Science Letters*, **5**: 320-324.

Appendix A: Major Element Concentrations (wt%)

Geochemical																	
Sample	Rock Name*	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	H ₂ O	CO ₂	LOI	Mg#	
SM98-23T(m)	quartz-monzodiorite	56.42	0.97	19.05	2.96	3.42	0.13	2.03	5.27	5.50	2.37	0.41	0.31	0.17	0.56	37.22	
SM98-23T(l)	diorite	56.16	0.62	20.09	1.88	2.76	0.08	2.09	5.34	6.16	2.07	0.46	0.36	0.45	1.16	45.55	
SM98-69(m)	granodiorite	67.57	0.62	15.10	2.53	2.56	0.07	1.40	3.18	3.72	2.02	0.15	0.25	0.22	0.77	34.03	
SM98-69(l)	granodiorite	67.97	0.48	15.41	2.16	2.09	0.05	1.11	3.24	4.02	2.10	0.16	0.27	0.33	0.95	32.9	
SM98-112	monzogranite	75.18	0.14	12.97	0.61	0.53	n/d	0.27	0.80	2.76	5.99	0.04	0.12	0.19	0.62	30.84	
SM98-118	quartz-monzodiorite	59.33	0.85	18.83	2.34	2.65	n/d	1.63	4.16	5.20	3.49	0.29	0.06	0.07	0.34	37.92	
SM98-106	gabbro	50.71	0.74	18.94	3.46	5.59	n/d	4.94	9.22	4.26	0.61	0.20	0.12	0.28	0.64	50.28	
SM98-88D	quartz norite	54.21	2.19	15.32	4.34	6.31	n/d	4.17	6.55	3.57	1.25	0.43	0.15	0.48	0.83	42.11	
SM98-129	quartz diorite	55.04	1.21	18.58	2.06	5.51	0.12	2.25	5.31	4.24	2.82	0.37	0.07	b.d.l.	0.46	35.25	

*based on Streckeis (1976)

n/d = not detected

b.d.l. = below detection limits

Appendix B: Trace Element Concentrations (ppm)

Sample*	Cr	Ni	Co	Sc	V	Cu	Pb	Zn	Rb	Cs	Ba	Sr	Ta	Nb	Hf	Zr	Y	Th	U
SM98-23T(m)	8	12	12	15	94	16	16	143	48	0.32	1167	867	n/d	10.0	n/d	385	43	3.80	0.58
SM98-23T(l)	7	10	10	26	51	15	20	122	48	0.26	479	644	n/d	11.8	n/d	459	82	1.77	0.31
SM98-69(m)	14	12	9	9	68	18	12	98	73	1.21	581	415	n/d	8.1	n/d	304	17	6.10	1.00
SM98-69(l)	11	10	6 _u	5	63	9	14	84	67	1.08	632	461	n/d	7.2	n/d	269	15	14.70	1.10
SM98-112	n/d	b.d.l.	b.d.l.	1	10	42	n/d	22	119	0.20	n/d	451	0.08	2.9	3.25	79	10	4.69	0.51
SM98-118	n/d	9	9	12	9	14	n/d	117	86	0.60	n/d	724	0.71	15.1	3.64	111	46	4.96	0.74
SM98-106	n/d	59	33	21	60	39	n/d	74	3	0.04	n/d	406	0.13	2.6	1.29	34	26	0.72	0.26
SM98-88D	n/d	65	35	15	178	87	n/d	142	24	0.28	n/d	933	0.38	7.2	0.86	18	26	1.42	0.34
SM98-129	n/d	15	16	15	70	21	n/d	108	58	0.22	n/d	589	0.58	14.6	2.57	83	43	6.88	0.43

*corresponding rock names are those listed in Appendix A

n/d = not detected

b.d.l. = "below detection limit"

Appendix C: Rare Earth Element (REE) Concentrations (ppm)

Sample*	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Be
SM98-23T(m)	41.26	88.22	11.28	49.51	10.19	2.28	8.80	1.26	7.76	1.46	4.11	0.53	3.55	0.55	b.d.l.
SM98-23T(l)	62.40	201.79	20.92	93.91	20.08	2.48	17.20	2.48	14.91	2.80	7.78	0.94	5.42	0.73	4
SM98-69(m)	35.96	73.48	8.85	34.13	5.86	1.19	4.18	0.54	3.28	0.60	1.88	0.26	1.89	0.30	b.d.l.
SM98-69(l)	56.81	111.38	11.97	42.20	6.42	1.24	3.93	0.48	3.07	0.52	1.64	0.20	1.34	0.21	b.d.l.
SM98-112	27.42	46.91	5.79	20.64	3.40	0.77	2.53	0.35	2.01	0.36	0.93	0.11	0.66	0.10	b.d.l.
SM98-118	60.47	122.86	15.30	55.91	9.69	2.23	8.01	1.24	7.56	1.55	4.64	0.70	4.49	0.72	b.d.l.
SM98-106	14.18	32.61	4.61	19.90	4.38	1.24	4.28	0.67	4.37	0.92	2.60	0.39	2.52	0.39	b.d.l.
SM98-88D	24.93	54.40	7.84	34.82	7.75	2.67	6.90	0.94	5.40	0.97	2.43	0.32	1.94	0.28	b.d.l.
SM98-129	62.82	124.77	15.80	59.11	10.46	3.49	8.55	1.29	7.56	1.52	4.46	0.62	4.25	0.66	b.d.l.
<hr/>															
Total LREE (La to Sm)=	1770.6														
Total HREE (Gd to Lu)=	204.76														

*corresponding rock names are those listed in Appendix A
b.d.l.= "below detection limits"

**Appendix D: Chondrite normalizing factors
(Nakamura, 1974)**

Element	Normalizing Factor
La	0.329
Ce	0.865
Pr*	0.13
Nd	0.63
Sm	0.203
Eu	0.077
Gd	0.276
Tb*	0.0498
Dy	0.343
Ho*	0.077
Er	0.225
Tm*	0.0352
Yb	0.220
Lu	0.0339

*denotes factors provided by G.Jenner
(Currently at the Memorial University of
Newfoundland)

NOTE TO USERS

Oversize maps and charts are microfilmed in sections in the following manner:

**LEFT TO RIGHT, TOP TO BOTTOM, WITH
SMALL OVERLAPS**

UMI

5021750m. N.

271250m. E.

Map 3: Structural geology across the Bark Lake Transect

Scale: 1: 50 000

metres

miles

Structural symbols

- strike and dip of cross-cutting pegmatite dykes
- strike and dip of S_T
- trend and plunge of discrete extension lineations
- trend and plunge of penetrative extension lineations
- trend and plunge of mineral lineations
- trend and plunge of fold axes within F_{ST} ; M, Z, S
- indicate vergence
- trend and plunge of fold axes within F_{ST+1} ; M, Z, S
- indicate vergence
- strike and dip of shear zones
- crustal-scale shear zone; balls on the hanging wall side

Geographic symbols

- main road or highway
- secondary road
- dirt roads and trails
- 60 provincial highway
- swamps and boggy areas

NOTE TO USERS

Oversize maps and charts are microfilmed in sections in the following manner:

**LEFT TO RIGHT, TOP TO BOTTOM, WITH
SMALL OVERLAPS**

UMI

metres

5021750m. N.

271250m. E.

NOTE TO USERS

Oversize maps and charts are microfilmed in sections in the following manner:

**LEFT TO RIGHT, TOP TO BOTTOM, WITH
SMALL OVERLAPS**

UMI

Map 2 Geology of the Bark Lake area

Sheet 31 F/5

Scale: 1: 50 000

Legend

Quartz syenite intrusive rocks

a biotite, magnetite +/- hornblende-bearing
pre-, syn-, and post-tectonic pegmatites

b orthopyroxene-bearing pegmatites; interpreted
as very late syn-, to post-tectonic

Mafic rocks of uncertain protolith

a disrupted and attenuated biotite-bearing
amphibolite layers and boudins, locally
containing garnet, clinopyroxene and/or

5021750m. N.