METAMORPHISM, STRUCTURE, AND TECTONICS OF THE EASTERN CAPE SMITH BELT, NORTHERN QUEBEC.

by

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A THESIS

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My gold is sunshine and my gems,

The glint of dew on grassy stems...

Robert Service

ABSTRACT

The Cape Smith Belt of northernmost Quebec is a sedimentary/volcanic belt at the boundary of the dominantly gneissic Churchill and Superior structural provinces.

Within the Belt, three distinct phases of deformation and a metamorphic culmination are evident. Thrusting and imbrication of Belt rocks lead to the thermal culmination, a pattern similar to numerous Fold-Thrust Belts, including the eastern Alps. This is followed by an episode of coaxial folding. A late cross-folding event provides up to 12km of structural relief in the study area, at the eastern end of the Belt.

Metamorphism in the study area is of the hot-side-down Barrovian type. A systematic progression is seen from lower-greenschist to mid-amphibolite facies, from west to northeast across the study area. Minimum temperatures of 575°C and pressures of 5.5 kbar are suggested by the highest grade assemblages in pelitic rocks in the northeastern corner of the Belt.

Shallow-water ironstone and carbonate, deeper water siliciclastics, and rift volcanics are present stratigraphically from bottom to top. Restored to a pre-deformation state, this sequence is believed to represent a transition from passive margin to crustal rifting, from on- to off-shore, subsequently imbricated during collision of two crustal blocks. The rocks of the Belt are allochthonous, with the excption of those on the southern margin, having been thrusted southward onto the Superior craton during imbrication.

These findings are consistent with the current tectonic model for this area.

ACKNOWLEDGEMENTS

I would like to thank Dr. R.H. McNutt for his supervision of this study. His encouragement throughout the school year, and his constructive criticisms of preliminary written installments kept this study on track.

Dr. M.R. St-Onge of the Geological Survey of Canada suggested the study topic, and provided an exceptional amount of direction, field support, and enthusiastic encouragement during a thoroughly enjoyable field season. Marc also provided many helpful suggestions during the school year. Steve Lucas, Normand Begin, and Janien Schwarz, the rest of the field crew, helped make the summer so memorable.

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CHAPTER 1: INTRODUCTION

1.1 Purpose of Study

Among rocks of the Churchill structural province of the Canadian Shield, those of the Cape Smith Belt, northern Quebec, (see Figs. 1-1 and 1-2) remain unresolved with respect to their origin and relationship to the Churchill and Superior provinces. In 1985, the Geological Survey of Canada initiated a three-year, detailed study of the structure and lithology of the eastern end of the Cape Smith Belt, with the goal of establishing the structural relationship of those rocks to the underlying basement complex, and deducing the tectonic history of the area as a whole.

This thesis study, based on field observations made during the 1985 field season and subsequent laboratory investigations focuses on the lithology, metamorphic grade and structural style of the basal, dominantly sedimentary rocks at the eastern end of the Belt. The field study covered roughly 700 km^2, and contained numerous locations where stratigraphic sections could be measured in detail and sampled extensively.

1.2 Location and Access

The Cape Smith-Wakeham Bay Belt (Bergeron, 1957) is located on the Ungava Peninsula of northernmost Quebec (see Figs. 1-1, 1-2).

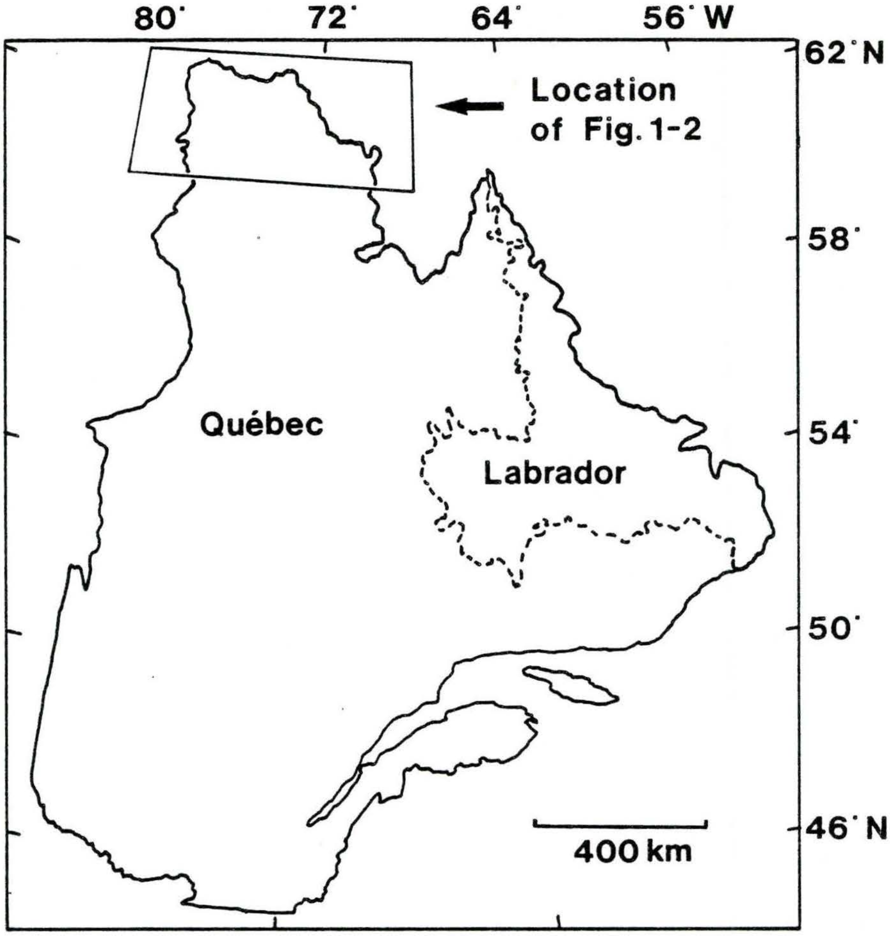
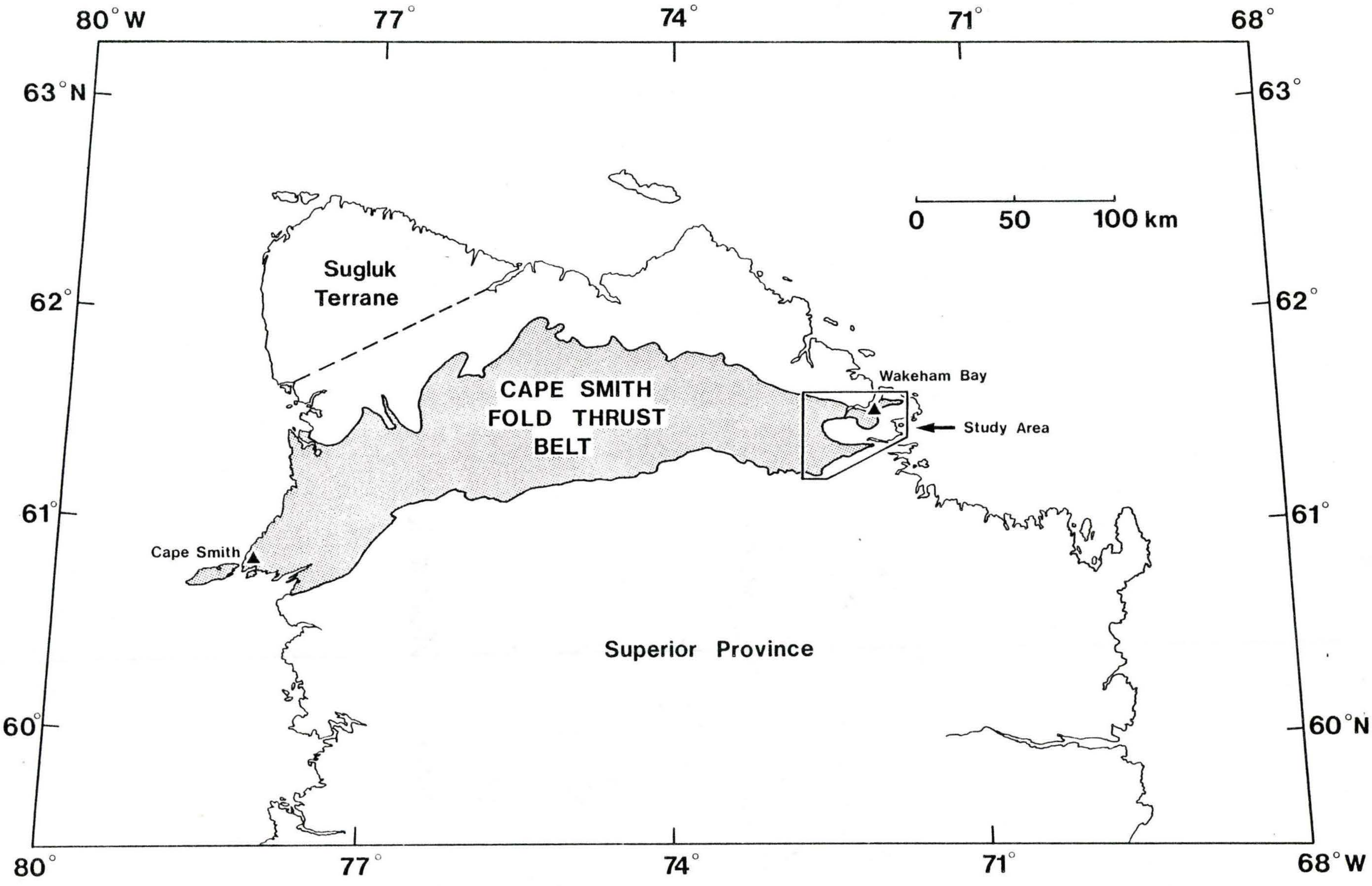


Figure 1-1: **Provice of Quebec and Labrador**, showing location of Figure 1-2.

Figure 1-2: **Location of Cape Smith Fold-Thrust Belt.** Outlined Study Area is shown in Figure 1-3.


80° W
77°
74°
71°
68°
63° N
63°
62°
62°
61°
61°
60°
60° N
80°
77°
74°
71°
68° W
0
50
100 km
Sugluk
Terrane
CAPE SMITH
FOLD THRUST
BELT
Wakeham Bay
Study Area
Cape Smith
Superior Province

The Belt trends roughly east-west and covers nearly 16 000 km^2. Cape Smith is a settlement at the western end of the Belt, while Wakeham Bay Village is located at its eastern end.

The metasediments are located around the perimeter of the Belt. This study concentrates on those found east of 72° 45′ W (Fig. 1-3).

The field work was done during the summer of 1985, when the author was employed as a Field Assistant, under the direction of Dr. M.R. St-Onge.

The logistical base for the study is Fort Chimo (Kuujjuaq), Quebec, serviced several times weekly by Nordair from Montreal and Quebec City. The study area is reached by scheduled Twin Otter service from Fort Chimo to Wakeham Bay Village (Kangiqsujuaq). Base camps were established by helicopter set-out. Daily traversing was by foot.

1.3 Previous Work

The Cape Smith-Wakeham Bay Belt was initially mapped at 1:1,013,760 in 1957 by the Quebec Department of Mines (Bergeron, 1957), following three years of private mineral exploration. Portions of the belt were mapped at 1:63,360 (De Montigny, 1959; Beall, 1959, 1960; Gold, 1962; Gelinas, 1962; Westra, 1972) (see Fig. 1-4). The entire region north of 61° N was mapped at 1:250,000 by

Figure 1-3: **Geological Map of Study Area,** Eastern Cape Smith Belt,

Wakeham Bay - Lac de la Grunerite Area. Thesis study focuses on Formations 1-4, described in detail in the text. (after St-Onge et al, 1986, Fig. 1.1).

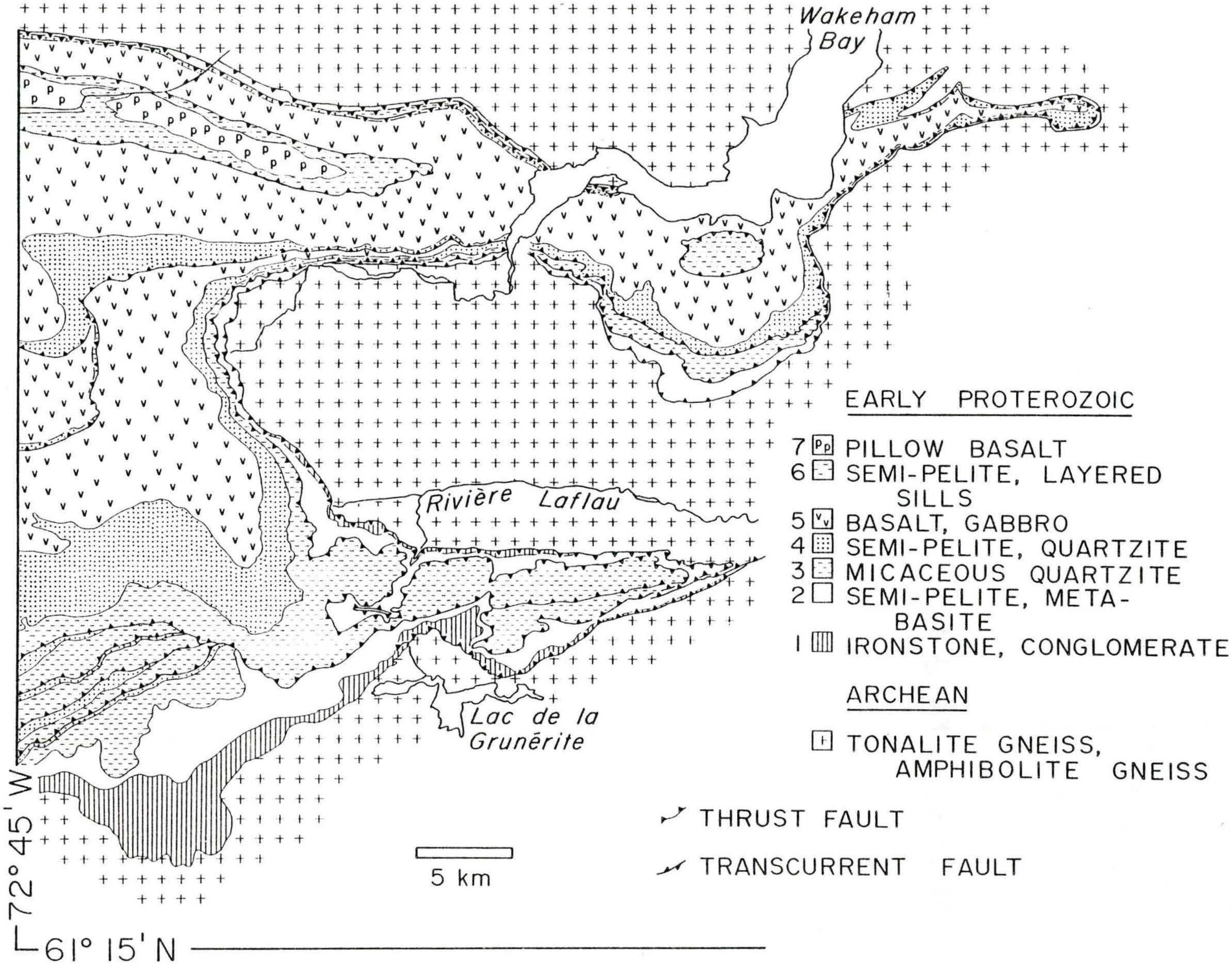
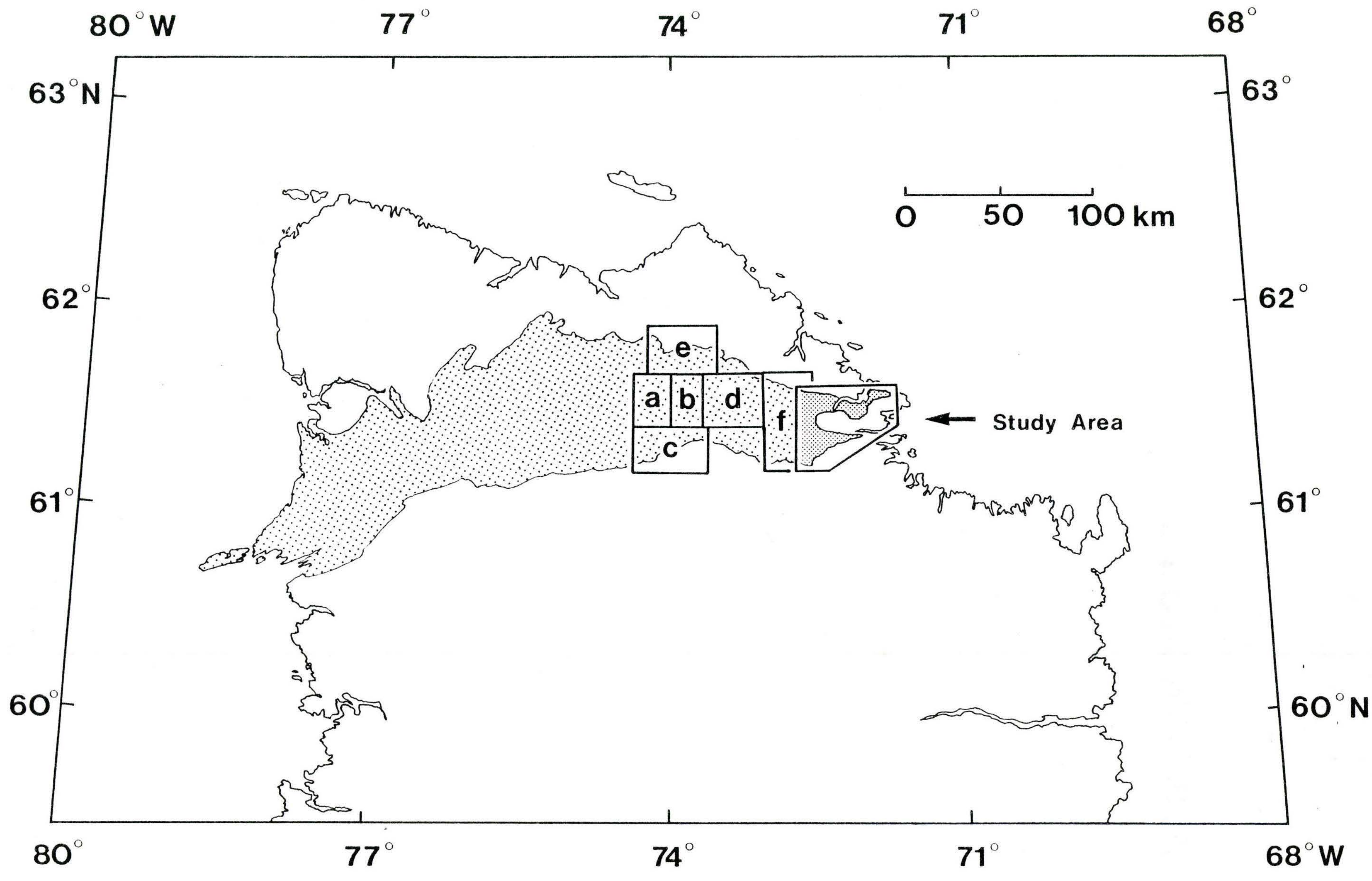

Wakeham Bay
Rivière Laflau
Lac de la Grunérite
EARLY PROTEROZOIC
7 PILLOW BASALT
6 SEMI-PELITE, LAYERED SILLS
5 BASALT, GABBRO
4 SEMI-PELITE, QUARTZITE
3 MICACEOUS QUARTZITE
2 SEMI-PELITE, META-BASITE
1 IRONSTONE, CONGLOMERATE
ARCHEAN
TONALITE GNEISS, AMPHIBOLITE GNEISS
THRUST FAULT
TRANSCURRENT FAULT
5 km
72° 45' W
61° 15' N

Figure 1-4: **Map Showing Location of Previous Work**

Entire Belt - light stipple - Bergeron, 1957; 1:1,013,760
Area a - Beall, 1959; 1:63,360
Area b - De Montigny, 1959; 1:63,360
Area c - Beall, 1960; 1:63,360
Area d - Gold, 1962; 1:63,360
Area e - Gelinas, 1962; 1:63,360
Area f - Schimann, 1972; 1:63,360
Dark Stipple- Study Area, St-Onge et al, 1986


80° W
77°
74°
71°
68°
63° N
62°
61°
60°
63°
62°
61°
60° N
80°
77°
74°
71°
68° W
0
50
100 km
a
b
c
d
e
f
Study Area

Taylor (1982). A regional metamorphic study covering the area mapped by Taylor, based on samples collected by Taylor, was published by Westra (1978).

In 1976 and 1977, a detailed study of the metamorphism in the Wakeham Bay area was done by Schimann (1978). A detailed stratigraphic, structural and petrological transect of the Belt at about 74° 30′ W was carried out in 1981 (Hynes and Francis, 1982; Francis et al, 1983).

The Quebec Ministere de l'Energie et des Resources is currently operating an ongoing, 1:50,000 scale mapping project west of 74° 30′W (Giovenazzo, 1985; Tremblay, 1985).

CHAPTER 2: GENERAL GEOLOGY

2.1 Regional Setting

The Cape Smith Belt is a synclinal structure, composed of well exposed, low- to medium grade metamorphosed sedimentary and volcanic rocks. The metasediments are dominantly terruginous clastics, ranging from orthoquartzites, micaceous quartzites and semi-pelites to pelites. Also present are impure carbonates, mafic volcaniclastics (meta-tuffs), extensive iron formations and a basal conglomerate. The metavolcanics vary from komatiitic and tholeiitic basalts to rare intermediate and rhyolitic flows. Gabbroic, peridotitic and layered gabbroic-peridotitic sills are also present.

Granodioritic to tonalitic rocks of Archean age surround the Belt, which was thought to represent the boundary of the Churchill (to the north) and Superior (to the south) structural provinces (Stockwell et al, 1976). However, field evidence from this study, in part, supports the recent model which suggests that the belt is, in fact, a klippe, allochthonously lying completely within the Superior province. (Hoffman, 1985). The basement gneiss complex is continuously exposed around the eastern end of the belt, with no evidence of any major features that would suggest the suturing of two gneissic blocks in that area. The contact between the Belt and underlying basement is highly strained, developing a shear foliation in the basement parallel to the dominant schistosity within the Belt.

This basal decollement (zone of detachment) forms part of a basal shear zone that decreases in thickness from north to south. On the northern margin, the zone is 240m thick, including 2-3m of reworked basement gneiss, while on the southern margin, east of Lac de la Grunerite, only 1-2cm of basement and 2-3m of cover are involved. (St-Onge et al, 1986). West of Lac de la Grunerite, the basal decollement "steps up" into the cover rocks, thus the basement-cover contact becomes unconformable.

Three discrete phases of deformation are identified within the belt. The first event (D_1) creates thrust faults and tight folds of bedding (F_1) that are characteristically isoclinal. A strong, synmetamorphic, bedding parallel, fold axial plane parallel schistosity (S_1) was developed, featuring a lineation (L_1) that is generally oriented steeply down dip, orthogonal to F_1 fold axes.

The second event (D_2) produced meso- to microscopic folds of S_1 that are chevron to rounded in style. Both D_1 and D_2 fold axes are oriented roughly east-west (Belt parallel).

The D_3 event is recognized primarily as macroscopic, km-scale wavelength cross-folds re-orienting D_2 fold axes. Oriented roughly north-south, these folds are thought to be responsible for the oblique crustal section exposed at the eastern end of the Belt.

2.2 **Table of Formations, Eastern Cape Smith Belt**
(tabulated from St-Onge et al, 1986).

TABLE OF FORMATIONS, Eastern Cape Smith Belt

Eon	Western Cape Smith Belt Correlative	Formation and thickness (m)		Lithology
Early Proterozoic	Chukotat Group	**Formation 7:** (min. 300)	Pillow Basalts	-pillowed basaltic lavas -minor gabbro and peridotite sills
		Formation 6: (400-500)	Layered Sills	-gabbro, peridotite, and layered gabbro-peridotite sills
	Upper Povungnituk Group		Semi-Pelites	-semi-pelites, micaceous quartzite, orthoquartzite
		Formation 5: (2100-4100)	Basalts	-locally pillowed lavas, massive flows, sills -rare semi-pelite
		Formation 4: (240-2200)	Orthoquartzites and Semi-Pelites	-white orthoquartzite -semi-pelite -metabasites, mafic sediments
	Lower Povungnituk Group	**Formation 3:** (70-270)	Miceaceous Quartzite	-massive, homogeneous micaceous quartzite
		Formation 2: (60-340)	Semi-Pelites and Metabasites	-semi-pelite, micaceous quartzite, pelite -metabasites, mafic sediment -minor dolomite and calcsilicate schist -minor hornblende-garnet-biotite schist -peridotite sill
		Formation 1: (50-500)	Ironstones and Basal Conglomerate	-interbedded magnetite-grunerite schist -ferruginous sandstone -grunerite-carbonate schist -garnet-biotite-hornblende schist -olgomictic pebble conglomerate
Archean	Basement Gneiss Complex			-tonalitic/granodioritic gneiss

2.3 Description of Basal Sedimentary Formations

The lowest four Formations in the Belt contain the bulk of the metasediments. As this thesis study focuses on these four Formations, this discussions omits Formations 5, 6, and 7. St-Onge (et al, 1986) describes these upper three Formations in detail. This paper is also the source of much of the following discussion.

Stratigraphic sections of the lowest four Formations were measured at five locations in the study area. The stratigraphic cross-section produced by the correlation of these measured columns roughly transects the Belt, and is shown in Figure X (pocket, back cover).

Formation 1: Basal Conglomerate and Ironstones

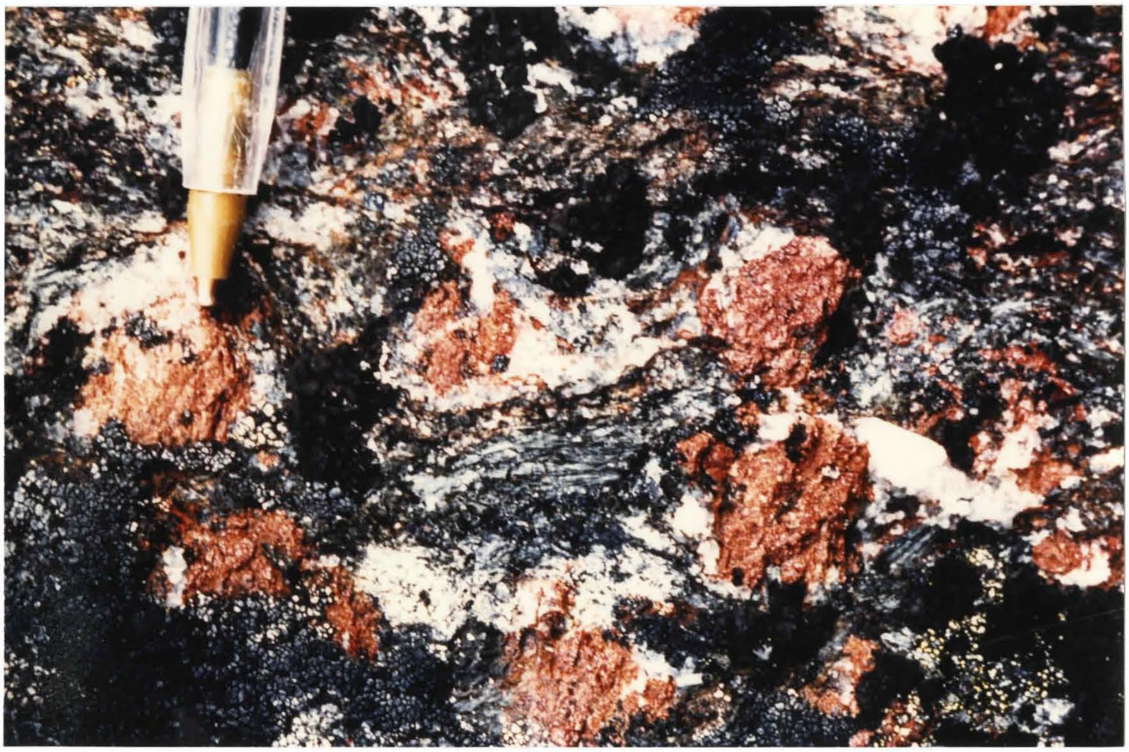
This formation is best developed and almost exclusively seen on the south margin of the Belt. A thin (less than 1m) oligomictic metaconglomerate is occasionally observed lying unconformably on steeply dipping tonalitic gneisses (Plate 2-1).

A wide variety of ironstone units was recognized. Lithologies varied from ferruginous sandstones, magnetite-grunerite-garnet schist, carbonate-rich grunerite schist, to coarse grained garnet-biotite-hornblende schist (Plate 2-2a,b). While voluminous and mineralogically distinct, these lithologies are discontinuous

Plate 2-1: Oligomictic basal metaconglomerate, Formation 1, at Lac de la Grunerite. Pen is 15cm in length.

Plate 2-2a: Magnetite-grunerite-garnet schist, Formation 1, at Lac de la Grunerite. Pen is 15 cm in length.

Plate 2-2b: Garnet-biotite-hornblende schist, Formation 1, at Lac de la Grunerite. Pen is 8mm in diameter.



laterally, hence are not sub-divided on the present map (Fig. 1-3). Some sub-division is shown in the Lac de la Grunerite stratigraphic section (Fig. X).

While these ironstones contain a significant amount of iron, most of it is bound in silicates such as grunerite, garnet and hornblende. Magnetite is present either as disseminated grains, or thin laterally discontinuous horizons. It is not concentrated in significant amounts. Thus, these ironstones are probably uneconomic at present.

<u>Formation 2: Semi-Pelites and Metabasites</u>

Formation 2 is dominated by semi-pelite and micaceous quartzite, with minor horizons of calcsilicate, dolomite and pelite. Interlayered with these are mafic sediments, as well as mafic and ultramafic flows/sills. Discontinuous horizons of hornblende-garnet-biotite schist are present (see Plates 2-3a-d).

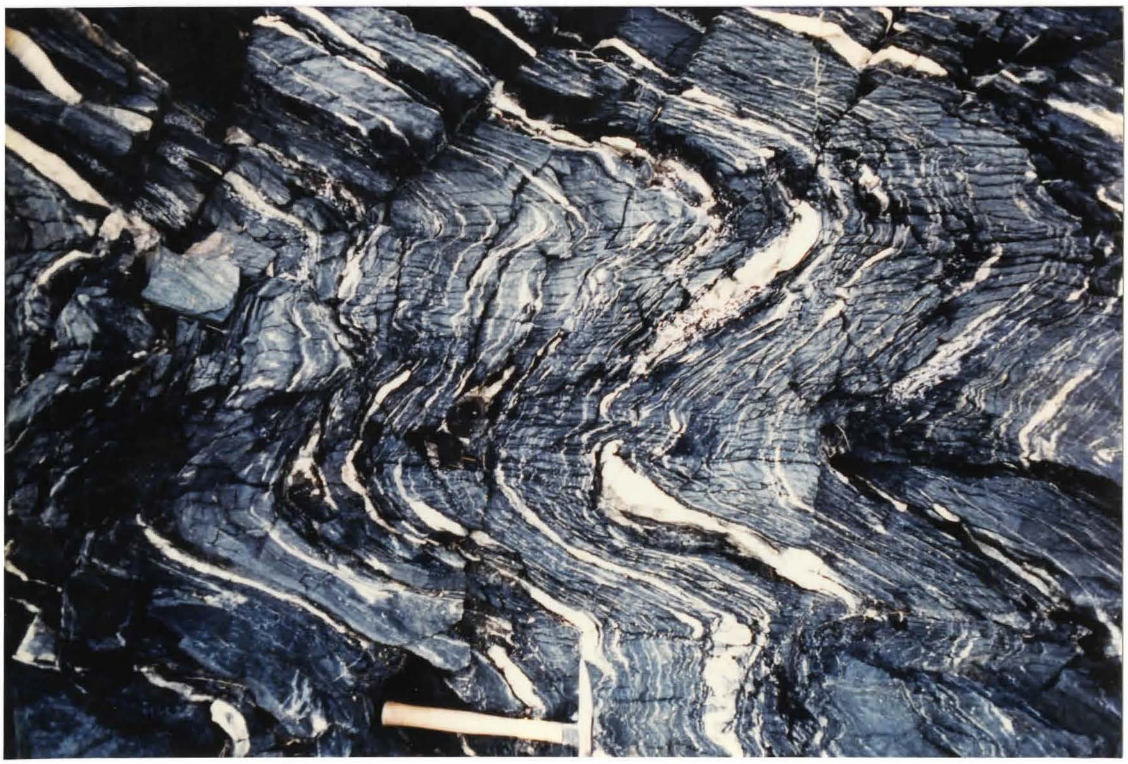
Fine compositional layering (sub millimeter to centimeter scale) in the siliciclastic units is noted both in the field and in thin section, and is thought to represent original bedding. Metamorphic growth of mica has defined a bedding parallel schistosity in these lithologies.

Plate 2-3a: Semi-pelite, composed of quartz + biotite + garnet. Formation 2, south of Wakeham Bay. Pen is 8mm in diameter.

Plate 2-3b: Pelite, with garnet + staurolite + kyanite. Formation 2, south of Wakeham Bay. Pen is 8mm in diameter.

Plate 2-3c: Calcsilicate showing large rosettes of tremolite. Formation 2, Cripple Creek section. Pen is 8mm in diameter.

Plate 2-3d: Mafic Sediment, with quartz veins. Formation 2, Wakeham Bay, west shore. Hammer is 40 cm in length.



The dolomite is regularly associated with the calcsilicates. Together they average 2-3m in thickness, rarely exceding 5m. This lithology is readily deformed, and occasionally contains fine layers of pelitic impuritites. This lithology has very poor lateral continuity, often traceable for no more than tens of meters.

A boudinaged, layered gabbro-peridotite sill, varying from 10-20 meters thick, was observed lying stratigraphically immediately below the carbonate horizon.

The mafic sediments are composed of quartz and biotite, with actinolite and hornblende. This is thought to indicate a volcanogenic component introduced into background clastic sediment.

Formation 3: Micaceous Quartzite

This formation is almost exclusively micaceous quartzite, interupted rarely by semi-pelite or mafic sediment horizons. The micaceous quartzite is a resistant, grey weathering rock, frequently with a silvery sheen on the S_1 surface due to the presence of muscovite. A characteristic feature is the pressure of discontinuous quartz veins/segregations, a few millimeters to several centimeters in width, 5-10cm in length, spaced roughly 5-25cm apart. These quartz segregations are evenly distributed throughout the unit, and lie parallel to S_1 (see Plate 2-4).

Plate 2-4: Micaceous quartzite, composed of quartz + biotite + muscovite, showing quartz segregations. Formation 3, Cripple Creek section. Pen diameter is 8mm.

Plate 2-5: Orthoquartzite, showing bedding on 5-10cm scale, Formation 4, southwest shore of Wakeham Bay. Pen is 15cm in length.

The minerology is typically 60% quartz, 10% feldspar (usually untwinned plagioclase), 15% biotite and 15% muscovite. Calcite, opaques and garnet may be present in trace amounts.

Formation 4: Orthoquartzites and Semi-pelite

Clean, white orthoquartzite and interbedded biotite-rich semi-pelite are the dominant clastic lithologies in this formation (see Plate 2-5). Numerous coarse grained gabbroic sills inflate the measured thickness of this formation.

At higher structural levels, graded bedding and original clasts can be seen in the orthoquartzites. The semi-pelites are bedded on a centimeter scale, showing parallel laminations at the tops of beds.

Meter thick rhyolitic lenses, showing little lateral continuity, are occasionally found in this formation.

CHAPTER 3: METAMORPHISM

The basal rocks of the Cape Smith Belt contain a wide variety of lithologies, some more useful than others in terms of systematic determination of regional metamorphic conditions. For maximum systematic utility, there should be good outcrop control throughout the study area, the lithology should have roughly constant bulk composition, and be understood in terms of experimental petrology. In light of these considerations, micaceous quartzites, semi-pelites and metabasites here present excellent opportunities for systematic study. Pelites, while only poorly exposed at a limited number of locations, are so well understood experimentally that they offer the best opportunity to obtain concise estimates of pressure and temperature, based on mineral assemblages.

Each of the above four lithologies can be sub-divided into relative metamorphic grade zones, defined by the first occurance of a specific mineral or mineral pair. The lowest grade zone is designated Zone A, higher grade zones are designated B and C respectively.

The dolomite and calcsilicate lithologies are not sub-divided into grade zones, due to their limited exposure and poor areal distribution. Based on significantly different mineral assemblages in adjacent outcrops, rocks mapped as "mafic sediment" and ironstone are each judged to encompass too much compositional variation to be here included in the systematic grade zone study.

3.1 Micaceous Quartzite

At the lowest grade observed in the study area, the micaceous quartzite assemblage consists of quartz + plagioclase + biotite + muscovite +/- opaques. As plagioclase is generally untwinned, representative thin sections were stained to aid in feldspar identification, using the method outlined by Houghton (1980). The micas are aligned in a well-developed schistose fabric, denoted S_1. This assemblage represents Zone A, defined by the appearance of biotite + muscovite. Zone B is defined by the addition of garnet to the above assemblage (see Fig. 3-1).

A proposed reaction for the transition to Zone A is purely speculative, as rocks below this grade are not observed in the study area. One reasonble reaction, proposed by several authors (Brown, 1971, 1975; Mather, 1970; Winkler, 1979) is:

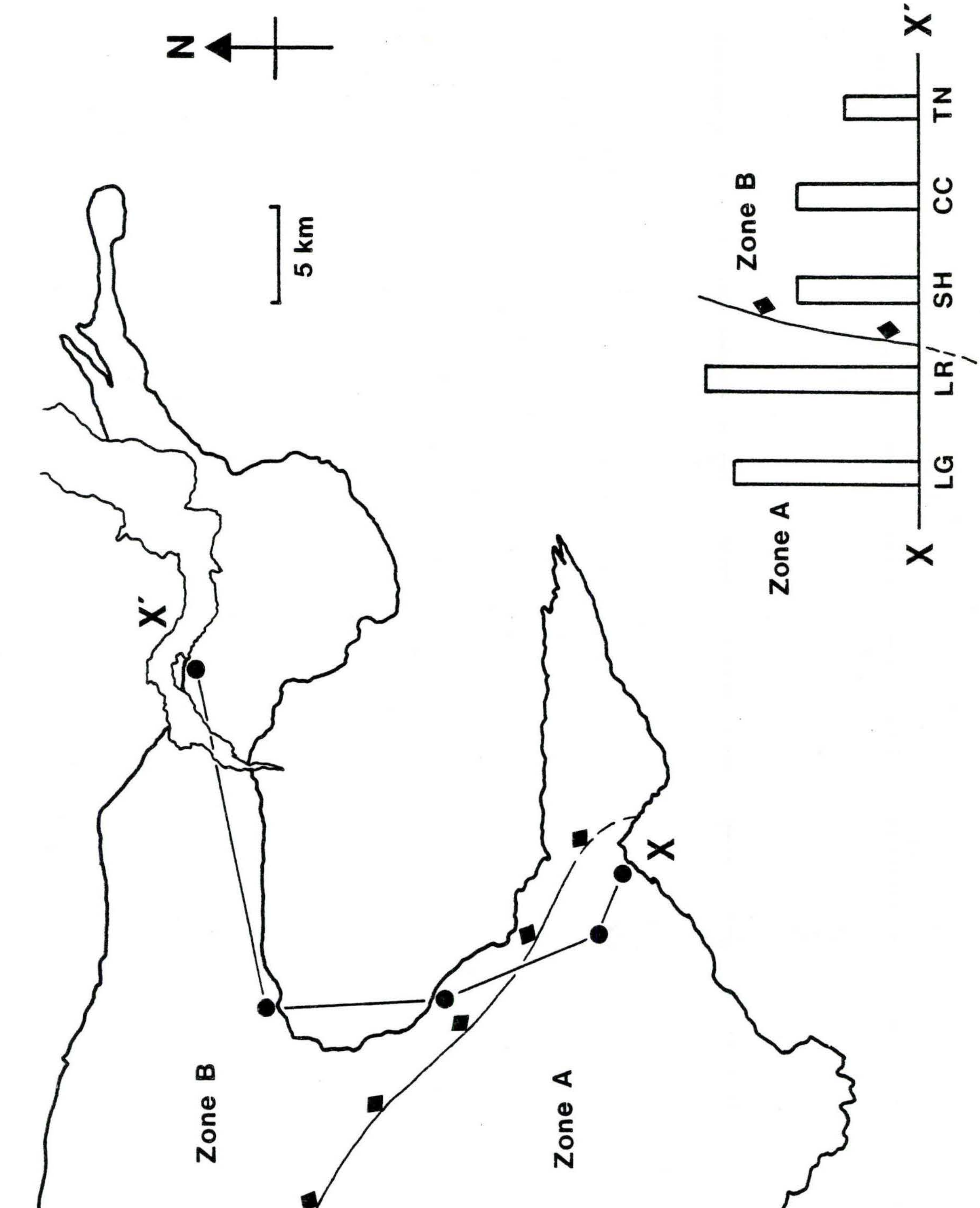
$$\text{microcline + chlorite} = \underbrace{\text{biotite + muscovite}}_{\text{Zone A}}$$

It is suggested that this reaction occurs at 400°C at 2 kbar (Brown, 1971), thus suggesting an estimate of minimum pressure and temperature conditions for the study area. Microcline is

Figure 3-1: **Micaceous Quartzite Isograd Map,** Eastern Cape Smith Belt.

Approximate traces of "garnet-in" isograd, map and cross-sectional views. Line of section X-X′ corresponds to Figure X (correlation of measured stratigraphic sections). Map area corresponds to that shown in Figure 1-3. Zones explained in text.

"garnet-in" isograd


N
5 km
X′
X
Zone B
Zone A
Zone B
Zone A
X
X′
LG
LR
SH
CC
TN

occasionally present in Zone A rocks, but relict chlorite is never observed. When chlorite is observed in these rocks, it is as a clearly late alteration product, after biotite or garnet.

The transition from Zone A to Zone B is defined by the appearance of garnet, assumed to be almandine-rich. A possible reaction is:

$$\underset{\text{Zone A}}{\text{biotite + muscovite}} = \underset{\text{Zone B}}{\text{almandine + K feldspar}} + H_2O$$

(D.M. Shaw, p.comm. 1986)

This reaction is not observed to go to completion, as the Zone B assemblage still contains both biotite + muscovite, in addition to quartz + plagioclase +/- opaques. Similar almandine forming reactions have an approximate range of T and P of 500-600^{o}C and 4-5 kbar (Winkler, 1979).

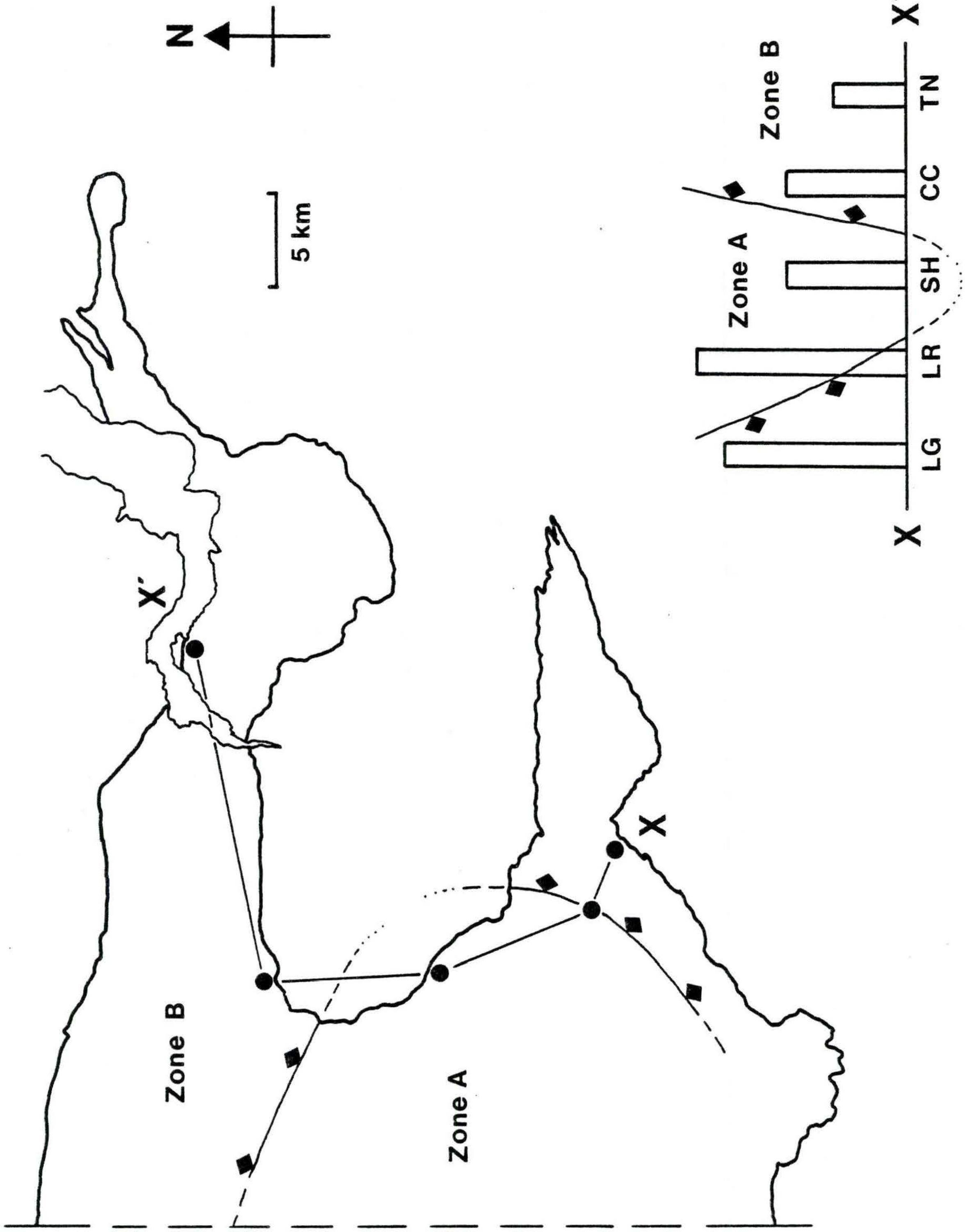
3.2 Semi-Pelite

This lithology is similar to the micaceous quartzite, but contains less quartz and has no muscovite. The Zone A assemblage consists of quartz + plagioclase + biotite +/- opaques. Zone B is defined by the appearance of garnet in the above assemblage (see Fig. 3-2).

Figure 3-2: **Semi-Pelite Isograd Map,** Eastern Cape Smith Belt.

Approximate traces of "garnet-in" isograd, map and cross-sectional views. Line of section X-X′ corresponds to Figure X (correlation of measured stratigraphic sections). Map area corresponds to that shown in Figure 1-3. Zones explained in text.

"garnet-in" isograd


N
5 km
X′
X
Zone B
Zone A
Zone A
Zone B
LG
LR
SH
CC
TN
X
X′

The transition from Zone A to Zone B, by addition of almandine-rich garnet, may be explained by the following reaction, as suggested by St-Onge (p.comm. 1986):

$$\text{quartz} + \text{biotite}_1 = \text{almandine} + \text{biotite}_2$$

This reaction is thought to occur at metamorphic conditions similar to those of the almandine forming reaction suggested for the micaceous quartzites in Section 3.1, as the two isograds are nearly coincident in physical space.

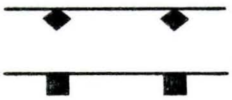
3.3 Pelite

Outcrop exposure of this lithology is poor, but the few observed locations have wide geographical distribution. Three grade zones have been established for this lithology. The Zone A assemblage is defined as quartz + plagioclase + biotite + muscovite, and is observed in the vicinity of Lac de la Grunerite. Zone B is defined by the appearance of garnet, observed in the vicinity of the central basement cored anticline (see Fig. 1-3). The Zone C assemblage contains staurolite + kyanite, in addition to garnet + quartz + plagioclase + biotite + muscovite, and is observed at several locations around Wakeham Bay (see Fig 3-3).

The transition from Zone A to Zone B can be described by the following reaction:

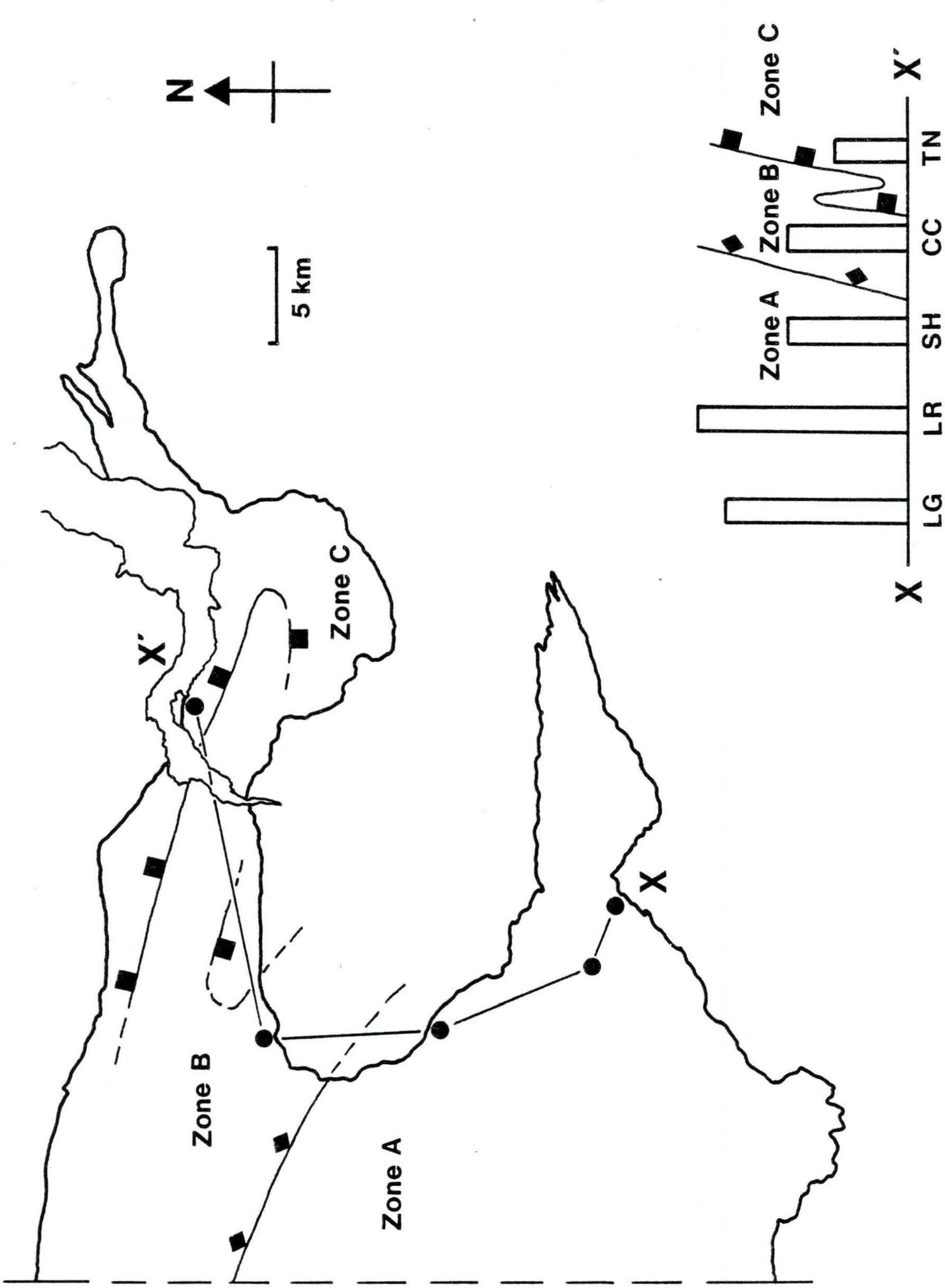
Figure 3-3: **Pelite Isograds Map,** Eastern Cape Smith Belt.

Approximate traces of "garnet-in" and "kyanite + staurolite-in" isograds, map and cross-sectional views. Line of section X-X′ corresponds to Figure X (correlation of measured stratigraphic sections). Map area corresponds to that shown in Figure 1-3. Zones explained in text.



"garnet-in" isograd

"kyanite + staurolite-in" isograd


N
5 km
Zone A
Zone B
Zone C
X
X′
LG
LR
SH
CC
TN

biotite + muscovite = almandine + K feldspar + H_2O

Zone A Zone B

This reaction occurs at a deeper structural depth than in the micaceous quartzite, (compare Figs. 3-1 and 3-3), thus presumably at higher grade. This may be due to production of a garnet of slightly different composition than those found in the micaceous quartzites.

The first appearance of staurolite occurs within Zone B. While the first occurence is not spatially well confined, it is a prograde feature. Hence, staurolite is adjacent to Zone C, rather than Zone A.

St-Onge (1984) suggests the following staurolite forming reaction:

garnet + muscovite + chlorite = staurolite + biotite + quartz + H_2O

Zone C is characterized by the addition of kyanite. This transition may be explained by the following reaction (St-Onge, 1984):

staurolite + muscovite + quartz = kyanite + garnet + biotite + vapour

Pelite samples observed petrographically contained the above six minerals, plus plagioclase. The co-existence of all the reactants and products indicates a univariant assemblage. This reaction is constrained experimentally to a minimum of 5.5 kbar and 575°C (St-Onge, 1984, Fig. 1-4).

3.4 Metabasite

Metabasic flows and sills are generally well exposed and widely distributed. Zone A is characterized by the presence of actinolite + hornblende + plagioclase + quartz +/- epidote, chlorite,and calcite. The western limit of this zone has not been determined. Zone B is defined by the disappearance of actinolite from the above assemblage. Zone C is defined by the appearance of garnet (see Fig. 3-4).

Raase (et al, 1986) suggest that in general, metabasic rocks do not show dramatic changes in their mineral assemblages over a wide range of P-T conditions, as do the metapelites. Rather, metabasites show more continuous modal variations and chemical changes, especially in amphiboles and plagioclase.

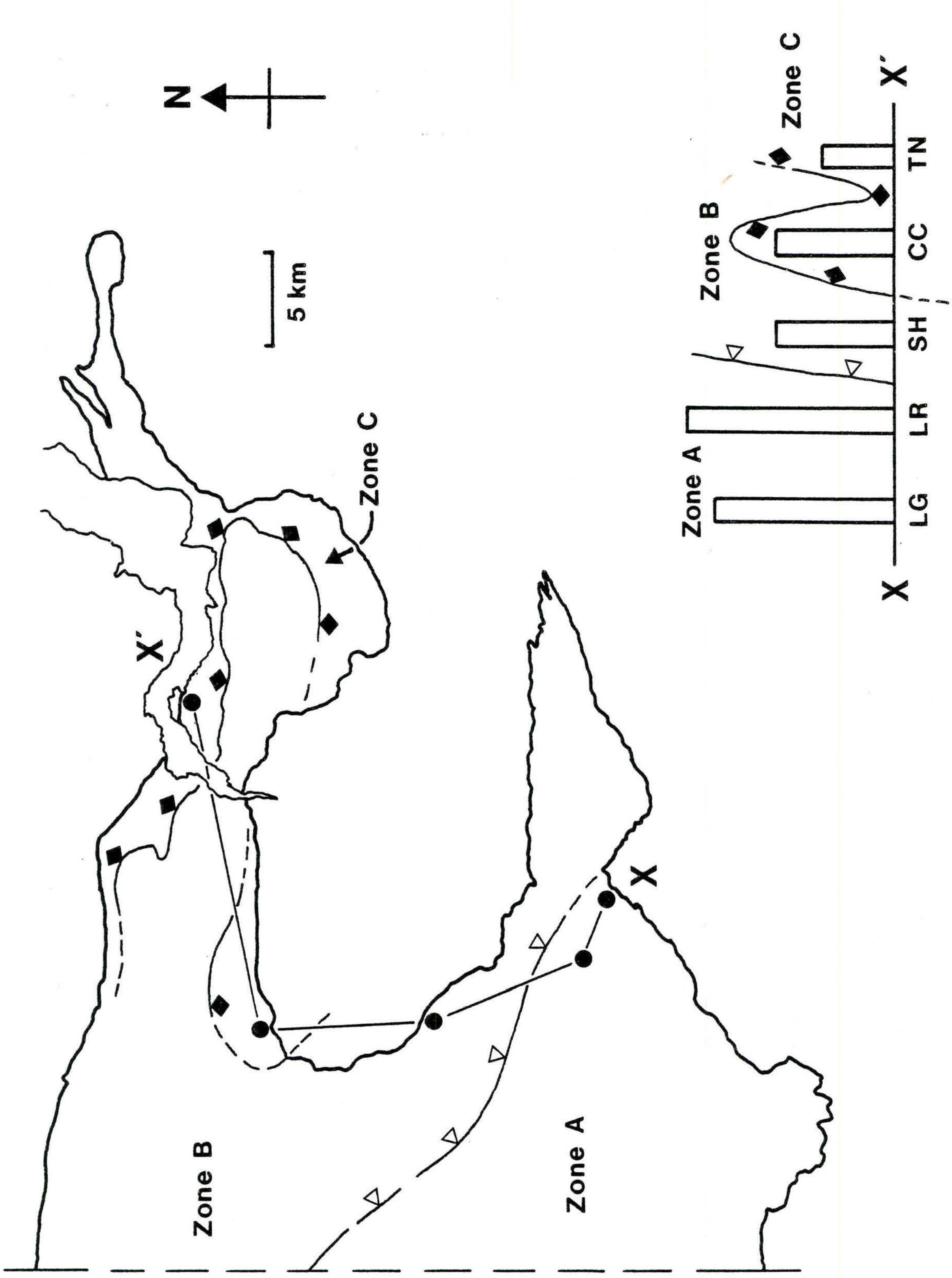
In the absence of detailed chemical analyses, statements regarding mineral compositional variation with pressure and temperature are not possible. It is assumed for the purposes of this study that the bulk composition of the metabasites is roughly constant. Thus, the various mineral assemblages are due to changes

Figure 3-4: **Metabasite Isograds Map,** Eastern Cape Smith Belt.

Approximate traces of "actinolite-out" and "garnet-in" isograds, map and cross-sectional views. Line of section X-X′ corresponds to Figure X (correlation of measured stratigraphic sections). Map area corresponds to that shown in Figure 1-3. Zones explained in text.

"actinolite-out" isograd

"garnet-in" isograd


N
5 km
Zone C
X'
X
Zone B
Zone A
Zone A
Zone B
Zone C
X
X'
LG
LR
SH
CC
TN

in metamorphic conditions, rather than variations in composition. This is supported by the general spatial conformity of the metabasite isograds to the isograds mapped in other lithologies.

Assemblages of Zone A are typical of those found in the greenschist facies, at the beginning of the transition from actinolite to hornblende (Raase et al, 1986).

The reaction:

actinolite + epidote + calcite = hornblende + plagioclase

(Raase et al, 1986)

is thought to be occurring within Zone A.

The transition to Zone B is marked by the completion of this reaction, limited by actinolite availability.

The formation of garnet (transition to Zone C) is speculative in the absence of chemical analyses. It is most likely related to partial consumption/composition change of hornblende and plagioclase (Raase et al, 1986).

A summary and comparison of the locations of the various metamorphic indicator minerals used in sections 3.1 to 3.4 is shown in Figure 3-5.

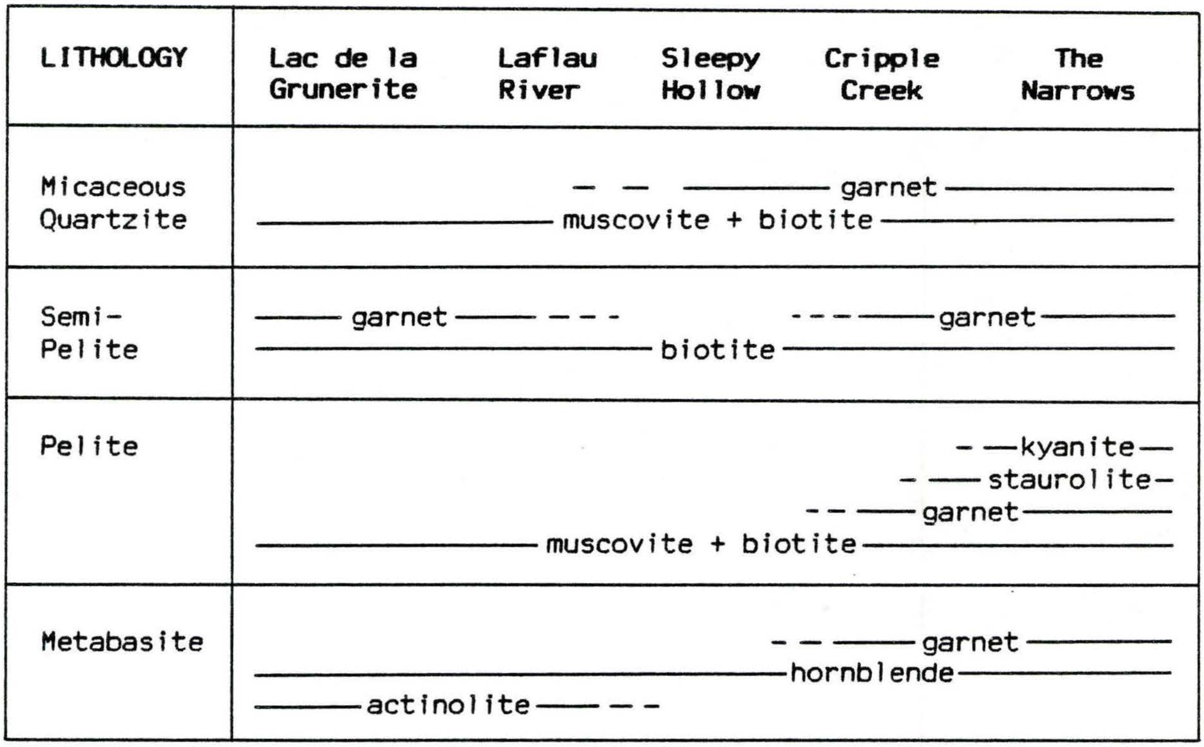


Figure 3-5: A comparison of the locations of the various metamorphic minerals used to determine the locations of the isograds shown in Figures 3-1, -2, -3, and -4.

3.5 Mafic Sediment

As previously mentioned, this lithology is interpreted as having significant variation in composition, often between adjacent outcrops. This is thought to reflect the origin of the lithology, as a combination of basic volcanic material and background terruginous sediments. For example, rocks mapped in the field as mafic sediment, in outcrops only a few hundred metres apart, have assemblages as varied as biotite + muscovite + quartz, and quartz + plagioclase + hornblende. In comparison, adjacent metabasic flows are hornblende + plagioclase + quartz + opaques. In these cases, mafic sediment assemblages may reflect metabasite assemblages quite closely. In these cases, analogous metamorphic reactions might apply.

3.6 Dolomite, Calcsilicate

Dolomite and calcsilicate lithologies are not extensively developed in the Belt. Laterally discontinuous, 2-3m thick horizons are noted in numerous locations. The majority of these, however, are very clean, pure, dolomite. Thin sections were stained to distinguish between calcite and dolomite, according to the method outlined by Friedman (1959).

In the Cripple Creek section (Fig. X), carbonate rocks are present approximately 150m from basement, with the assemblage calcite

+ dolomite + tremolite + phlogopite + biotite + talc. This assemblage may be explained by the following reaction:

$$\text{talc} + \text{calcite} = \text{tremolite} + \text{dolomite} + CO_2 + H_2O$$

(Winkler, 1979, Fig 9.5)

Winkler suggests that at a fluid pressure of 5 kbar, this reaction occurs at about 530°C for a mole fraction of CO_2 between 0.2 and 0.6. This estimate of pressure and temperature is similar to, but slightly lower than, that suggested by the pelitic assemblage found several kilometers to the northeast.

3.7 Ironstone

Considerable variation in assemblage in adjacent outcrops of this lithology suggests significant local variation in composition.

At Lac de la Grunerite, the dominant assemblages are:

a) grunerite + carbonate

b) grunerite + carbonate + garnet

c) grunerite + garnet + magnetite + biotite + quartz

d) grunerite + quartz + magnetite + carbonate + feldspar + garnet + stilpnomelane(?)/biotite.

Certain of these minerals are stable from diagenesis to the pelitic kyanite-staurolite zone (Fig. 3-3), i.e. calcite, dolomite/ankerite, siderite, hematite, magnetite and quartz. Others, such as grunerite, garnet, hornblende and stilpnomelane have a more restricted pressure - temperature range, thus may be used in a semi-quantitative sense, to estimate metamorphic conditions (Klein, 1978).

Grunerite [assemblage (a)] is stable from the Biotite Zone into the Staurolite + Kyanite Zone ("Zone" terminology from Klein, 1978, Fig. 4). However, with the addition of garnet to the assemblage, [assemblages (b) and (c)], the stability field of grunerite is restricted to the Garnet Zone/lower Staurolite + Kyanite Zone. The addition of stilpnomelane to the assemblage (d) pins the stability field to the Garnet Zone.

Slightly south of the Sleepy Hollow section (see Fig. X), the assemblage grunerite + carbonate + biotite is noted. According to Klein (1978), this assemblage is stable from the Biotite Zone up to the Staurolite + Kyanite Zone. In the Cripple Creek section (Fig. X), Formation 2, hornblende + garnet + quartz + biotite + carbonate is found. This assemblage is characteristic of the Garnet Zone.

As noted in section 3.1, the Biotite Zone corresponds to a minimum of roughly 400^{o}C at 2 kbar, and the Garnet Zone to a minimum of 500^{o}C at 4 kbar. The formation of grunerite from minnesotaite is thought to occur at 320^{o}C over the pressure ($P=P_{H2O}$) range of 2 to 6 kbar (Mel'nik, 1982). This gives minimum temperature and pressure estimates for the Lac de la Grunerite assemblages. Mel'nik (1982) also estimates the dehydration breakdown of grunerite to fayalite + quartz + H_2O at 660°C and P_{H_2O} of 4 kbar, suggesting maximum temperature and pressure conditions for the area.

CHAPTER 4 STRUCTURE

4.1 Initial Phase of Deformation

Structural field data is shown in detail on nine 1:50 000 scale field compilation sheets, as mapped by St-Onge, Lucas, Scott, and Begin during the 1985 field season. The reader is refered to these sheets (Maps A-I, pocket, back cover) for detailed documentation of the structural features discussed in this chapter.

The first deformational event (D_1) involves development of and transport along low angle, bedding parallel thrust faults. Tight, isoclinal folds of bedding (F_1) are formed during this event. A strong, synmetamorphic, bedding parallel schistosity (S_1) is formed, decreasing in intensity from north to south across the Belt, as well as in the up-section direction. This schistosity is interpreted as being axial planar to the tight folds of bedding (Plate 4-1a, b). A steep "down dip" mineral lineation was noted on S_1.

Detailed subdivision of lithologies at the eastern end of the Belt enabled the recognition of inter-formational stratigraphic repetitions. These repetitions, consistently placing older- on top of younger stratigraphic units, are thought to be due to the presence of thrust faults. This is clearly illustrated in the south-west corner of Figure 1-3, where formations 3 and 4 are repeated four times (see also Map Sheets C and D).

Plate 4-1a: Photomicrograph of thin section showing growth of micas (S_1) parallel to the axial plane of a fold of bedding (S_0), semi-pelite lithology. Field of view is 3cm by 2cm, plane light.

Plate 4-1b: Semi-pelite outcrop showing fold of bedding (S_0), and development of an axial planar cleavage. Pen is 1cm wide.

Four thrust faults are shown in Figure X. Their geometry is characterized by dominantly isostratigraphic flats, with occasional ramps cutting up-section towards the south. This flat-ramp-flat geometry is thought to be characteristic of thrusted continental margin sediments (Boyer and Elliot, 1982), and indicates a south directed transport direction. Where noted, L_1 is thought to represent a metamorphic manifestation of thrusting -- a transport direction-parallel stretching lineation. Piggy-back stacking (Butler, 1982), interpreted from field relationships between thrust faults and their contained thrust slices suggest normal, foreland-directed (southward) development and transport (Boyer and Elliot, 1982).

D_1 strain is thought to decrease stratigraphically up-section, as well as isostratigraphically north to south. A layered, peridotite-gabbro sill lying 50-80m stratigraphically above basement in the northern half of the map (see Map E) is seen boudinaged at map scale. Individual boudins range from 50-500m in length, and are generally <20m thick, and are 50-700m apart. Margins of individual boudins are sheared, and conformably wrapped by S_1. Minimum extension is estimated at 100% (St-Onge et al, 1986). A similar layered sill in Formation 6, lying physically and stratigraphically above the boudinaged sill location, (Map F), provides a marked contrast in strain. At this elevated level, the sill shows no sign of penetrative D_1 deformation. Rather, it is gently folded into a broad F_2 syncline.

A thrust surface roughly 10m stratigraphically above the level of the boudinaged sill is folded, conformable to the individual boudins. This relationship would suggest the thrust surface predates the high strain associated with the boudinaging event. Both the thrust and the boudinaging are interpreted as D_1 events.

Development of penetrative S_1 fabric is less advanced in all lithologies with increasing stratigraphic height. While present at lower structural levels, S_1 is absent in thick quartzites and metabasites at higher levels. However, these units are in formations above those being studied in detail here, and will not be discussed further.

On the northern margin of the Belt, the basal shear zone (zone of high shear strain) has a total width in excess of 240m, including 3m of reworked basement gneiss (see Plate 4-2, also Map H). This width systematically decreases southward, resulting in a zone only several meters wide, involving 1-2cm of reworked basement. The zone is defined by those rocks showing highest D_1 strain, based on several qualitative strain indicators. Included were strain-induced grain size reduction, presence of a stretching lineation, intensity of S_1 and its parallelism to bedding, and boudinaging of competent bodies such as peridotite sills and quartz veins (St-Onge et al, 1986).

At Lac de la Grunerite, the lowest thrust (basal decollement) steps up, out of the basement and into the cover (see Fig. 1-3,

Plate 4-2: Reworked basement gneiss (left of hammer) transposed into parallelism with cover rocks (right of hammer). Northeast of Wakeham Bay Village. Hammer is 40cm in length.

Plate 4-3: Penetrative, transverse stretching lineation (L_1) defined by quartz and garnet, on transposed S_1 surface. Reworked basement gneiss, west shore of Wakeham Bay. Pen is 15cm in length.

also Map B). South and west of this lake, cover rocks rest uncomformably on the basement, ubiquitously composed of alternating, centimeter-scale bands of granodioritic and tonalitic material. The gneissosity ranges from moderate- to steeply dipping, but is completely transposed into parallelism with the basement-cover contact and S_1 in the above-mentioned reworked zone, immediately below the contact. Within this transposed zone, distinct grain size reduction is noted, and is attributed to strain-induced progressive recrystallization (St-Onge et al, 1986). This notion is supported by petrographic observations. The reworked zone is characterized by the growth of garnet and S_1-parallel muscovite. A penetrative, transverse stretching lineation (L_1), defined by smears of quartz/feldspar and garnet, is commonly noted on S_1 surfaces (see Plate 4-3). This smearing lineation is oriented orthogonal to the general E-W axes of isoclinal folds of reworked gneissosity, suggesting this lineation is sub-parallel to the direction of D_1 transport.

4.2 D_2 Structures

The second deformational event folds the S_1 fabric, and gives the Belt its generally synclinal form. More specifically, the rocks at the eastern end of the Belt form two west-plunging synclines separated by a basement-cored anticline (see Maps A, B, F, G, H).

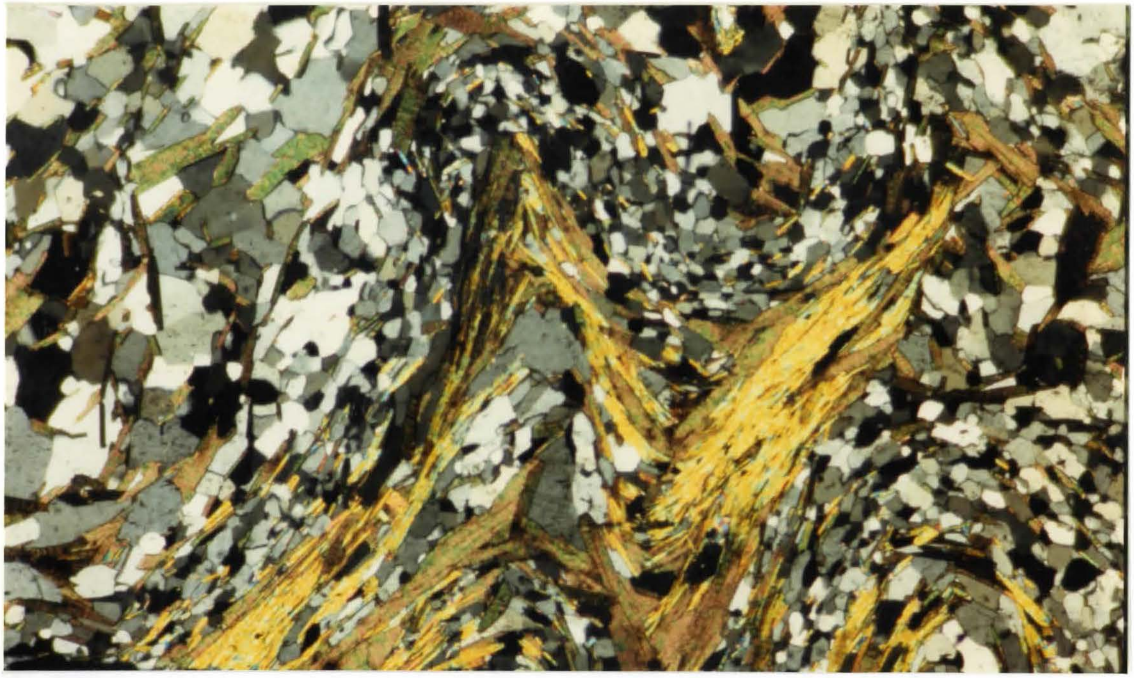
For purposes of correlation, D_2 folds observed in outcrop have been subdivided into two major groups (Hobbs et al, 1976). Angular- to chevron-type folds are best developed in micaceous lithologies such as mafic sediment, semi-pelite, mica-rich bands in micaceous quartzite, and occasionally in metabasites (Plate 4-4). Rounded-type folds are more ubiquitous (Plate 4-5). Interlimb angles of chevron folds are generally tight, rounded folds being tight to open. Further subdivision was made on the basis of fold amplitude. Approximate categories were established for sub cm-, cm-, 10cm-, m- and 10m amplitude folds. Chevron folds were noted in only the first two size divisions, while rounded folds were noted at all scales.

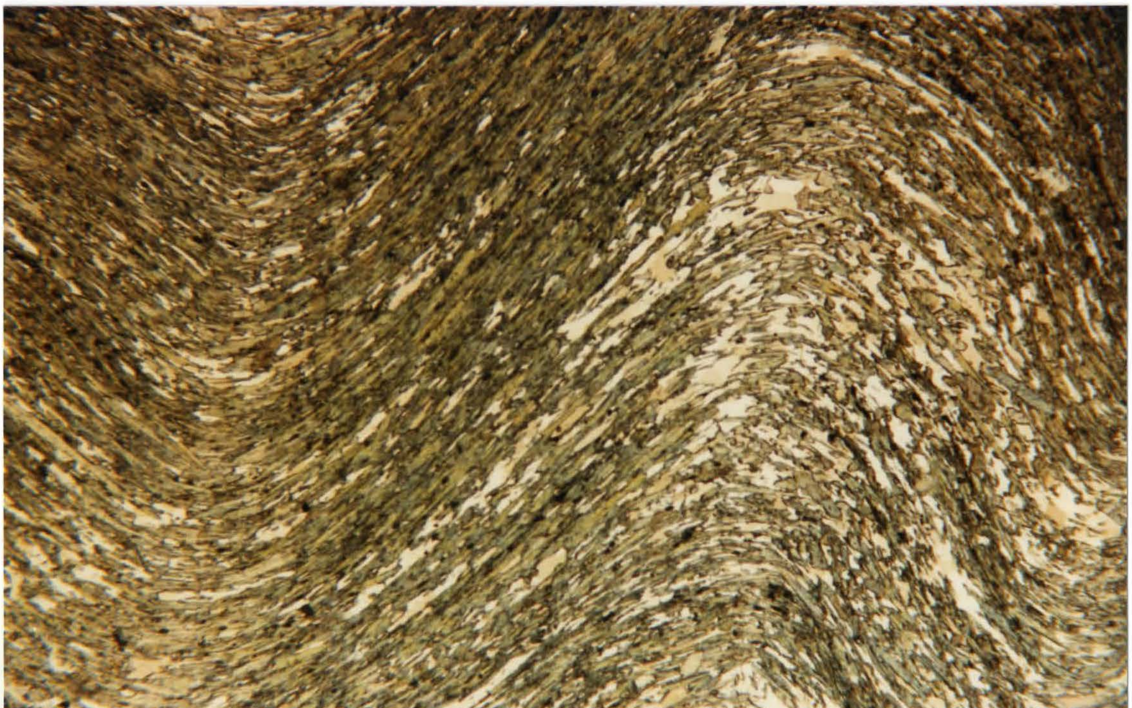
For a given lithology, intensity of folding decreases up stratigraphic section, as does the range of sizes and styles. In the semi-pelites at the base of the Laflau River section (unit 1, lower unit 2, see Fig. X), sub cm- and cm-scale chevron and sub cm- and 10cm scale open folds are found in abundance. In the upper half of the section, fewer total folds are present, chevron folds are absent, and rounded folds still range from sub cm- to 10cm scale in amplitude. In the uppermost part of the section, 10cm-scale rounded folds are rare, and sub cm- to cm-scale rounded folds are greatly reduced in number. Chevron folds are absent.

While this is generally the trend, some exceptions do exist. Occasionally, size and morphology remain roughly constant in a given lithology throughout an entire section, the only variation being an

Plate 4-4: Photomicrograph showing tight, angular chevron folds (D_2) of muscovite-rich layers (S_1) in micaceous quartzite. Field of view is 1.5mm x 2.5mm, crossed polars.

Plate 4-5: Broad, open, rounded folds (D_2) of biotite-hornblende defined schistosity (S_1) in a metabasite. Note enrichment of quartz in fold nose relative to fold limb. Field of view is 2mm by 3mm, plane light.





overall decrease in number of folds per unit area with increasing height (See Fig. X, Sleepy Hollow section, mafic sediment lithology).

The stratigraphic section measured at The Narrows (See Fig. X) is anomalous with respect to D_2 folds. Very few mesoscopic folds of S_1 are observed in outcrop. The first cm-scale rounded folds are noted in micaceous quartzite 160m above basement. Further folds of this variety are observed again, in orthoquartzite, approximately 190m above basement. Aside from these two, D_2 deformation is conspicuous in its absence. One possible explanation is that the outcrop face was oriented such that D_2 folds were only rarely seen in profile. Secondly, all D_1 thrusts are rooted in this section. The large amount of displacement on thrusts contained within this limited height of section may have caused extensive distributed shearing within the section. The high metamorphic grade of assemblages within this section (kyanite-staurolite-garnet in pelites) suggests that a rather ductile, distributed deformation may have occurred. This could possibly have resulted in work hardening of the rocks, inhibiting later, D_2 fold development (Johnson, 1980).

The Lac de la Grunerite section (see Fig. X) shows an inverted strain gradient-- lowest at the base, higher moving up-section. In the basal autocthonous section, S_1 is poorly developed and evidence of D_2 is non-existant. Above the basal thrust, S_1 gradually becomes the dominant fabric. Continuing upsection, folds of S_1 become more numerous. These folds are limited to mafic sediment and are generally sub-cm scale. At the top of the section, a wide range of sizes of rounded folds are developed in micaceous quartzite and grunerite-carbonate iron formation (Plate 4-6a, b).

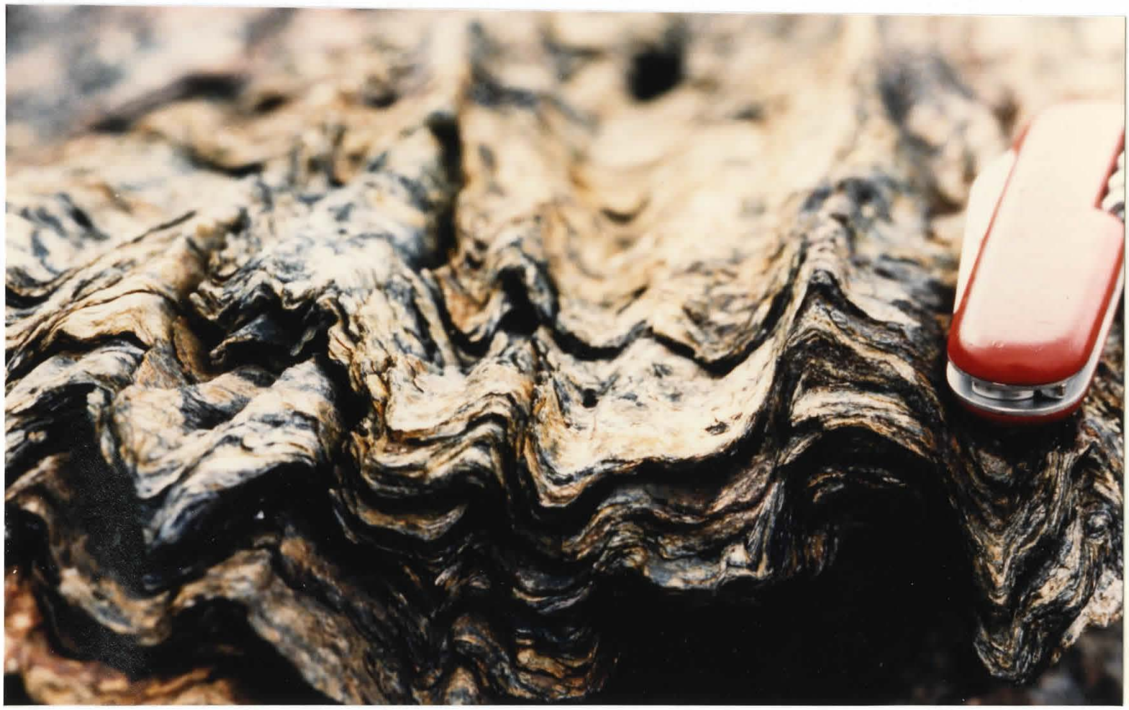
This apparent contradiction to the general trend is perhaps due to variable S_1 development. In this section, D_2 fold development shows a positive correlation with S_1 development. Both S_1 and D_2 folds are best developed toward the top of the Lac de la Grunerite section.

The correlation of poor D_2 fold development in weakly S_1 foliated rocks is observed in the other measured sections, but to a lesser degree. More massive lithologies (such as the metabasites) commonly display less D_2 folding that adjacent, more micaceous lithologies.

In order to develop angular-nosed chevron folds, a well developed pre-existing tectonic anisotropy is required (Ramsay, 1967). Best developed chevron folds are normally seen in micaceous lithologies and micaceous layers in more quartzose lithologies. Once folds are initiated, slip occurs on the micaceous S_1 surface. Chevron folds are usually tighter (i.e. smaller apical angle) than corresponding rounded folds, as once the folds are initiated and slip is occuring, it becomes progressively easier to continue folding, rather than initiate new folds (Ramsay, 1967).

Plate 4-6a: Rounded folds (D_2) of grunerite-carbonate schistosity (S_1), near Lac de la Grunerite. Hammer head is 15cm long.

Plate 4-6b: Angular (left) and rounded (right) D_2 folds in grunerite-carbonate schist, near Lac de la Grunerite. Pen knife is 2cm wide.



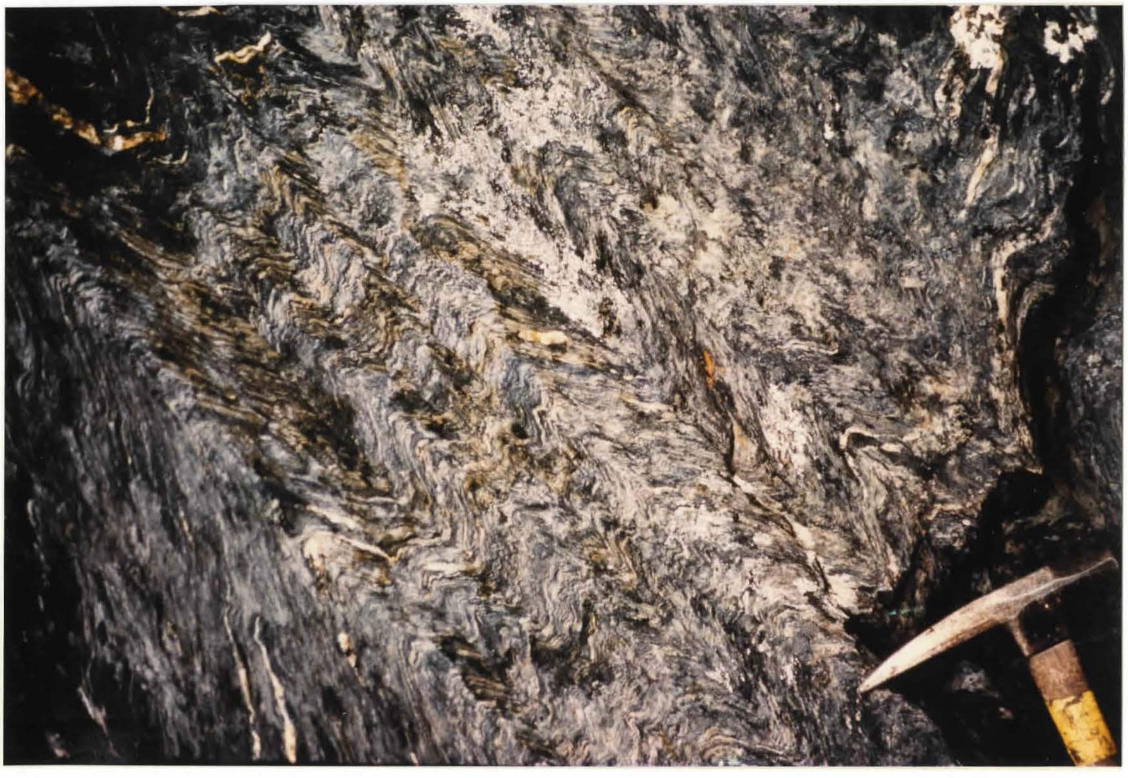
The F_2 mesoscopic folds are generally not cylindrical, but form continuous fold trains at outcrop scale. These folds are characterized by distinct limb asymmetry (Plate 4-7). In the northern macroscopic synform, these mesoscopic asymmetries are different on opposite sides of the synform (see Maps G, H, I). In both cases, the relative sense of movement suggested is cover rock off-basement, toward the macro-synform core. This may be the combined result of diapiric rise of the antiforms' tonalitic basement core/gravitationally induced sinking of overlying mafic-ultramafic thrust slices, and north-south compression and resultant folding during D_2.

However, the change in asymmetry from one limb to the other does not coincide with the mapped hinge trace of the macrscopic synform. While the northern macrolimb contains mesolimbs of one asymmetry (south verging), the south macrolimb contains both north- and then south-verging mesoscale folds.

The D_2 folding event develops an axial planar fabric (S_2) in favourable lithologies, usually semi-pelite and micaceous quartzite. In outcrop, it is generally seen as an irregular, spaced cleavage (see Plate, 4-8). Examination of thin sections confirms that S_2 is notable only in more micaceous lithologies, and can generally be classified into two types, according to Gray (1977a,b). Wide, diffuse cleavage zones with indistinct boundaries produced by

Plate 4-7: Macroscopic D_2 fold limb asymmetry, suggesting top side over to the right. Orthoquartzite, southwest shore of Wakeham Bay. Happy slug is 1.85m tall

Plate 4-8: Folds of S_1, showing development of an axial planar spaced cleavage (S_2), in micaceous quartzite. South of Cripple Creek section. Hammer head is 15cm long.



microfolding are known as zonal type (Plate 4-9a), while thin, sharply defined cleavage discontinuities which truncate the initial (S_1) fabric are known as discrete type (Plate 4-9b).

Some controversy exists as to whether the discrete type of crenulation cleavage is the most advanced stage of zonal type crenulation cleavage (Hanmer, 1979; Gray, 1977a, 1979a), or a feature developed either pre-, syn-, or post-microfolding (Swager, 1985). In sections examined, a nearly continuous spectrum from zonal to discrete crenulation cleavage is observed (Plate 4-10).

Where zonal crenulation cleavage is insipient, rather broad microfolds of S_1 are noted (see Plate 4-5). In this case, the concentration of quartz/feldspar in microfold noses is only slightly higher than on the limbs. As cleavage generation continues, the limbs become distinctly quartz/feldspar poor (see Plate 4-9a), as micas are rotated into a new orientation, roughly axial planar to the microfolds (Gray, 1979b). The limbs are routinely asymmetrical, long limb defining S_2 cleavage (Hobbs et al, 1976). Ultimately, with continued compression, no quartz/feldspar is present in the realigned mica seams, which are now rotated such that they truncate the earlier S_1 fabric (see Plate 4-9b).

The compositional variation from quartz/feldspar rich-microfold noses to quartz/feldspar poor microfold limbs/S_2 cleavage traces suggests preferential removal of quartzofeldspathic material from the

Plate 4-9a: Zonal-type D_2 crenulation cleavage in micaceous quartzite, showing a diffuse S_2 cleavage zone. Field of view is 1.5mm by 2.5mm, crossed polars.

Plate 4-9b: Discrete-type D_2 crenulation cleavage, showing a sharply defined S_2 surface truncating the pre-existing S_1 surface in micaceous quartzite. Field of view is 1.5mm by 2.5mm, crossed polars.

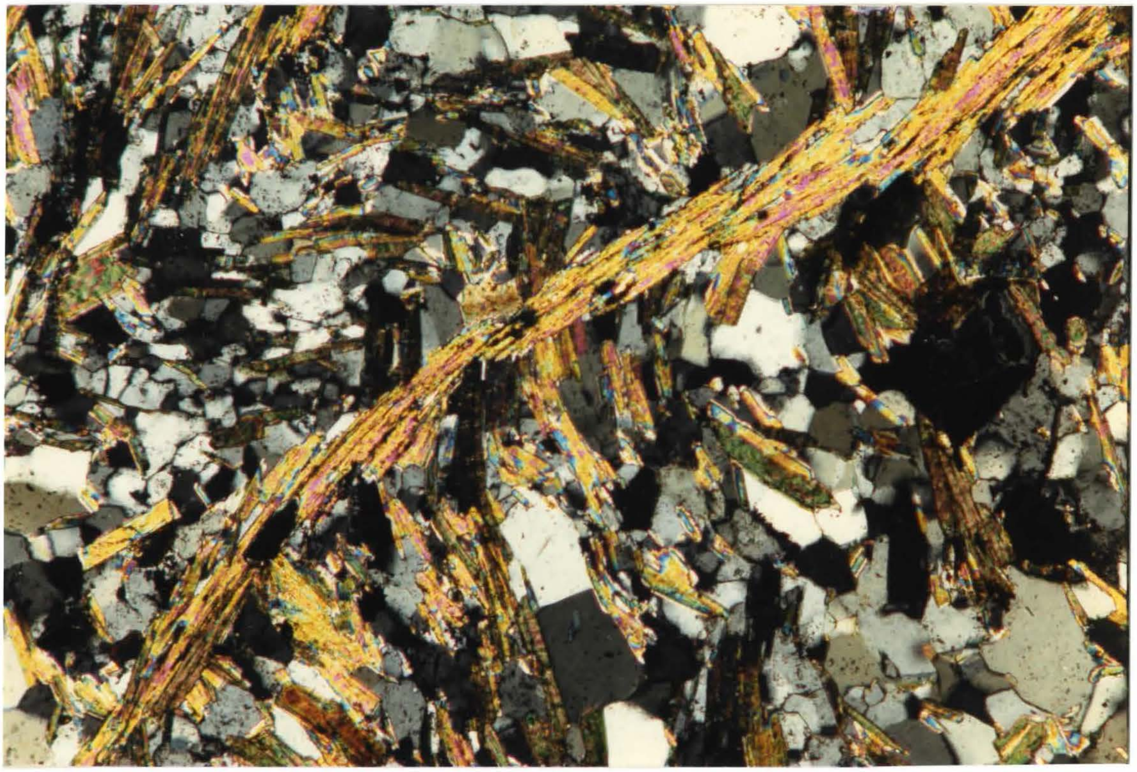


Plate 4-10: Continuous spectrum of crenulation cleavage development, from broad, zonal-type, to truncating, discrete-type. Semi-pelite, field ofview 2.0mm by 3.0mm, plane light.

limbs, passively concentrating residual micaceous material. This, coupled with the ultimate local truncation of S_1 fabric, supports the idea of dissolution and diffusion of soluble minerals (quartz and feldspar) from fold limbs -- a mechanism known as pressure solution (Gray, 1979b). The dissolved material may be deposited locally, in areas of relatively lower stress. This mechanism may account for the widespread occurance of the quartz veins/segregations so commonly observed in the semi-pelite and micaceous quartzite lithologies.

4.3 D_3 Structures

The third major deformational event (D_3), is a late, northwest-trending cross folding event, producing folds with typically kilometer-scale wavelengths. The effects of D_3 are recognized by the systematic plunge reversal of D_2 fold axes. D_3 fold axis traces are readily recognizable at map scale, dividing areas of east- and west-plunging D_2 fold axes (see Maps F, G, H). This basement involved folding is thought to be responsible for the minimum 15 km of crustal thickness exposed in oblique sections at the eastern end of the Belt.

CHAPTER 5: SYNTHESIS

5.1 Discussion

Interpretation of Isograds

According to Miyashiro (1961), each metamorphic terrain is characterized by a certain metamorphic facies series. The sequence of metamorphic grade zones at the eastern end of the Cape Smith Belt is Barrovian in nature (Barrow, 1893). Zones observed in pelitic rocks correspond to Barrovian Biotite -, Almandine- (Tilley, 1925), Staurolite-, and "Cyanite"- (Kyanite) zones (Barrow, 1893). This sequence is "kyanite-sillimanite" type, in the terminology of Miyashiro (1961).

The above zones are identified by the first occurrance of the index mineral whose name the zone bears. The surface separating adjacent zones is known as an isograd. Tilley (1924), in defining the term isograd, incorrectly assumed that isograds joined points of equal pressure and temperature (Thompson, 1976). Rather, they delineate the occurance of a specific, mineral-producing reaction that may be occuring over a range of pressure and temperature conditions.

The locations of the isograds shown in Figures 3-1 to 3-4 are somewhat approximate. Subsequent detailed mapping will undoubtedly refine their locations. The garnet isograd shown in Fig. 3-4 is exceptional in this respect, as its delineation was the subject of several days of detailed mapping in 1985 by Normand Begin (see St-Onge et al, 1986, Fig. 1.20).

As viewed in cross section, the various mineral isograd surfaces have a generally southwestward-dipping orientation. This, combined with their sequence of increasing grade to the northeast, indicate that metamorphism in the belt is the normal, "hot-side-down" type (St-Onge, 1981; 1986).

Several of the isograds are seen to intersect one another, such as the garnet isograds in Figs. 3-1 and 3-2. Garnet is produced by a different reaction in each lithology (from biotite + muscovite in 3-1, quartz + biotite in 3-2). The observed intersection in physical space implies intersection in P-T space (Carmichael, 1969; 1970). Only at the intersection point (in both P-T and physical space) are both reactions occuring at the same conditions of presssure and temperature.

Lower grade isograds, such as the pelitic "garnet-in" isograds (Figs. 3-1, -2, -3), and the metabasite "actinolite-out" isograd (Fig. 3-4) are not as severely folded by D_2 as the higher grade pelitic "staurolite + kyanite-in" isograd (Fig. 3-3) and metabasite "garnet-in" isograd (Fig. 3-4). The higher grade isograds conform well to macroscopic D_2 fold structures, such as the northern synform (see Fig. 1-3). This may indicate that the higher grade isograds were "set" prior to D2. Occurring in the same metamorphic event, the lower grade isograds were not set until late- to post-D_2, as uplift and cooling continued.

D_1 strain decreases stratigraphically upwards, and iso-stratigraphically from north to south. Within the D_1 event, thrusting predates high strain in the basal shear zone, as shown by the boudinaged sill described in section 4.1. Thrust geometry suggests southward transport and imbrication of the thrust slices.

Development of D_2 folds is most pronounced in rocks with a strong metamorphic schistosity (S_1). D_2 fold morphology varies from rounded- to chevron-style with increasingly micaceous S1 schistosity. Intensity, size range, and style variation of D_2 folds decreases stratigraphically upwards. If intensity of folding (i.e. number of folds per unit area or volume) can be interpreted as correlating positively with intensity of deformation (strain), and hence applied stress, then stress decreases stratigraphically upward.

D_1 deformation culminates at higher pressure and temperature conditions than does D_2. This is supported by development strongly schistose metamorphic fabric, (S_1), and high ductile shear strain during D_1. In contrast, D_2 and D_3 are not characterized by metamorphic mineral growth, and strain is primarily displayed as relatively open folds. It is suggested that the D_3 event occurs at lower P-T conditions than D_2, as there is no evidence of mineral growth or planar fabric development during this event. D_3 fold geometry is very broad and open.

Timing of Metamorphism

Pressure and temperature maxima during metmorphism are not necessarily reached at the same time (England and Richardson, 1977). In a terrain thickened by overthrusting, maximum pressure will be reached before maximum temperature. Transmission of applied stress through a rock will be almost instantaneous compared to the rate of temperature increase. The rate of upward relaxation of the isotherms depends primarily upon the thermal conductivity of the thrusted pile, in the absence of igneous intrusions (England and Thompson, 1984).

In many orogenic belts, it is possible to deduce metamorphic pressure-temperature (P-T) pathways. These "piezothermic arrays" (Richardson and England, 1979) may be determined in a quantitative manner using appropriate geothermometer-geobarometer pairs and microprobe analyses of zoned minerals (Selverstone et al, 1984). It is also possible to generate a P-T path using microprobe data from a single zoned, poikiloblastic mineral. Poikiloblastic garnets, containing inclusions of biotite, plagioclase, quartz and Al_2SiO_5 polymorphs, were used in a study by St-Onge (1986; St-Onge and King, 1986). Appropriate geothermometer/geobarometer pairs were appplied to analyses of sub-assemblages composed of entrappped inclusions and adjacent host garnet in pelitic schists from Wopmay Orogen, N.W.T.. Quantitative treatments, using reasonable estimates of thermal conductivity values, rates of loading and available time have been developed for the Eastern Alps (Oxburg and Turcott, 1974; Oxburgh and England, 1980).

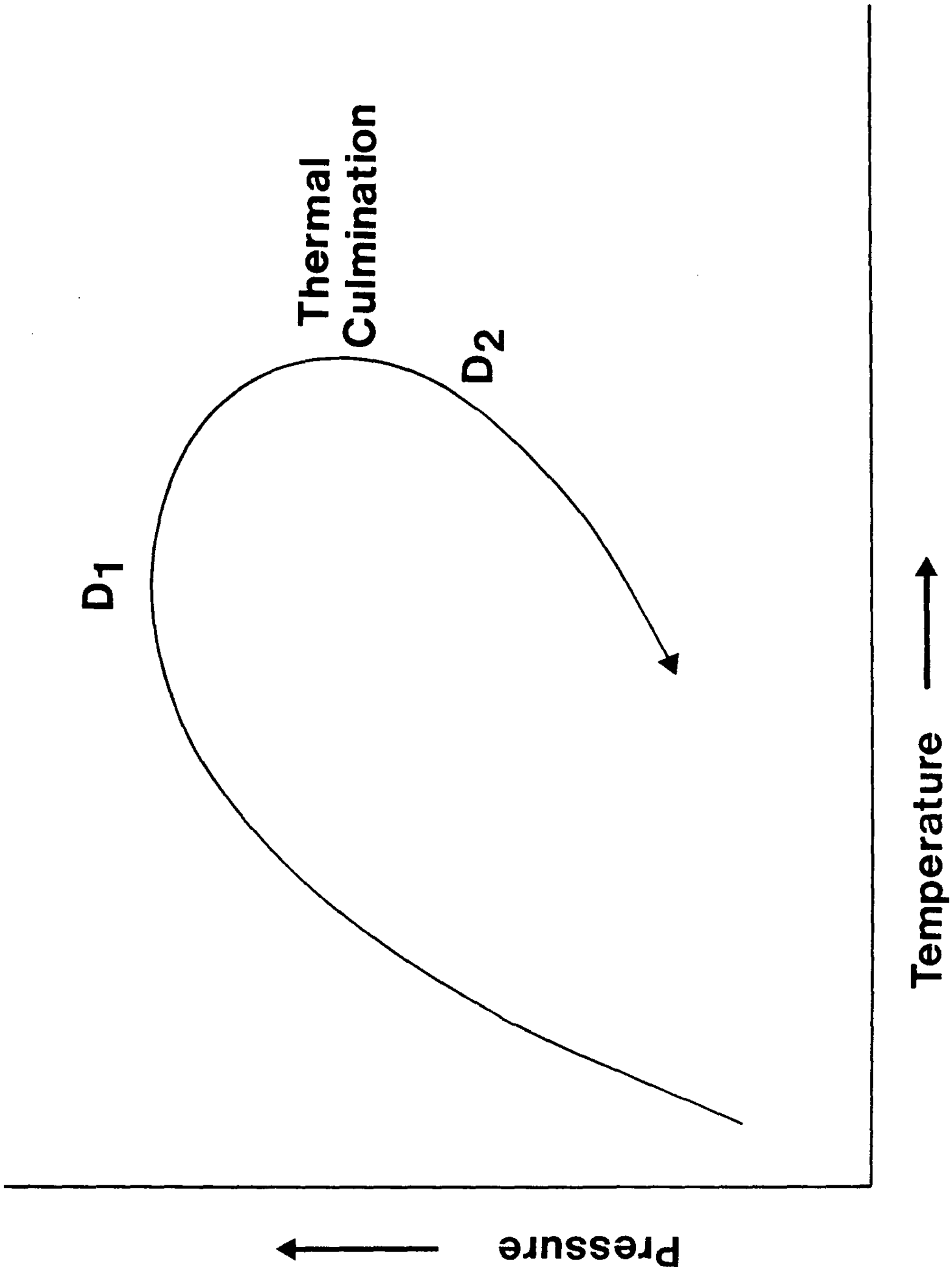
Postulated P-T Pathway: Eastern Cape Smith Belt

In the absence of the microprobe analyses required for precise geothermometry/geobarometry, an approximate P-T pathway is proposed, based on the interpretation of textural relationships in thin section. However, as this method relies on data gathered from more than one sample location, it is open to some ambiguity (Oxburgh and England, 1980). Each mineral grain within the belt undergoes its own P-T history. While this history will be practically identical for adjacent grains, significant variation in P-T history will occur at various points throughout an orogenic belt. (St-Onge and King, 1986, Fig. 16). Despite this, the qualitative path generated from samples of restricted areal distribution will yield a first order approximation of the shape of the P-T path.

Figure 5-1 shows the general form of the P-T path postulated for the eastern Cape Smith Belt, based on interpretation of textural relationships observed in samples gathered in the basal, thrusted sequence.

Figure 5-1: **Piezothermic Array,** Eastern Cape Smith Belt

Postulated, relative Pressure-Temperature pathway through time, as generalized for the Wakeham Bay area. Initially, pressure increases and peaks (D_1) before the thermal culmination. A second, lower stress event (D_2) postdates the thermal culmination.



Thermal Culmination
D_2
D_1
Temperature
Pressure

During imbrication (D_1), pressure rises more rapidly than does temperature. Mineral growth begins once the appropriate thermal threshold for a given metamorphic mineral is reached. The load stress exerted by the thrusted slices causes the alignment of platy minerals, creating the metamorphic schistosity (S_1) (see Plate 4-1a) axial planar to tight folds of bedding.

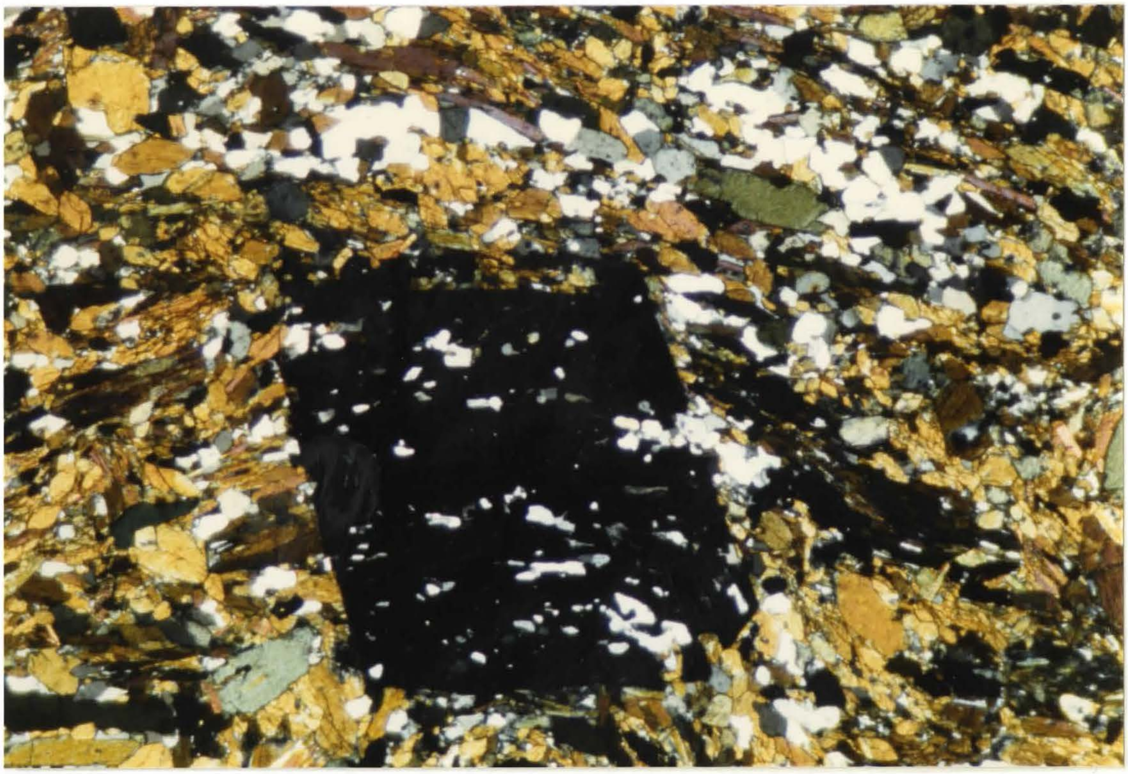
The thermal culmination occurs after peak D_1 stress. This is interpreted from samples in which the highest grade minerals in the assemblage are overgrowing the S_1 fabric (Olesen, 1978; Vernon, 1978; Zwart, 1960) (See Plate 5-1a,b). The decrease in pressure is likely due to the cessation of imbrication, followed by the onset uplift and erosion, while the increasing temperature is due to the upward relaxation of isotherms.

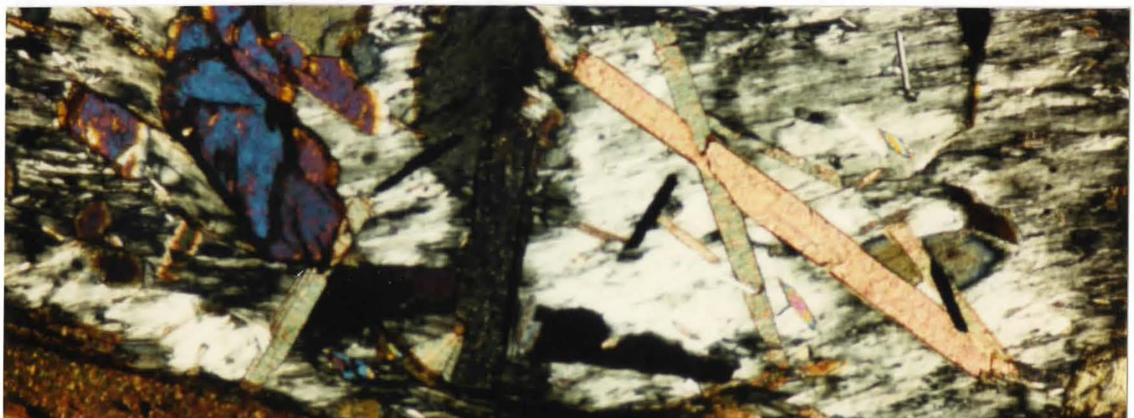
The second deformation (D_2) folds S_1, creating mesoscopic folds, with crenulations visible in thin section (Plate 5-2). Growth of new minerals is generally not associated with the formation of S_2. Rather, the deformation tends to either rotate S_1 minerals into an axial planar S_2 fabric (See Plate 4-9a) or create cleavage by pressure solution (See Plate 4-9b).

Some mineral growth does postdate D_2. Muscovite is rarely seen randomly overgrowing D_2 structures (see Plate 5-3a) Chlorite is a common late metamorphic mineral, commonly seen altering biotite, hornblende, garnet and muscovite (see Plate 5-3b). Both the

Plate 5-1a: Poikiloblastic, euhedral garnet (black rhomb) overgrowing (post-dating) the dominant, S_1 schistosity. Matrix mineralogy includes hornblende (gold and green), biotite (brown laths), and quartz/plagioclase (white and grey). Metabasite, field of view 1.5mm by 2.5mm, crossed polars.

Plate 5-1b: Biotite laths (browns + greens) and hornblende (blue) randomly overgrowing (post-dating) the dominant, S_1 schistosity defined by chlorite (off-white). Mafic sediment, field of view 1.5mm by 2.5mm, crossed polars.





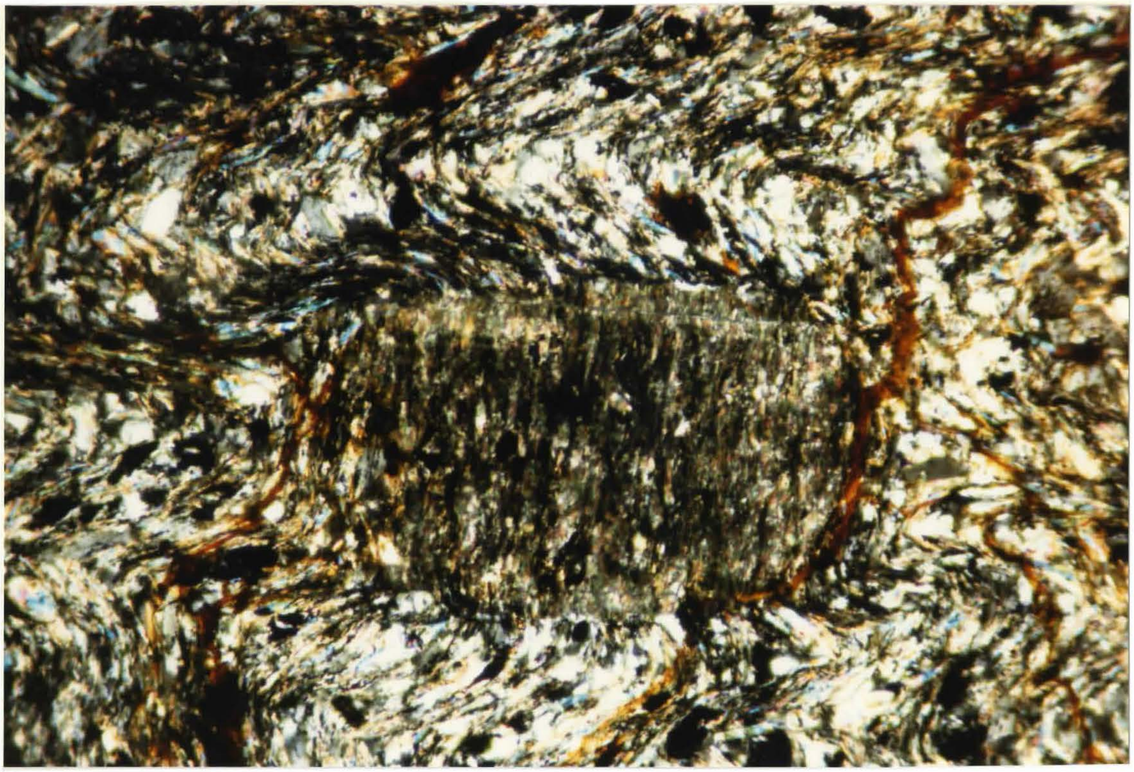
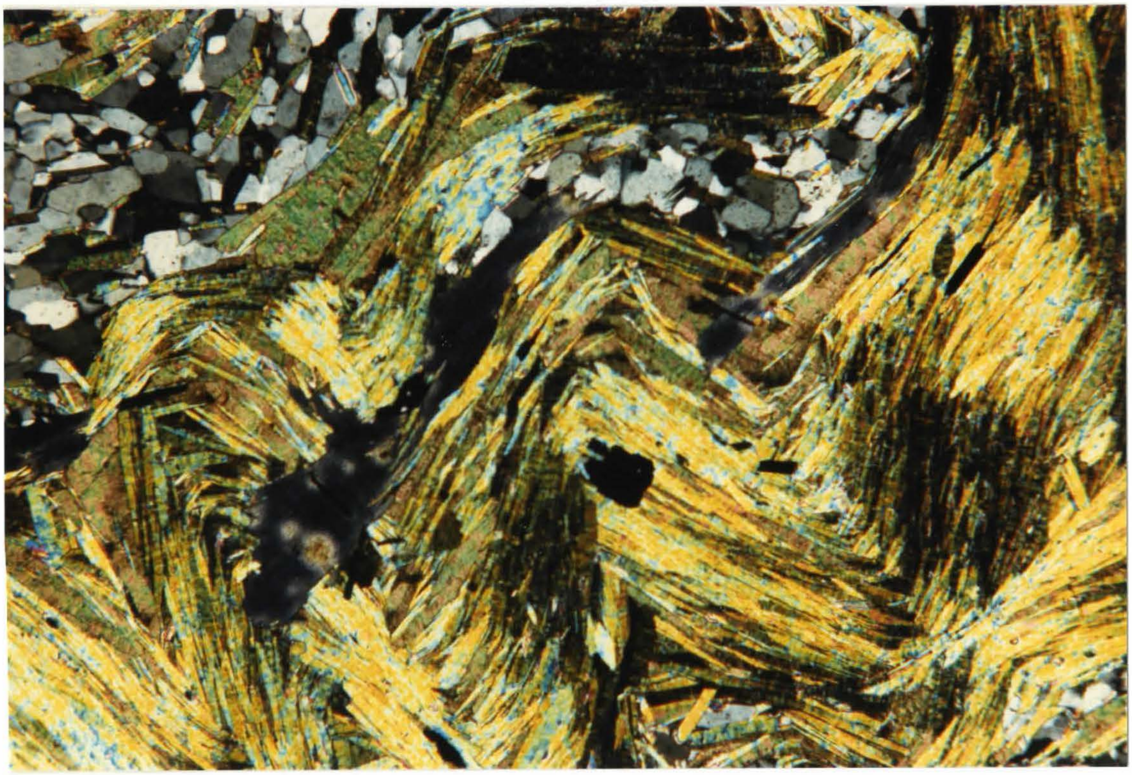


Plate 5-2: Large, interkinematic biotite poikiloblast with straight inclusion traces (SI) in a matrix deformed during D_2. Poikiloblast grew post-D_1, pre-D_2. Matrix is composed mainly of quartz, with biotite and muscovite defining S_1. Mafic sediment, field of view is 0.9mm by 1.5mm, crossed polars.

Plate 5-3a: Muscovite laths (pale blue) overgrowing (post-dating) D_2 folds, defined by S_1 laths of biotite (brown, pale green). Mafic sediment, field of view 0.9mm by 1.5mm, plane light.

Plate 5-3b: Chlorite (dark blue with yellow patches) growing after D_2-folded biotite (yellow and green) in micaceous quartzite. Field of view is 1.5mm by 2.5mm, crossed polars.



muscovite and chlorite must have grown post D_2, as they are not preferentially aligned, and they cross cut D_2 structures.

5.2 Environmental Interpretation

The ironstones on the southern margin have been classified as "Metazoan-poor, Extensive, Chemical-Sediment-rich, Shallow Sea Iron Formation" (MECS-IF) by Kimberley (1978). Characteristics of the Lac de la Grunerite formations, such as thickness (up to 500m) and areal extent, correspond well to this type. Similar ironstones can be recognized 150km to the west (Moorhead, 1985). MECS-type iron formations are dominant in the Lower Proterozoic, commonly composed of grunerite when metamorphosed.

This rather cumbersome classification nomenclature was simplified by Gross (1980). Under his scheme, the Lac de la Grunerite formations would correspond to silicate and carbonate facies of Superior-type iron formation, indicative of near-shore, shallow water, continental shelf environments (Gross, 1980; Kimberley, 1978; Windley, 1984). This model is supported by the presence of a thin, conglomeratic horizon lying stratigraphically below the iron formation. Because very little detrital material is found in the Belt iron formations, a chemical precipitation of iron and silica is indicated (Kimberley, 1978; Gross, 1980). The source of the iron may be local fumarolic or hydrothermal activity, or leaching of Fe-rich rocks (Gross, 1980), located further offshore in this case.

Formation 2 micaceous quartzites and semi-pelites are interpreted as greywackes and siltstones respectively. This clastic input is capped by a sparsely developed carbonate horizon. The development of these carbonates suggests a quiet environment, essentially devoid of clastic detrital input (Blatt et al, 1980), such as a marginal basin shelf (Miall, 1984).

Overlying the carbonate horizon are the thick, homogenous, micaceous quartzites of Formation 3, also interpreted as greywackes. Their origin is uncertain. The most probable derivation would be from the dominantly quartzofeldspathic Superior province to the south, but sialic crustal material to the north, possibly the Sugluk Terrane (see Fig. 1-2), cannot here be ruled out.

The orthoquartzites and semi-pelites of Formation 4 may represent an offshore, deep water environment (Walker, 1984). The orthoquartzites may represent turbidites, grading into deep water muds, preserved as semi-pelites. This interpretation will be tested by subsequent mapping of lower grade, less strained equivalents.

Overlying these four Formations are the dominantly basaltic, massive to pillowed flows and sills of Formations 5-7, comprising the rest of the stratigraphy of the Belt. The voluminous ironstones of

Formation 1 and the "mafic sediment" units throughout Formations 2-4 are probably a reflection of intense, offshore volcanic activity, occurring during extensional thinning of the crust during continental rifting (Dickinson, 1980). Leaching of the basic volcanics so produced is a possible iron source for the ironstones of Formation 1 (Gross, 1980). The mafic sediment horizons may be air-fall or submarine volcaniclastic debris (Carey and Sigurdsson, 1984).

A non-balanced palinspastic restoration of the thrusted blocks suggests that the vertical sequence observed in the Belt may actually be the lateral, on- to off-shore sequence of the pre-thrust basin. Tentatively, Formation 1 conglomerate and ironstones represent the shoreline/shallow water environment. Formation 2 mudstones and carbonate represent the build-up of a shallow water shelf. Formation 3 greywackes represent a deeper water environment, and Formation 4 orthoquartzites and mudstones are deep, basinal representatives. The overlying volcanics of Formations 5-7 represent a zone of intense crustal rifting. This oversimplified interpretation requires further sedimentalogical work on lower grade, less strained equivalents.

5.3 Conclusions

Metamorphism in the eastern Cape Smith Belt is Barrovian in nature, with Biotite, Almandine, Staurolite and Kyanite Zones present in sequence. Pressure and temperature, and hence metamorphic grade, systemically increase from southwest to northeast across the study area. Pressure and temperature estimated from the lowest grade assemblages in the study area are on the order of 2-4 kbar and 400°C. Estimates from the highest grade assemblages suggest minimum pressures of 5.5 kbar and temperatures of 575°C. Thus, metmorphic conditions in the area may be classified as varying from low- to medium-grade (Winkler, 1979), or lower-greenschist to mid-amphibolite facies.

The first identifiable deformation event produced signficant crustal shortening by imbrication along low-angle reverse faults or thrusts. A second, roughly co-axial event folded both cover and basement at all scales of observation. A late cross folding event (D_3) provides an oblique view through the crust on a major D3 fold limb in the study area. The structural relief of over 12km so provided allows the direct observation of the results of the two preceding deformation events.

The generalized P-T pathway determined fits the pattern established for deformed belts, such as the eastern Alps, in which crustal thickening by thrusting and imbrication (D_1) leads to a thermal culmination. This is followed by the late- to post-metamorphic co-axial folding event (D_2) during cooling by uplift and erosion. Thrust geometry suggests an overall southward transport of thrust slices.

The observations and results of this study are consistent with the current tectonic model proposed for the area (see Hoffman, 1985). The vast majority of the Belt (all parts north of Lac de la Grunerite) are allochthonous, the result of southward directed thrusting. Strictly speaking, the Belt is not a kilppe, as suggested by Hoffman (1985) (H. Helmstaedt, p. comm., 1986). The entire Belt is not fully isolated from its source area -- the autochthonous section on the southern margin is not tectonically transported. The area is best described as a Fold-Thrust Belt.

Tectonically, the rocks of the Belt may represent a passive marginal to crustal rifting transition, imbricated during collision of two crustal blocks -- the Superior block to the south and the Sugluk Terrane to the north.

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FIGURE X
Tectonostratigraphic Correlation
Eastern Cape Smith Belt