STRUCTURAL ANALYSIS

OF THE

PAINT LAKE DEFORMATION ZONE,

NORTHERN ONTARIO

by

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The Paint Lake Deformation Zone (PLDZ), located within the Superior Province of Canada, demarcates a major structural and lithological break between the Onaman-Tashota Terrane to the north and the Beardmore-Geraldton Belt to the south. The PLDZ is an east-west trending lineament, approximately 50 km in length and up to 1 km in width, comprised of an early ductile component termed the Paint Lake Shear Zone and a late brittle component known as the Paint Lake Fault.

Structures associated with PLDZ development including S-, C- and C'-fabrics, stretching lineations, slickensides, C-C' intersection lineations, Z-folds and kinkbands indicate that simple shear deformation dominated during a NW-SE compressional event. Movement along the PLDZ was in a dextral sense consisting of an early differential motion with south-side-down and a later strike-slip motion.

Although the locus of the PLDZ may in part be lithologically controlled, mylonitization which accompanied shear zone development is not dependent on the lithological type. Conglomerate, intermediate and mafic volcanic units exhibit similar mesoscopic and microscopic structures where transected by the PLDZ.

Field mapping, supported by thin section analysis, defines five strain domains increasing in intensity of deformation from shear zone boundary to centre. A change in the dominant microstructural deformation mechanism from
dislocation creep to diffusion creep is observed with increasing strain during mylonitization. C'-fabric development is temporally associated with this change. A decrease in the angular relationship between C- and C'-fabrics is observed upon attaining maximum strain intensity.

Strain profiling of the PLDZ demonstrates the presence of an outer primary strain gradient which exhibits a simple profile and an inner secondary strain gradient which exhibits a more complex profile. Regionally metamorphosed lithologies of lower greenschist facies outside the PLDZ were subjected to retrograde metamorphism during deformation within the PLDZ.
Thanks to all faculty, staff and students who have made my term at Brock most enjoyable. Greg Finn and Rick Cheel freely imparted their knowledge in their respective fields. Peter Brown provided drafting assistance and Candy Kramer completed final stages of thin section preparation. Office-mates Shane Buck and Danny Soo stimulated much discussion concerning problems in Archean geology. Gerry Lawson provided field assistance for some of the data collection. Special thanks to Howard Williams for his time, patience and guidance during his role as thesis supervisor. Finally, hugs and kisses for my wife Cheryl who was fully supportive throughout this study.
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CHAPTER I - INTRODUCTION

I.I LOCATION AND ACCESS

The study area is located in northern Ontario, approximately 220 km northeast of the city of Thunder Bay (Fig. 01). More specifically, it lies between Beardmore and Geraldton just north of the community of Jellicoe. Secondary Highway 801 defines the western boundary of the map area and the Camp 40 Road is the eastern limit. Both are gravel roads branching north from Highway 11 in the south. The study area measures approximately 25 km x 5 km and lies within the northern segment of Walters, Leduc and Legault Townships.

Access to the study area is provided by a lake and river system in which transportation was made possible by canoe. Outcrop in the western half of the area is continuous along the shore of Paint Lake which parallels the regional lithological strike. However, across strike away from the lake shore, much of the bedrock is covered with lichen and overburden. In the eastern half the quantity and quality of outcrop decreases as Paint Lake connects with the Namewaminikan River where swamp and overburden cover increases. Therefore, most of the detailed mapping was accomplished along the lakeshore of Paint Lake. It is reasonable to assume that perhaps some of the most interesting geology in the area lies at the bottom of Paint Lake.

Additional to lakeshore and riverbank mapping, short
Fig. 01. Location of study area.
north-south trending pace-and-compass traverses were made on foot. Roadcuts along secondary Highway 801 and the Camp 40 Road provided further outcrop.

I.II PREVIOUS WORK

Geological work began in the region as early as 1869. Bruce (1937) and Laird (1937) were the first to publish a comprehensive report on the region in which they made reference to the activities of the early workers (p. 2-3). Subsequent geological mapping in the region was completed by Horwood and Pye (1951), Peach (1951), and Pye (1951). In 1966 Pye et al. published a geological compilation map of the Tashota-Geraldton area. Reports by Mackasey (1976) and Mackasey and Wallace (1978) of the Ontario Geological Survey have provided an excellent database from which the base maps for this study were constructed. More recently, mapping has been completed by Devaney and Fralick (1985), Barrett and Fralick (1985), Kehlenbeck (1986), Williams (1986), Kresz and Zayachivsky (1986) and Devaney (1987).

I.III DEFINITION OF THE PAINT LAKE DEFORMATION ZONE

This zone was first named the Devil's Walk-Paint Lake Fault by Tyson (1945). Mackasey (1976) retained the name Paint Lake Fault. However, this structure is more than a mere fault, or fault zone, and it will be referred to as the Paint Lake Deformation Zone (PLDZ) in this study. The PLDZ consists of
two major components: an early ductile structure, here labelled the Paint Lake Shear Zone (PLSZ) and a late brittle structure, known as the Paint Lake Fault (PLF). The PLSZ is identified by a sequence of foliations which increase in intensity toward the shear zone centre, while the PLF is recognized by steep faces of lakeshore outcrop along Paint Lake and the truncation of late Proterozoic diabase dikes which abut against this lineament. Figure 02 is a simplified sketch of the PLDZ and its two main components. The centre of the PLSZ has been faulted by the PLF and both lie at the bottom of Paint Lake, hence, the terms shear zone centre and PLF will be used synonymously when referring to location within the PLDZ.

The PLDZ forms an east-west trending lineament which can be traced approximately 50 km from Lake Nipigon to the eastern boundary of Leduc Township using Landsat imagery. It is seen as a topographic depression on aerial photographs, typically connecting small lake and river systems. The PLDZ anastomoses on a regional scale with a wavelength of approximately 20 km and an amplitude up to 1 km. The PLDZ generally behaves as one long, continuous zone except for a splay, known as the Brookbank Fault, which diverges from the main zone to the west of the study area and splays from the PLDZ in the eastern boundary of the map area which may represent termination of the PLDZ.
Fig. 02. Definition of PLDZ = PLSZ + PLF.
I.IV SCOPe OF STUDY

Structural, lithological, and metamorphic mapping at 1:15,840 scale, including detailed mapping of structurally significant outcrops, was supplemented with thin section studies to analyze and document the following features in the map area:

1) geometry and dimensions of PLDZ
2) sense and amount of displacement
3) type and timing of deformation mechanism(s) responsible for development of PLDZ
4) lithological and structural changes across PLDZ
5) mesoscopic and microscopic structures associated with PLDZ
6) effects of increasing strain in volcanic and sedimentary rocks with progressive deformation
7) relationship of metamorphism to deformation
8) significance of PLDZ in the evolution of Archean supracrustals in the Superior Province of Canada
CHAPTER II - GENERAL GEOLOGY

II.I TWO DISTINCT TERRANES

The general geology of the study area is divided into two distinct terranes separated by the PLDZ (Fig. 03). South of the PLDZ metasediments form three east-west trending belts which are separated by three major mafic metavolcanic belts. Mackasey (1976) mapped the northern metasedimentary belt (NMB) as predominantly conglomerate; the central metasedimentary belt (CMB) as conglomerate, pebbly sandstone, siltstone, and argillite; and the southern metasedimentary belt (SMB) as interbedded argillite, siltstone and wacke sandstone. Devaney and Fralick (1985) have informally classified the rocks within the NMB and the CMB as the Namewaminikan Group. Ironstone is present in the CMB and more commonly in the SMB as described by Barrett and Fralick (1985). The mafic metavolcanic rocks are dominantly massive, but zones containing amygdules and pillows are present. This metasedimentary-metavolcanic sequence is known as the Beardmore-Geraldton Belt (BGB).

The BGB extends from the east shores of Lake Nipigon to the Geraldton area, however, correlation of the NMB in Legault Township with the metasedimentary belt in Geraldton is hindered by overburden. Devaney (1987) mapped these metasedimentary belts as two separate units. Metasedimentary belts similar to the BGB may be lithostratigraphically correlated along the regional strike west of Lake Nipigon.
Fig. 03 Regional geology of BGB and OTT (after Williams 1986).
The geology north of the PLDZ is dominated by intermediate to felsic metavolcanics in conformable contact with an underlying subordinate mafic sequence (Mackasey and Wallace 1978). Intermediate pyroclastic rocks are the most abundant metavolcanic rock and include pyroclastic breccia and less commonly lapilli tuff, crystal tuff and porphyritic flows. Felsic metavolcanics occur in some outcrops north of the study area (Mackasey and Wallace 1978). Mafic metavolcanics consist of massive and amygdaloidal lavas, pillow lavas and flow breccia. The metavolcanics have been intruded by small subcircular massive stocks of trondhjemitic composition (Mackasey and Wallace 1978; Kresz and Zayachivsky 1986). This terrane is known as the Onaman-Tashota Terrane.

All lithologies of Archean age, excluding Proterozoic diabase dikes, have been subjected to lower greenschist facies metamorphism in the study area. Hereafter, the prefix "meta" will be omitted. A brief description of lithologies within the study area is given below and reference should be made to Map 1 - Regional Geology. More extensive descriptions are presented by Mackasey (1976), Mackasey and Wallace (1978), Devaney and Fralick (1985), Barrett and Fralick (1985), Kresz and Zayachivsky (1986), Devaney (1987) and Devaney and Williams (manuscript).
II.II GEOLOGY OF BEARDMORE-GERALDTON BELT

II.IIa MAFIC VOLCANICS

Massive mafic volcanics are the most abundant rock type in the northern mafic volcanic belt. Weathered surfaces are medium green in colour and fresh surfaces are dark green. These rocks have a relatively homogeneous, equigranular texture in outcrop and become strongly foliated towards the centre of the PLDZ. Grain size generally varies from 0.1 mm to 1.0 mm, however, grain sizes up to 3.0 mm were measured. These coarse-grained rocks were interpreted as flow centres. Where primary volcanic structures or intrusive contacts were not observed, any distinction between coarse-grained flows and fine-grained gabbroic bodies was not possible. Vesicles and amygdules are locally present and commonly of spheroidal shape measuring 0.5 cm to 1.0 cm. Composition of the amygdules varies from quartz, carbonate, epidote and chlorite. Vesicles and amygdules are elongate, parallel to the stretching lineation, within the deformation zone.

Pillow basalts are common throughout the study area and are indicative of a sub-aqueous environment of lava deposition. Undeformed pillows are generally closely packed, ranging in size from 40 x 20 cm to 100 x 50 cm. Structural tops interpreted from pillow lavas indicate tops to the north (Fig. 04). Aphanitic selvages are a darker green colour than the interior of the pillows and maintain a constant thickness of 0.5 cm to 2.0 cm. Deformed pillows are elongate with their
Fig. 04. Relatively undeformed pillow lava indicating top to the north.

Fig. 05. Sheared pillow lava exhibiting length to width ratio exceeding 20:1.
long axes parallel to the foliation plane. Initial length to width ratios of approximately 2:1 now exceed 20:1 as pillows measure 100 x 5 cm to 150 x 10 cm (Fig. 05). Pillow selvages which are preserved in deformed pillows are up to 20 cm thick in the ends.

The northern mafic belt could not be subdivided into separate flow units because of irregular outcrop distribution and lack of well preserved primary volcanic structures. No evidence was found to interpret the presence of amygdules and vesicles as flow tops.

Most of the original minerals have been completely altered or replaced. However, primary textures may still be preserved. Intergranular, pilotaxitic and sub-ophitic textures are most common. Euhedral to subhedral lath-shaped plagioclase comprises up to 60% of the mafic volcanics with an average grain size of approximately 0.3 mm. Plagioclase grains have been altered by albitization, saussuritization and sericitization but retain their original shape (Fig. 06). Where albite twins are preserved, plagioclase grains were originally labradorite.

Amphibole occupies an interstitial position relative to the plagioclase or locally encloses plagioclase grains sub-ophitically. Pleochroism of the amphibole ranges from a medium green to a pale green and commonly colourless, characteristic of actinolite, probably after pyroxene. Actinolite grains are subhedral to anhedral in shape, may be fibrous, average 0.5 mm
Fig. 06. Plagioclase grains in mafic volcanic have been altered by albitization, saussuritzation and sericitization but retain their original shape (field of view ≈ 7mm).
in length and constitute approximately 15% of the rock. Fine-grained flakes of chlorite (25 μm) typically form aggregates (0.5 mm) which are interstitial to the plagioclase and comprise up to 25% of the mafic volcanics. Accessory minerals include quartz and apatite. Calcite is indicative of weak carbonatization which is pervasive. Fine-grained disseminated pyrite may also be present.

**II.IIb POLYMICTIC CONGLOMERATE**

A clast-supported polymictic conglomerate dominates the NMB. The conglomerate is typically massive, however, crude bedding is rarely defined by beds of different maximum clast size. Clast sizes range from pebbles to boulders (Fig. 07). Clast composition is variable but due to the close proximity of the NMB adjacent to the PLDZ centre, intense shearing makes some clasts indistinguishable from matrix. Detailed mapping by Devaney (1987) west of the study area, where the NMB diverges from the PLDZ, gives the following clast composition: felsic volcanic (67.2%), mafic volcanic (14.6%), granitoid (13.7%), felsic porphyry (2.9%), chert (0.5%) and quartz (0.3%). The conglomerate also possesses a sand-size fraction which constitutes the matrix. Hence, gravel size clasts and sand size matrix indicate a bimodal grain size distribution.

D₁₀ measurements (average length/width ratios in cm of the ten largest clasts at a site measured in the XZ plane of the finite strain ellipse) reached up to 65/14. The ten
Fig. 07. Clast size ranges from pebbles to boulders in conglomerate.

Fig. 08. Clasts exhibit a primary oblate shape which is accentuated in the plane of foliation during shearing.
Fig. 09. Quartz and feldspar grains are subangular in shape and poorly sorted in conglomerate (field of view ≈ 7mm).
largest clasts were generally of granitoid composition and less commonly felsic volcanics. A more extensive analysis of D\(^{18}O\) measurements is given by Devaney (1987). Undeformed clasts exhibit a primary oblate shape which is accentuated in the plane of foliation during shearing (Fig. 08). Clasts were not only elongated in the shear direction but also folded, particularly the less competent clasts. Interbedded pebbly sandstone constitutes a minor component of the NMB. These beds are massive and average less than a metre in thickness.

Matrices for the conglomerate and pebbly sandstone are medium grained (0.5 mm) feldspathic arenites. The matrix is comprised of approximately 40 % quartz, 30 % plagioclase, 15 % rock fragments and a finer microcrystalline fraction of quartzo-feldspathic material. Quartz and feldspar grains are subangular in shape and poorly sorted (Fig. 09). Plagioclase grains are weakly sericitized and albitized, but albite twinning is well preserved in relatively undeformed rocks. Other alteration minerals observed are chlorite and carbonate. Pyrite is generally present.

II.IIc CHERT AND JASPER LENSES

A thinly bedded (1-3 cm) chert lens, up to 3 m in thickness, is present 0.5 km north of Pasha Lake on the east side of Highway 801. The weathered surface is white to buff in colour and the fresh surface is light grey. Bedding is defined by alternating layers of light coloured chert and dark
ferruginous layers rich in fine grained (<0.5 mm) disseminated pyrite (Fig. 10). The iron-rich layers are non-magnetic. Bedding parallels the foliation and the contact with the mafic volcanic is concordant, yet the chert unit is discontinuous along strike forming a lens as it is displaced by the host mafic volcanic. Mackasey (1976) interpreted this structure as rafted portions of a chert layer that has been caught up in a lava flow.

From the same locality a relatively massive jasper unit is exposed on the west side of Highway 801. Weathered surfaces are light grey and fresh surfaces are pink in colour. The jasper layer measures approximately one metre in thickness and is bordered by chert layers similar to those described on the east side of the road. Bedding parallels foliation, contacts are concordant with the surrounding mafic volcanic and the jasperitic chert exhibits a lens shape.

Chert layers are composed predominantly of a microcrystalline (25 μm) to cryptocrystalline matrix of equigranular anhedral quartz grains. Alteration minerals include chlorite, epidote, and sericite. A peculiar "football-shaped" structure was observed in thin section, averaging 0.6 mm in length with chloritized rims approximately 40 μm thick (Fig. 11). These were interpreted as mineralization artifacts (Bill Parkins pers. comm.) possibly after an Archean organism.
Fig. 10. Bedding is defined by alternating layers of light coloured chert and dark coloured ferruginous layers.

Fig. 11. "Football-shaped" structures (mineralized artifacts) with chloritized rims in chert lens (field of view $\approx 7\text{mm}$).
II.III GEOLOGY OF ONAMAN-TASHOTA TERRANE

II.IIIa INTERMEDIATE VOLCANICS

Pyroclastic rocks are dominant in this terrane, commonly pyroclastic breccia (Mackasey and Wallace 1978). Pyroclastic breccia weathers pale green in colour with light green to grey fresh surfaces. Heterolithic fragments range in size from 2 x 1 cm to 25 x 5 cm and are generally more felsic than the matrix. These fragments are lenticular in shape, typically rounded to subrounded with their long axes parallel to the foliation (Fig. 12), however, feathery clasts were observed in some outcrops. Fragments may constitute up to 50% of the rock and some are porphyritic. The matrix is homogeneous, inequigranular and fine grained. Feldspar grains within the matrix measuring 2 to 3 mm produces a porphyritic texture. Crystal tuff may be intercalated with the pyroclastic breccia.

Porphyritic units lacking primary volcanic structures or intrusive relationships are present throughout the intermediate succession and are interpreted to be lava flows. Plagioclase phenocrysts, and less commonly quartz phenocrysts, range from 2 mm to 4 mm in size and sit in a fine grained matrix. Phenocryst density may reach up to 40% of the volcanic, but phenocryst distribution may be variable. Phenocrysts have a random orientation, except where aligned in the shear direction during deformation. A strong foliation develops approaching the PLF.

Thin section studies of typical intermediate volcanics
Fig. 12. Heterolithic pyroclastic breccia with lenticular fragments more felsic than matrix material.

Fig. 13. Plagioclase phenocrysts exhibit porphyritic, locally glomeroporphyritic, texture in pyroclastic breccia (field of view ≈ 7mm).
reveals approximately 25% subhedral plagioclase phenocrysts, with few quartz phenocrysts. Plagioclase phenocrysts are typically lath-shaped, rarely glomeroporphyritic, originally oligoclase-andesine in composition and commonly completely altered during albitization, saussuritization or sericitization (Fig. 13). The groundmass (<0.1 mm) ranges from a granular microcrystalline mixture of quartzo-feldspathic material to plagioclase microlites which exhibit a trachytic or pilotaxitic texture. Up to 10% actinolite and epidote is interstitial to matrix plagioclase. Minor amounts of chlorite, apatite, quartz and carbonate are also present in the groundmass.

Based on the colour index (mafic content) of these rocks, their porphyritic texture, and plagioclase composition the intermediate volcanics are interpreted to be of andesitic composition. Chemical analyses of the intermediate volcanics by Mackasey and Wallace (1978) indicate a calc-alkaline affinity. For reasons similar to those relating to the volcanics in the BGB, separate units could not be defined in the OTT.

II.IIIb MAFIC VOLCANICS

Massive, amygdaloidal, and pillow lavas similar in texture and composition to those south of the PLDZ are found to the north. Structural tops interpreted from undeformed pillows indicate tops to the south. The results of a chemical
analysis by Mackasey (1976) of a pillow lava north of the PLDZ indicates a tholeiitic basaltic composition.

Volcanic breccia, not observed south of the PLDZ, is present as a series of prominent outcrops jutting out from Kingston Island into Expansion Lake. Fragments are poorly sorted, loosely packed and angular to subrounded in shape (Fig. 14). These monolithic fragments consist of scoriaceous mafic lava averaging 15 x 10 cm in size in an aphanitic mafic matrix. Differential weathering of the matrix produces a rough surface characteristic of aa-aa flows (Mackasey 1976). Each individual outcrop protruding into Expansion Lake may represent separate flow units.

II.IIIc TRONDHJEMITE STOCK

A sub-circular stock, known as the Coyle Lake Stock (Mackasey and Wallace 1978), intrudes the volcanic rocks north of Paint Lake. The intrusion is a massive, medium grained, light pink to medium grey granitoid, which is locally porphyritic. Plagioclase phenocrysts up to 5 mm in size exhibit a random orientation. A primary igneous foliation was not observed. Narrow contact zones mapped by Mackasey (1976) and Mackasey and Wallace (1978) consist of hybrid intrusive rocks of dioritic composition intruded as dikes and tongues into the volcanics. They also reported granitoid rocks engulfing large volcanic inclusions.

Subhedral plagioclase grains of oligoclase-andesine
Fig. 14. Fragments of mafic volcanic breccia are poorly sorted, loosely packed and angular to subrounded in shape.

Fig. 15. Anhedral quartz grains interstitial to the plagioclase grains in trondhjemitic stock (field of view ≈ 7 mm).
composition average 2 mm in size and comprise up to 60 % of the intrusion. Albitization, saussuritization and sericitization is common and may completely alter plagioclase grains. Approximately 30 % quartz, averaging 1 mm in size, is present as anhedral grains interstitial to the plagioclase (Fig. 15). Up to 10 % dark green pleochroic hornblende approximately 0.75 mm in size and a minor amount of biotite was also found to be interstitial. Potassic feldspar was not observed, however, it was reported in some of the samples analyzed by Mackasey and Wallace (1978). Chlorite and epidote are common alteration products and acicular apatite grains are accessory. The dominant mineralogy is that of a trondhjemite or granodiorite composition which was confirmed by whole-rock chemical analysis by Mackasey and Wallace (1978).

II.IV INTRUSIVE ROCKS PRESENT IN BGB AND OTT

II.IVa GABBRO

Gabbroic lenses are the dominant mafic intrusive bodies. Although intrusive contacts were not observed, excepting a sheared contact described below, these units were interpreted to be gabbros based on their grain size, texture and degree of deformation. Gabbro outcrops as elongate bodies trending parallel to the foliation. Lengths measure hundreds of metres, while widths are typically tens of metres. They are generally massive, homogeneous equigranular bodies with a medium to coarse grain size. Weathered gabbro is dull brown in colour
and often has a rough, knobby surface. Fresh surfaces may exhibit a salt and pepper texture. Other mafic intrusions south of the study area are of diorite and quartz-diorite composition (Mackasey 1976; Lawson 1986).

Within the PLDZ the gabbroic intrusions retain an undeformed centre and deformation is accommodated in sheared margins (Fig. 16). Figure 17 shows a sheared margin of a gabbro only centimetres away from an undeformed segment of the same body as seen in Figure 18. Closer to the PLDZ centre, shear fabrics and kinkbands were observed in the same outcrop.

Subhedral, lath-shaped plagioclase grains comprise approximately 60% of the gabbro. Plagioclase has been extensively altered to saussurite, sericite, and albite, but retains its original shape. Plagioclase composition is deduced to be labradorite where albite twins are preserved. Average grain size of plagioclase grains is 0.4 mm. Remnant clinopyroxene grains are preserved in coarse-grained bodies and constitute up to 30% of the rock. Typically, clinopyroxene and hornblende are altered to actinolite. Average grain size of amphibole grains is 2 mm to 5 mm, but grains up to 8 mm were measured in the field. Amphibole typically contains plagioclase in a subophitic texture.

A micrographic texture consisting of an intergrowth of quartz and feldspar makes up a minor component of the gabbro. Fine grained alteration minerals include aggregates of flakes of chlorite, epidote and zoisite. Quartz, leucoxene and pyrite
Fig. 16. Gabbroic body retaining an undeformed centre (right) as deformation is accommodated in sheared margins (left).
Fig. 17. Sheared margin of a gabbro (field of view ≈ 7mm).

Fig. 18. Undeformed centre of a gabbro (field of view ≈ 7mm).
are also present.

II.IVb LAMPROPHYRE DIKES

An amphibole porphyry, interpreted as a lamprophyre, outcrops on a small peninsula on the south shore of Paint Lake. Weathered surfaces are medium green to grey and fresh surfaces are dark green in colour. Contacts with the host volcanics were not exposed but the width of the dike is estimated at a few metres. Long euhedral to subhedral lath-shaped amphibole (hornblende) grains measure up to 5 mm in length and comprise approximately 40% of the rock. The groundmass consists of a fine-grained (50 µm) mosaic of subhedral andesine grains which exhibit albite twinning. Alteration is minimal involving saussuritization and sericitization of plagioclase grains. Minor amounts of magnetite, apatite, chlorite and carbonate are present. Large altered phenocrysts of chlorite composition rimmed with fibrous quartz probably represent either altered pyroxene or hornblende or perhaps represent amygdules (Fig. 19).

Although only one lamprophyre dike was identified in the map area, more are believed to be present as they are difficult to recognize in the field when contacts are not observed, especially where hosted by mafic volcanics. Mackasey and Wallace (1978) have identified lamprophyres and other mafic dikes north of the study area. This rock is believed to be relatively young because it has suffered minimal alteration
Fig. 19. Large altered phenocrysts of chlorite composition rimmed with fibrous quartz present in lamprophyre (field of view \( \approx 7\text{mm} \)).

Fig. 20. Felsic porphyry displaying concordant contact with foliation.
and deformation.

II.IVc FELSIC PORPHYRIES

Feldspar and quartz-feldspar porphyries are found as medium-coarse grained dikes/sills throughout the map area. Weathered surfaces are white to buff and fresh surfaces are light grey in colour. Widths of the intrusives are generally less than one metre and exposed lengths measure tens of metres. Contacts with host rocks may be baked. A 40 cm wide felsic porphyry outcropping on the north shore of Paint Lake is concordant with the foliation of the host volcanics (Fig. 20).

Plagioclase phenocrysts, generally oligoclase, comprise approximately 40% of the porphyries, average 3 mm in size but have been measured up to 8 mm. Grains are euhedral to subhedral, typically kinked and twinned due to deformation and only weakly sericitized (Fig. 21). Subhedral to anhedral quartz grains make up approximately 5% of the rock and average 0.4 mm in size. Other phenocrysts include amphibole (hornblende) and biotite. A microcrystalline matrix (<25 μm) comprises an estimated 60% of the porphyry and is quartz-feldspathic in composition. A minor amount of chlorite, carbonate, apatite and disseminated pyrite is present.

Felsic porphyries may represent apophyses of the larger trondhjemitic stocks in the area, however, a lack of significant alteration and deformation suggests that this rock
Fig. 21. Kinked plagioclase grains in felsic porphyry (field of view ≈ 7mm).

Fig. 22. Typical diabasic texture (field of view ≈ 7mm).
type may be the granophyric equivalent of late diabase dikes (Mackasey 1976).

II.IVd DIABASE DIKES

North-trending diabase dikes intrude all rocks within the map area. Although contacts were not observed, dike widths are estimated at metres to tens of metres and magnetic maps indicate strike lengths of several kilometres. Weathered surfaces have a dull to dark brown colour, while fresh surfaces exhibit a massive, medium-grained, equigranular, salt and pepper texture. Outcrops exhibit a blocky appearance believed to have been produced by two fracture sets. The diabase dikes show no signs of internal deformation, however they have been displaced by the Paint Lake Fault (Mackasey 1976; Stott 1984a,b).

Randomly orientated euhedral to subhedral lath-shaped plagioclase grains of labradorite constitute an estimated 50% of the rock. Grain size is typically 1 mm to 2 mm. Plagioclase grains are only weakly sericitized and relatively fresh grains exhibit albite twinning. Approximately 40% of the diabase is clinopyroxene of similar grain size as the plagioclase and has undergone limited alteration to actinolite and chlorite. Accessory minerals include magnetite, apatite and micrographic intergrowths of quartz and feldspar. A typical diabasic to sub-diabasic texture is well preserved (Fig. 22).

Diabase dikes in the region have been dated by Wanless
(1970) as Middle to Late Precambrian age. Age determinations indicate at least two ages of 1545-1569 Ma and 1125 Ma.
CHAPTER III - STRUCTURES ASSOCIATED WITH PLDZ DEVELOPMENT

The PLDZ exhibits a wide assemblage of structures including foliations, lineations, folds, kinkbands, tension gashes and step fractures. This chapter will demonstrate that the dominant deformation mechanism was one of predominantly ductile, simple shear (PLSZ) followed by a later brittle fault movement (PLF). Reference should be made to Map 2 - Foliations and Map 3 - Minor Structures for distribution and orientation of these structures within the PLDZ.

III.I FOLIATIONS

Three main types of foliations are associated with the development of the PLSZ. These structures are expressed in the field as a succession of deformation stages commencing with weakly to moderately deformed rocks near the shear zone boundary, yielding to strongly to intensely deformed rocks in the shear zone centre with progressive simple shear deformation. S-fabric, C-fabric, and C'-fabric foliations, respectively represent an observed gradient of low shear strain in the margins and high shear strain in the centre of the PLDZ (Fig. 23). A fourth type of foliation identified as a conjugate cleavage set is observed only in the eastern boundary of the map area and suggests a localized component of pure shear in this region.
Fig. 23. Foliations associated with shear zone development in the PLDZ; a = S-fabric, b = C-fabric, c = CS-fabric, d = C'-fabric (scale may represent metres to 100's of metres).
III.Ia S-FABRIC

During initial stages of a deformation a plane of preferred mineral orientation develops defining a schistosity. This schistosity has been termed an S-fabric (schistosité) by Berthé et al. (1979) and lies within the XY-plane of the finite strain ellipsoid (Fig. 23a). On the south side of the PLDZ the S-fabric exhibits a sigmoidal shape on a macroscopic scale as it rotates from the shear zone boundary into parallelism with the shear zone centre. The angular relationship between the S-fabric and the trend of the shear zone is given the value $\theta$ (Ramsay and Graham 1970). On the north side of the PLDZ a preexisting foliation (believed to have been produced during a pre-PLDZ shear deformational event) has been deflected during the formation of the S-fabric. Thus, a realignment of the preferred mineral orientation of the older foliation to coincide with a new tectonic regime also produces a sigmoidal geometry as the value of $\theta$ decreases with increasing strain. The axial surface of asymmetric folds and the alignment of some conglomeratic clasts parallel the S-fabric (Fig. 24).

III.Ib C-FABRIC

As strain increases from a shear zone boundary towards its centre, discrete planes of relative movement are formed. Lithologies cut by this foliation typically display a mylonitic layering (Fig. 25). Such foliations have been
Fig. 24. Conglomeratic clasts aligned parallel to S-fabric with pencil parallel to C-fabric.

Fig. 25. Mylonitic layering defines C-fabric in conglomerate.
defined by Berthé et al. (1979) as a C-fabric (cisaillement), which they believe to have developed as a result of activation of the S-fabric as the angle \((\theta)\) between the S- and C-fabric diminishes. Platt (1984) argues that the formation of the C-fabric simply reflects a change in the orientation of the finite strain axes with increasing shear strain and the C-fabric has now become the local XY-plane of the finite strain ellipsoid (Fig. 23b).

The S-fabric is believed to have been formed before the C-fabric based on the following field observations:

1) The S-fabric is observed in weakly to moderately deformed rocks near the shear zone boundary and the C-fabric is present in strongly to intensely deformed rocks near the centre of the shear zone.

2) Nowhere has the S-fabric and the C-fabric been found in the same outcrop. However, microscopic analysis reveals a weak S-fabric in some moderately deformed rocks which possess a C-fabric. The S-fabric is preserved as the alignment of the long axes of plagioclase grains oblique to the C-fabric, yet the C-fabric does not cross-cut or displace the S-fabric.

3) An increase in the density of the C-fabric approaching the shear zone centre can be explained by the rotation of the S-fabric into parallelism with the C-fabric. Therefore, the C-fabric grows at the expense of the S-fabric or by gradual increase in the number of discrete planes. The S- and C-fabric cannot be distinguished at this stage, similar to the
conditions which led Berthé et al. (1979) to coin the term CS-fabric (Fig. 23c). This foliation plane behaves as a C-fabric and will be referred to as such.

Fig. 26 is a plot of the poles to S- and C-fabric planes. The mean foliation derived from both S- and C-fabric measurements is 077/86, however, it is the construction of foliation trajectories, shown in Fig. 27, which best displays the foliation geometry. The PLDZ exhibits an anastomosing pattern which can be traced through the study area. This is illustrated on the stereonet in Fig. 26 as the strike of the PLDZ varies from 075 to 105.

III.Ic C'-FABRIC

Within the innermost portion of the PLDZ, approximately 100 m - 200 m from the shear zone centre, there exists a crosscutting foliation which lies oblique to the C-fabric. This foliation is manifested as a single set of small ductile shear zones which has a relatively consistent orientation (clockwise) with respect to the trend of the PLDZ. White (1979) called these structures shear bands, Platt (1979) named them extensional crenulation cleavages, and Ponce de Leon and Choukroune (1980) preferred the nomenclature C'-fabric (Fig. 23c).

The C'-fabric overprints the C-fabric and the apparent sense of displacement on the C'-fabric is always such to cause extension along the C-fabric (Platt and Vissers 1980).
Fig. 26. Stereonet plot of poles to planes for both S-fabric and C-fabric (mean orientation= 077/86).

Fig. 27. Foliation trajectories constructed from both S-fabric and C-fabric.
Carbonate infilling along the C'-fabric is indicative of a component of dilation. Preferential weathering of the carbonate material leaves the C'-fabric as a foliation of negative relief in the field (Fig. 28). The spacing between individual C' foliation planes ranges from millimetre to centimetre scale, averaging 3-4 cm in width.

It is evident from the trend of the C'-fabric that its formation is a product of shortening at a high angle to the C-fabric (White et al. 1980; Platt and Vissers 1980). The orientation of the C'-fabric is postulated to be controlled by the bulk shear direction (Platt and Vissers 1980) and/or the attitude of the C-fabric (Harris and Cobbold 1984). The C'-fabric developed during the same deformation that produced the C-fabric (White et al. 1980) and C'-fabric generation reflects the final stage of ductile deformation in a shear zone before entering the field of dominant brittle deformation (Gapais and White 1982; Passchier 1984). The absence of conjugate and multiple sets of C'-fabric implies non-coaxial deformation produced by a bulk simple shear mechanism. Hence, the sense of shear along a single C'-fabric plane is the same as that of the main shear zone and can be used to deduce the overall sense of shear (Simpson and Schmid 1983; Weijermars and Rondeel 1984). Poles to C'-fabric planes, plotted in Fig. 29, reveal a mean orientation of 279/69.
Fig. 28. C'-fabric in intermediate volcanic exhibits negative relief due to preferential weathering of carbonate infilling.

Fig. 29. Stereonet plot of poles to planes for C'-fabric (mean orientation = 270/69).
III.Id CONJUGATE CLEAVAGE SET

A conjugate cleavage set outcrops in four localities in the eastern boundary of the map area. The orientations of the conjugate set planes average 050/90 and 097/90. Conjugate foliation planes are preferentially weathered and exhibit a negative relief. No displacement was observed along either cleavage plane and they appear to have developed as a fracture set. The conjugate set displays an anastomosing geometry with relatively undeformed lithons measuring approximately 8 x 3 cm bounded by the cleavage planes (Fig. 30).

The presence of a symmetric conjugate cleavage set is an anomalous structure in a bulk ductile simple shear environment (Platt and Vissers 1980). The spatial and temporal relationship between this foliation and the PLDZ is uncertain because the exact position PLDZ and its eastern extent is problematic, however, the conjugate cleavage planes are symmetrical about the C-fabric. C'-fabric which is ubiquitous throughout the study area is not found in this region. Correlation of the PLDZ with the Barton Bay Deformation Zone to the southeast requires detailed mapping which is hindered by overburden. Therefore, it is possible that the PLDZ terminates, perhaps through splays of the PLF, in the eastern limits of the study area (see Map 2 - Foliations). A conjugate cleavage set may have been produced in the waning stages of deformation in which the eastern tip(s) of the PLF was subjected to a component of pure shear.
Fig. 30. Conjugate cleavage set displaying an anastomosing geometry with relatively undeformed lithons bounded by the cleavage planes.
III.II LINEATIONS

Four main types of lineations were mapped in the study area: 1) stretching, 2) slickensides and sickenfibres, 3) intersection, and 4) mineral elongation.

III.IIa STRETCHING LINEATION

Within the central portion of the PLDZ a pervasive, statistically subhorizontal lineation corresponds to the projection on the C-fabric (XY-plane) of a stretching direction related to minerals deformed in the S-fabric which have rotated into the C-fabric. The stretching lineation parallels the direction of movement. Metamorphism and deformation has altered volcanic and conglomerate primary mineralogical assemblages to a very fine-grained assemblage of chlorite, white mica, carbonate, and quartzo-feldspathic material. Such an assemblage is not conducive to developing a well defined lineation, thus, a weak stretching lineation is observed despite evidence for a strong component of subhorizontal shear. Only clasts of the conglomerate exhibit a strong stretching lineation. The lineation:foliation (L:S) ratio is generally L<<S in the PLDZ. L>S, L>S, L=S, L<S, L<<S refer to visual estimates of the relative prominence to the stretching lineation (L) in comparison with the foliation (S) (Schwerdtner et al. 1977).

Figure 31 is a plot of stretching lineations. Values average 04/248, however, lineations were recorded plunging to
the west and to the east. Although the regional trend of the PLDZ is east-west, lineations depict a predominantly ENE trend. This may be explained by the anastomosing geometry of the PLDZ and the abundance of data obtained along Paint Lake where the PLDZ strikes ENE.

In a vertical to subvertical shear zone such as the PLDZ, a subhorizontal stretching lineation demonstrates that strike-slip movement prevailed, thus defining a transcurrent or wrench shear zone. Similar shear zones with subhorizontal stretching lineations were mapped by Platt et al. (1978), Berthé et al. (1979), Ponce de Leon and Choukroune (1980) and Jegouzo (1980).

III.IIb SLICKENSIDES AND SLICKENFIBRES

Slickensides are polished and commonly striated shear surfaces (Fleuty 1975) on the same surfaces and generally parallel to the stretching lineation. These surfaces are planes of relative movement in which the striations or grooves parallel the subhorizontal direction of movement. Unlike stretching lineations, slickensides are not a penetrative feature. Quartz or carbonate growth fibres associated with slickenside surfaces result from crystal growth and not abrasion. The long axes of these minerals are oriented subhorizontally. Wise et al. (1984) labelled these fibrous growths slickenfibres. Interpretation of movement from slickenfibres is analogous to that of slickensides.
Fig. 31. Stereonet plot of stretching lineations (mean orientation= 04/248).

Fig. 32. Stereonet plot of mineral elongation lineation (mean orientation= 73/248).
III.IIc INTERSECTION LINEATION

The lineation produced by the intersection of the C-fabric with the C'-fabric measured in the XY-plane of the C-fabric plunges 50 to 70 degrees to the east. The C'-fabric is typically expressed in the field as a foliation of negative relief due to preferential weathering of carbonate infilling, the intersection of the C'-fabric on the C-fabric results in a lineation of negative relief. The oblique trend of this east plunging lineation in the C-plane is produced because the dip of the C'-fabric (approximately 70 degrees north) is less than the dip of the C-fabric (typically vertical). This relationship is significant and will be addressed in the section on movement along the PLDZ.

III.IId MINERAL ELONGATION

A mineral elongation lineation is present only in the OTT and is not observed in the BGB. It is defined by the preferred elongation of plagioclase, actinolite and chlorite minerals in the volcanics. A stereonet plot of this lineation shown in Figure 32 reveals an average orientation of 73/248. Formation of the mineral elongation lineation is presumed to be a result of an earlier pre-shear deformation because it is found only in the OTT and does not extend across the PLDZ into the BGB. It plunges steeper than the stretching lineation of the PLDZ and is better developed than the stretching lineation, displaying LiS.
Within the PLDZ, a rotation of the mineral elongation lineation into parallelism with the shear direction (X) was not observed. No consistent trend in the plunge of this lineation was displayed approaching the shear zone centre. Detailed mapping is required further north of the PLDZ to better understand the extent of this lineation and the tectonic environment responsible for its development.

III.III FOLDING

Folding within the PLDZ is restricted to the central portion of the shear zone, approximately 200 m from shear zone centre, where a well developed C-fabric is present. C-fabric foliation planes are a prerequisite for folding. Folds are mesoscopic in scale and are presumed to have developed during one deformation event dominated by progressive simple shear. Two main types of folding were identified in the PLDZ: 1) asymmetric folds, 2) kink folds. Although the C-fabric typically exhibits a wavy or undulatory geometry it can not be proven whether this curviplanar foliation was produced during the shear process or was a product of gentle, low amplitude folding representing a transition from ductile asymmetric folding to brittle kink folding.

III.IIIa ASYMMETRIC FOLDS

Asymmetric folds are small in size and few in number. Fold amplitudes and wavelengths are centimetre scale,
interlimb angles measure tight to close. Asymmetric folds, as seen Fig. 33, exhibit a Z-shaped geometry indicative of dextral shear (Berthé and Brun 1980; Cobbold and Quinquis 1980). Asymmetric folds are present in conglomerate, intermediate and mafic volcanic units. Within the conglomerate, competent boulder-size clasts of granitoid composition serve as a nucleation site from which an asymmetric fold forms during shear along the C-fabric (Fig. 34). Fold amplitude dies out along the axial plane. Less competent elongate clasts, typically of volcanic composition, behave similarly to the ductile matrix and are folded.

Fold axes were difficult to measure because of smooth, flat outcrop surfaces. Consequently, any relationship between the fold axes and the stretching direction in the PLDZ could not be documented. Average orientation of the traces of axial planes is 039 as shown on the rose diagram in Figure 35. Axial planes are subparallel to the orientation of the S-fabric and theoretically, behave in a similar manner with progressive shear, rotating towards the C-fabric. Because of the very limited number of asymmetric folds observed, such a rotation can not be statistically proven.

Where asymmetric folds are observed, C-fabric is always folded. However, the C'-fabric was not found to be folded. Therefore, asymmetric folding can be bracketed to have developed post-C-fabric and pre-C'-fabric. This type of folding is labelled "active" folding as it is related to the
Fig. 33. Z-fold in mafic volcanic (above pencil).

Fig. 34. Nucleation of Z-fold around competent granitoid clast in conglomerate.
Fig. 35. Rose diagram illustrating trend of axial traces for Z-folds (mean orientation (r.v.) = 0.39; probability that distribution is random (p) = 0.00014; interval = 10).
Asymmetric folding is best developed when competency contrast between layers is high (Johnson 1979) or when local heterogeneities, such as clasts, are present within host lithologies (Berthé and Brun 1980). The absence of significant competency contrast between layers in each lithology and the absence of exposed lithological contacts between different lithologies explains the low number of asymmetric folds observed in the PLDZ. Isoclinal and sheath folds, which are characteristic of high strain shear zones, were not found for the same reasons.

**III.IIIb KINK FOLDING**

Single asymmetric kinkbands prevail in the PLDZ, however, only four conjugate asymmetric sets were mapped (Fig. 36). Kinkbands were found in conglomerate, gabbro, intermediate and mafic volcanics. Like asymmetric folds, kinkbands are not lithologically controlled but are dependent on their proximity to the shear zone centre where a preexisting anisotropy (C-fabric) is present. Measurements recorded in the field from kinkbands included the orientation of the kinkband boundary, the width of the kinkband and the acute angle between the kinkband boundary and the C-fabric. These measurements are listed in Table 1.

From Table 1 it is evident that single asymmetric kinkbands are statistically similar to asymmetric kinkbands
Fig. 36. Conjugate set of kinkbands.

Fig. 37. A stereonet plot of poles to kinkband boundary planes displays a regional conjugate set.
Table 1. Field measurements of kinkbands.

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</tr>
<tr>
<td>number of observations</td>
<td>4 4</td>
<td>53 20</td>
</tr>
<tr>
<td>orientation of kinkband boundary</td>
<td>122 (101-131) 022 (015-025)</td>
<td>121 (+- 16) 028 (+- 17)</td>
</tr>
<tr>
<td>acute angle between kinkband boundary and C-fabric</td>
<td>47 (36-59) 53 (42-66)</td>
<td>52 (+- 16) 49 (+- 17)</td>
</tr>
<tr>
<td>width of kinkband (cm)</td>
<td>3 (2-5) 3 (2-4)</td>
<td>4 (+- 2) 3 (+- 3)</td>
</tr>
<tr>
<td>host lithology: conglomerate</td>
<td>0 0</td>
<td>17 2</td>
</tr>
<tr>
<td>intermediate volc.</td>
<td>1 1</td>
<td>18 6</td>
</tr>
<tr>
<td>mafic volc.</td>
<td>2 2</td>
<td>17 12</td>
</tr>
<tr>
<td>gabbro</td>
<td>1 1</td>
<td>1 0</td>
</tr>
</tbody>
</table>
observed in conjugate sets. Data for conjugate sets are presented as a range of measurements because of few data points, while data for single kinkbands are provided with standard deviations of measurements. A stereonet plot of poles to kinkband boundary planes (Fig. 37) displays a conjugate set which has developed on a regional scale. Sinistral kinkbands have an a mean orientation of 124/90 and dextral kinkbands average 020/90. Kinkband boundaries dip vertical to subvertical and where a measurement could not be obtained from an outcrop it was assumed to be vertical.


Many geologists, including Anderson (1964), Paterson and Weiss (1966), Donath (1968), Gay and Weiss (1974), have experimentally shown that conjugate kinkbands develop when maximum compression (σ1) parallels the layering. In theory, σ1 bisects the obtuse angle between kinkband boundaries for conjugate sets. Wallace and Clifford (1983) stated that kink folding is not related solely to layer parallel shortening (pure shear), but possibly to a component of simple shear as well. The significance of simple shear in producing kinkbands
and any interpretation of a stress regime from these structures is discussed below.

Pure shear experiments by Gay and Weiss (1974) report that conjugate kinkbands develop if the layering is inclined at angles less than 5 degrees to \( \sigma_1 \). Single kinkbands form if the angle is between 5 degrees and 30 degrees. With an increase in the angle between the layering and \( \sigma_1 \), one set of conjugate kinkbands is reduced until layer parallel slip increases and no kinkbands form.

Reches and Johnson (1976) subjected layered material to layer parallel pure and simple shear and have shown that asymmetric sinistral kinkbands prevail during dextral shear and dextral asymmetric kinkbands dominate during sinistral shear. They state that the formation of symmetric conjugate sets or single asymmetric kinkbands is dependent on the inclination of the direction of maximum compression to layering, similar to Gay and Weiss (1974).

Experimental studies by Hoeppener and Schwarz (1980) involved only simple shear of layered material. Their results contrast with earlier studies of pure shear and pure plus simple shear. When layering and \( \sigma_1 \) were parallel, no kinkbands formed and only slip along layering was observed. An angle between 5 degrees and 30 degrees produced kinkbands antithetic to shearing, similar to pure shear experiments. Conjugate kinkbands were formed when layering was inclined 45 degrees to layering and angles between 50 degrees and 70 degrees produced
synthetic kinkbands. Hoeppener et al. (1983) go as far to say that the orientation of the antithetic kinkbands parallel \( \sigma_1 \).

Any interpretation of stress fields from the orientation of kinkbands must be made carefully and full consideration must be given to the dominant deformation mechanism which has produced the kinkbands. The determination of \( \sigma_1 \) in a pure shear environment differs from a simple shear environment. Given the field data from the PLDZ (Fig. 37), \( \sigma_1 \) is equal to 072/00 if pure shear or 117/00 if simple shear. Minor structures within the PLDZ indicate that simple shear was the primary deformation mechanism. However, the presence of synthetic and conjugate kinkbands in addition to the dominant antithetic kinkbands makes interpretation of \( \sigma_1 \) complex. At best, it can only be implied that \( \sigma_1 \) is orientated approximately northwest to southeast and is a subhorizontal stress. This is manifested as dextral shear transcurrent movement along the PLDZ.

A transition from a dominantly ductile environment to a dominantly brittle environment is required to produce kinkbands. According to Johnson (1979), the contact strength between layers is an important factor in determining if ductile or brittle folds will develop. He defines contact strength as a combination of friction, cohesion and pore fluid pressure between layers. During ductile deformation contact strength is minimal, permitting easy slip between layers and favouring the formation of asymmetric folds. As the contact
strength increases, the ease with which slip occurs between layers decreases. During brittle deformation contact strength is high, thus inhibiting slip between layers. Kink folding is favoured at this stage of deformation. Honea and Johnson (1976) illustrated that a preexisting anisotropy first develops an undulatory geometry before it is actually kinked. Hence, transition from a ductile environment to a brittle environment must be accompanied by an increase in friction between layers, an increase in cohesion between minerals and a gradual decrease in pore fluid pressure. A change in these parameters is probably due to continual uplift during progressive deformation or sudden localized increases in pore fluid pressure which will result in hydrofracturing (Stott and Schneider 1983; Henderson 1983).

Nowhere were kinkbands observed to be overprinted by another structure and they may represent the youngest structure associated with transcurrent movement. Similar spatial and temporal relationships between kinkbands and deformation zones are present throughout the Superior Province. Williams (in press) explained this relationship in terms of stick-slip movement along fault planes. During the waning stages of deformation, stick on the fault plane results in a localized compressional regime on the leading edge of the hanging wall and a localized extensional regime on the trailing edge. Kinkbands are produced during compression, while extensional structures such as C'-fabric and tension
gashes develop during extension (Williams, in press). Hanmer (1979) also recognized a spatial and temporal relationship between C'-fabric and kinkbands.

III.IV TENSION GASHES

III.IVa EN ECHELON VEIN ARRAYS

Tension gashes or extension fissures within the PLDZ vary in length from 4 cm to 75 cm, averaging approximately 40 cm. Widths range from 0.5 cm to 4 cm, averaging about 3 cm. Where measurable, tension gashes dip steeply to vertically. Quartz and less commonly carbonate material infills the tension gashes. These extensional structures are found in all major lithologies, typically in an en echelon array. Conjugate sets were not observed.

A rose diagram representing the trend of individual tension gashes in Figure. 38 demonstrates that although many of these structures are orientated roughly north-south, the probability that data represent a uniform distribution is 0.16. The probability that vein arrays, as shown in Figure 39, represent a random distribution is higher (0.68). Most of the tension gashes are undeformed or exhibit a sigmoidal geometry indicative of dextral shear (Fig. 40). However, few exhibit a sinistral sigmoidal geometry.

The majority of tension gashes were observed outside the central portion of the PLDZ, where C- and C'-fabric are absent and only an S-fabric, if any fabric, is present. One of the
Fig. 38. Rose diagram representing the trend of individual tension gashes (mean orientation (r.v.)= 357; probability that distribution is random (p)= 0.16; interval= 30).

Fig. 39. Rose diagram representing trend of vein arrays containing tension gashes (mean orientation (r.v.)= 093; probability that distribution is random= 0.69; interval= 20).
Fig. 40. Tension gashes exhibiting a sigmoidal geometry indicative of dextral shear.

Fig. 41. Pull-apart structure in granitoid clast of conglomerate.
very few tension gashes which was observed in the central portion of the PLDZ is infilled with carbonate material which has crystallized with its long axes parallel to the maximum elongation direction.

According to Ramsay (1967) tension gashes may develop perpendicular to the maximum elongation at an angle of 45 degrees to the shearing surfaces. This geometry is well illustrated by Beach (1975) and Ramsay and Huber (1983). Nicholson and Ejiofor (1987) agree with the initiation of tension gashes parallel to the maximum compression direction but have shown that these sigmoidal structures need not be formed through ductile, simple shear parallel to the vein arrays. They believe that displacement parallel to the edges of echelon arrays was instead a consequence of dilation. Pollard et al. (1982) have also produced experimental sigmoidal arrays without shear.

Tension gashes are not a common minor structure associated with the development of the PLDZ. Their presence in the less deformed portion of the PLDZ contrasts with the spatial relationship of many of the other minor structures which are concentrated in the highly deformed central portion where a strong anisotropy has developed. Either tension gashes are not well preserved where C- and C'-fabric dominate or they were simply not produced during or after the most intense stages of deformation. Alternatively, tension gashes may have been produced in the absence of any remote shear component.
This may explain the lack of consistency in orientations and conflicting sigmoidal geometry. Such randomness may be also be explained by the presence of fractures existing in the rock before simple shear, which may be opened during shearing and their orientations are different from that produced during simple shear (Ramsay 1967).

Tension gashes represent an extensional structure which may have formed at various stages throughout the development of the PLDZ or perhaps during a period when ductile simple shear was not active. Thus, the timing of these structures may represent many generations from early to late in the deformational history of the PLDZ. One explanation involves stick-slip movement (Williams, in press).

### III.IVb PULL-APART STRUCTURES IN CLASTS

Pull-apart structures are defined as quartz-filled fractures which are present in many of the gravel-size conglomeratic clasts. Only the competent granitoid clasts exhibit this structure and a strong anisotropy is a prerequisite in the host conglomerate. The walls of these fractures are perpendicular to the maximum extension direction and individual fibrous quartz grains are orientated parallel to this direction (Fig. 41). The fractures typically extend the width of the clast but do not propagate into the matrix. The more competent clasts, sitting in a ductile matrix, undergo some elongation then yield by fracturing during
extension parallel to the shear direction producing this pull-apart structure.

II.V STEP FRACTURES

Slickenside surfaces may contain small (centimetre scale) step fractures oriented normal to the subhorizontal striations with the steep sides of the steps (risers) facing in one direction. These step fractures were observed in only a few outcrops on the southern wall of the PLF consistently facing west which is the direction of movement of the opposite (north) block as interpreted from other kinematic indicators. Norris and Barron (1969) report that step fractures originated in at least one of three ways:

1) as irregularities in the surface due to original fracture
2) due to plucking of plates from the slip surface
3) due to faults intersecting the slip surface

Early textbooks by Hills (1940), Nevin (1949) and Billings (1954) state that the steps faced in the direction of movement of the opposite block. Later experimental and field studies by Paterson (1958), Tjia (1964), Rieker (1965) and Gay (1970) has shown that the steps oppose the movement of the opposite block. This prompted Hobbs, Means, and Williams (1976) to conclude that there is no general rule concerning the kinematic significance of the steps. However, detailed studies by Norris and Barron (1969) in the field and in the lab have demonstrated that step fractures do not necessarily
face in the direction of movement of the opposite block, as inferred from regional structures, but step fractures with this orientation should predominate.
CHAPTER IV - STRAIN ANALYSIS

IV. I STRAIN DOMAINS

The presence or absence of a foliation serves as an excellent field tool which may be used to estimate the relative degree of deformation in sheared rocks. Progressive deformation during a bulk ductile simple shear event furnished the PLDZ with a succession of foliations which serve as the basis for classifying the shear zone into distinct strain domains as follows:

Domain 1 = weak deformation
  = S-fabric present

Domain 2 = moderate deformation
  = S-fabric present
  = C-fabric present

Domain 3 = strong deformation
  = C-fabric present

Domain 4 = intense deformation
  = C-fabric present
  = C'-fabric present

Domain 5 = highest degree of deformation
  = C-fabric present
  = C'-fabric present
  = decrease in $\beta$ measurements

The difference between Domain 4 and Domain 5 is based on the angular relationship ($\beta$) between the C-fabric and the C '-fabric. Field measurements on an outcrop scale reveal that some C'-rich zones have a smaller $\beta$ value than others. Domain 4 angular measurements average 37 degrees (S.D. = 4.4), while Domain 5 measurements average 27 degrees (S.D. = 4.8). These domains with a smaller $\beta$ value are much more fissile than
those with a higher β value and their original lithology is most difficult to ascertain. Such domains have developed with increasing shear strain (White et al. 1980) and depict the highest degree of ductile deformation in the PLDZ. The presence of a weak S-fabric in Domain 2 was identified in thin section and the boundaries between Domains 1 and 2 and between Domains 2 and 3 may be gradational as they could not be delineated in the field. Conversely, the boundaries between Domains 3, 4 and 5 are discrete. Domain 5 is less common than the others. Fig. 42 is a schematic sketch illustrating strain domains present in the PLDZ.

Each strain domain is observed on both sides of the PLDZ in conglomerate, intermediate and mafic volcanic rocks, although they are best exposed on the north side. Because of the observed increase in the degree of deformation approaching the shear zone centre, primary textures and mineral assemblages are obliterated by shear fabrics. Assuming deviatoric stress, confining pressure, pore fluid pressure, temperature and grain size were equal in each domain, the strain domains are believed to have been formed in originally homogeneous lithological units. These assumptions are difficult to confirm, especially the restriction of grain size. Any primary variation in grain size is obscured due to grain size reduction which accompanies an increase in deformation. However, the presence of equally well developed shear fabrics in conglomerate, intermediate and mafic volcanic
Fig. 42. Schematic sketch illustrating the five strain domains observed in the PLDZ (scale varies from metres to 100's of metres).
units suggests that the development of successive strain domains is not lithologically controlled, but rather is dependent on the proximity of the rock type to the shear zone centre. Their development could therefore be a function of strain rate during a bulk simple shear deformation.

IV.II STRAIN PROFILES

Various methods have been used to determine shear strains and total displacement in large scale shear zones. These include the reorientation of regional scale dike swarms (Escher et al. 1975), the statistical change of fold axial planes and axes (Bak et al. 1975), the statistical change in the interlimb angles of folds (Grocott and Watterson 1980), and the calculation of principal strains at individual exposures to give the total displacement in a large shear zone (Mitra 1979).

Each of the these methods of strain profiling requires some special circumstance. The Ramsay and Graham (1970) model has been applied successfully to numerous mesoscopic shear zones and may be used as a general application for all shear zones, but as Naruk (1986) reported, attempts to apply the model quantitatively to kilometre scale zones are rare (Coward et al. 1973) and generally unsuccessful.

Ramsay and Graham (1970) formulated the following equations to calculate shear strain and total shear displacement, respectively:
\[ \gamma = \cot \alpha' - \cot \alpha \]  
Eqn.1

\[ \gamma = \text{shear strain} \]

\[ \alpha = \text{initial angle with } x\text{-direction before deformation} \]

\[ \alpha' = \text{angle with } x\text{-direction after deformation} \]

\[ S = \int_{-d}^{d} \gamma \, dx \]  
Eqn.2

\[ S = \text{total shear displacement} \]

\[ d = \text{distance (width)} \]

The location of three primary strain profiles across the PLDZ is given in Fig. 43. Shear strain / distance graphs were constructed using Equations 1 and 2 for the three sections (Fig. 44). Measurements of the shear zone width range from 0.96 km to 1.37 km, calculations of shear strain values exceed 55, and total shear displacement varies from 30 km to 51 km. Point distribution for each strain profile is presented in Appendix I and shows an increase in shear strain toward the shear zone centre, similar to strain profiles of classical shear zones (Ramsay and Graham 1970). However, the distribution of points within the centre of the strain profile deviates from classical shear zones where strain / distance graphs exhibit point distribution only in the margins of the strain profile. Foliations from profile C-D are shown in Fig. 45, because this section perhaps best portrays the foliation geometry and hence, best represents shear strain values and the total amount of shear displacement in the PLDZ.
Fig. 43. Location of three primary strain profiles across the PLDZ.

Fig. 44. Shear strain / distance graphs constructed from sections AB, CD and EF.
Fig. 45. Foliations from which strain profile C-D was constructed.
Many considerations must be weighed when attempting to strain profile a shear zone at this scale. Computation of shear strain and total displacement hinge on the following factors as listed in Table 2.

Large shear zones characteristically have a complex strain profile according to Grocott and Watterson (1980). They state that the strain rate varied with time and such a complexity is the result of a longer history rather than a feature of rheological or mechanical significance. Small shear zones are more easily studied and can provide information directly applicable to larger examples. Small scale shear zones typically have simple strain profiles, suggesting a single displacement event, perhaps with a constant strain rate.

A recent study of a decimetre scale shear zone by McGrath (1986) produced a shear strain profile with point distribution and shear strain values similar to profiles constructed from the PLDZ. Thus, shear strains may reach exceedingly high values with similar point distribution irrespective of the size of the shear zone. Despite the enormous differences in calculated total displacement, it is concluded that kilometre scale shear zones may be represented by a relatively simple primary strain profile similar to small shear zones. However, within the central portion of the PLDZ, illustrated in Fig. 44 as a dashed line at the top of each strain profile, there exists a region which behaves less uniformly than the primary
Table 2. Factors controlling successful strain profiling in PLDZ.

1) The deformation mechanism must be one of simple shear strain. Pure shear and volume change will distort foliation measurements. Although, mesoscopic structural analysis demonstrates a major component of simple shear, a minor amount of pure shear is implied from conjugate cleavages restricted to the eastern extent of the map area. The component of pure shear cannot be evaluated due to the absence of strain markers present during ductile deformation, but it is certain to induce an apparent increase in shear strain values and related displacement calculations.

2) An important requirement is that the shear zone be suitably exposed. Shear zone boundaries cannot be accurately demarcated because of lack of continuous outcrop, thus, the width of the shear zone can only be approximated. This uneven distribution of outcrop may well explain the modest asymmetry of the shear strain profiles about the shear zone centre. Because the centre of the shear zone lies at the bottom of Paint Lake, its position and orientation can only be approximated.

3) Error or deviation in measurement of $\theta$ in the field will substantially affect shear strain values, especially at high strain (i.e. small $\theta$ values).

4) Late Proterozoic diabase dikes are the only independent strain markers which may be used to compare displacements obtained from shear strain profiles. The dikes have a consistent dextral sense of displacement and, although correlation of the dikes across the PLDZ is difficult, displacements measure up to 5 km. This value is much less than those obtained from shear strain profiles and may be explained by different deformation processes. Because the dikes are truncated abruptly by the PLF and microscopic analysis displays undeformed textures near the truncation, it is implied that the dikes were faulted and not sheared. Hence, displacement of the dikes may represent an estimate of brittle displacement as opposed to ductile displacement.

5) Mesoscopic structures show that displacement is affected by the problems of oblique section. Although displacement appears to be dominated by strike-slip movement, there is evidence for an early dip-slip movement.

6) A significant deficiency of the strain profile is the lack of time control (Grocott and Watterson 1980). The strain rate or state is not known at any stage in the displacement history.

7) Shear strain values represent minimum displacement. Rotation of an S-fabric into parallelism with the C-fabric can be measured, yet an unknown quantity of movement along these C-fabric planes results in immeasurable shear strain and total displacement values. No method has been devised to calculate similar values from the C'-fabric.
strain profile. This allows for separation of the PLDZ into a primary and secondary strain gradient, which is discussed below.

IV.III PRIMARY AND SECONDARY STRAIN GRADIENTS

Many authors, beginning with the work of Ramsay and Graham (1970), have documented the presence of a strain gradient from shear zone boundary to shear zone centre. Such a gradient is observed within the PLDZ and is referred to as a primary strain gradient. Numerous authors including White (1979), Platt (1979, 1984), Ponce de Leon and Choukroune (1980), White et al. (1980), Platt and Vissers (1980), Gapais and White (1982), Passchier (1984), Harris and Cobbold (1984), and Weijermars and Rondeel (1984) have demonstrated that the C'-fabric generally develops within the centre of the shear zone where strain is greatest. This is also true for the PLDZ, except that the C'-fabric does not develop as a homogeneous sector, but rather as a heterogeneous sector containing alternating C'-rich and C'-poor zones which implies a secondary strain gradient.

Within the region of highest shear strain, approximately 100 m - 200 m from the shear zone centre, distinct zones rich in C'-fabric are intercalated with zones poor in C'-fabric. C'-rich zones are those which contain both a well defined C-fabric and an equally well defined C'-fabric. C'-poor zones possess a well developed C-fabric and a faint C'-fabric that
can only be detected in thin section. Such field relationships demonstrate that generation of C'-fabric is spatially heterogeneous. Contacts between these contrasting domains are discrete and zone widths range from decimetres to metres, typically 1-2 m wide. These zones are planar to curviplanar and exhibit a continuous strike trending subparallel to parallel to the PLDZ.

Figures 47 and 48 best illustrate typical shear foliations within the central portion of the PLDZ. The location of two outcrops and their position relative to the PLF is given in Figure 46. These schematic block diagrams of intermediate volcanic outcrops display the spatial relationships of C'-rich and C'-poor lithologies as observed in the field. Thus, strain partitioning not only produces a primary strain gradient during the initial formation of PLDZ, but continues to play an active role during the final stages of ductile deformation where a secondary strain gradient is observed between C'-rich and C'-poor zones in the centre of the shear zone (Fig. 49). These observations further exemplify the heterogeneous nature of deformation.

It is important to note that although the primary strain gradient increases in intensity from shear zone boundary to shear zone centre, the secondary strain gradient does not comply with the same configuration. C'-poor zones are commonly found separating C'-rich zones from the shear zone centre producing a fluctuating strain profile (Fig. 50). The C'-rich
Fig. 46. Location of outcrops P237 and P235 and their position relative to the PLF.
Fig. 47. Outcrop P237 exhibiting shear foliations
Fig. 48. Outcrop P235 exhibiting shear foliations.
Fig. 49. Primary and secondary strain gradient.

Fig. 50. Strain profile of secondary strain gradient.
zones where able to accommodate an imposed strain rate which
the country rock could not accommodate by bulk deformation
(White et al. 1980). Therefore, the development of C'-rich
zones is analogous to the development of individual shear
zones, yet they are not individual shear zones themselves.
There is no evidence that the PLDZ is composed of numerous
smaller shear zones. Field mapping indicates the presence of
only one major shear zone which exhibits a primary and
secondary strain gradient.

External shear zone boundaries demarcate the outer limits
of the PLDZ. On the south side of the PLDZ, the external
boundary is marked by the onset of an S-fabric which crosscuts
lithological contacts as it rotates into parallelism with the
shear zone centre. On the north side of the PLDZ, the position
of the external boundary is problematic because of the
presence of a preexisting foliation. The two foliations can be
separated only on the basis that the preexisting foliation
parallels lithological contacts, whereas the S-fabric lies
oblique to lithological contacts and exhibits a sigmoidal
geometry similar to that observed on the south side. The
primary strain gradient is enveloped by the external shear
zone boundaries (Fig. 51).

The internal shear zone boundaries are located within the
central portion of the PLDZ. These boundaries are not clearly
defined and are estimated to be positioned approximately 100 m
- 200 m from the shear zone centre. The internal boundaries
Fig. 51. External and internal shear zone boundaries.
mark the onset of the heterogeneous sector of the PLDZ consisting of C'-rich and C'-poor zones. This region represents the highest shear strain in the PLDZ and delineates the secondary strain gradient (Fig. 51).

Despite many shortcomings encountered when calculating the amount of strain in large scale shear zones, it is not simply an exercise in futility. The Ramsay and Graham (1970) model gives an estimate of the magnitude of shear strain and total displacement in the PLDZ, albeit subject to much variation. More importantly, strain profiling illustrates the presence of a simple, primary strain gradient and a complex, secondary strain gradient. Strain analysis enables the division of the PLDZ into domains corresponding to increasing deformation which are microscopically analyzed in the next chapter.
CHAPTER V - MICROSTRUCTURAL ANALYSIS

V.I MICROSTRUCTURES AND TERMINOLOGY

Five strain domains were recognized in the PLDZ during field mapping and thin section study. Each domain is defined by the type of foliation(s) and the intensity of deformation. A detailed microstructural analysis was performed on a representative thin section, one which best exhibited the foliation(s) type from each domain, to examine and document microstructural changes during progressive deformation. Sections were cut perpendicular to the foliation and parallel to the lineation (i.e. XZ plane). Microstructures were observed with an optical microscope capable of 2.5 x 10 to 40 x 10 magnification. Measurements of mean grain size (length and width) and mean aspect ratios (length/width) were made with a Apple IIe computer-interfaced graphics tablet which has an error of approximately 4%.

Consideration must be given to primary variations within each lithology in a study like this. In other words, is the original lithology of Domain 1 the same as the original lithology of Domain 5. Because of the width of the PLDZ and the lack of continuous outcrop, there is no guarantee that the sample selection is from an isotropic rock. Outcrops that best represent each domain were selected across the strike of the shear zone. Some of these domains were located adjacent to each other in outcrop, others were located up to hundreds of
metres along strike. As always, with increasing deformation, the original lithology is often obscured and/or obliterated. This chapter will demonstrate that consistent microstructural trends observed from shear zone boundary to centre in a conglomerate and an intermediate volcanic unit can be best explained by a tectonic origin.

Microstructures were divided into three categories based on cataclasis, dislocation and diffusion creep deformation mechanisms. This classification scheme, presented in Table 3, is modified from the work of Bell and Etheridge (1973) and White (1976).

Microstructural analysis enables classification of strain domains using the terminology proposed by Wise et al. (1984): protomylonite, mylonite, and ultramylonite. Ultramylonites in the PLDZ have been subdivided into α-mylonites and β-mylonites to distinguish Domain 4 from Domain 5, respectively.

**V.II MYLONITIZATION OF A CONGLOMERATE**

**V.IIa DOMAIN 1 - WEAKLY DEFORMED**

Domain 1 best represents the primary mineralogical assemblage of the conglomerate because deformation is minimal. Gravel-size clasts, predominantly felsic volcanic and granitoid in composition, sit in a matrix of sand-size quartz and plagioclase grains which exhibit a crude equigranular texture. A minor mud-size fraction of quartzo-feldspathic material is also present. The gravel-size clasts will be
Table 3. Deformation mechanisms and microstructures.

<table>
<thead>
<tr>
<th>Cataclasis</th>
<th>Dislocation Creep</th>
<th>Diffusion Creep</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Microfracturing</td>
<td>1) Deformation structures - undulose extinction - deformation bands</td>
<td>1) Pressure shadows</td>
</tr>
<tr>
<td></td>
<td>2) Recovery structures - polygonization - subgrains</td>
<td>2) Bubble trails parallel grain boundaries</td>
</tr>
<tr>
<td></td>
<td>3) Recrystallization - new grains - serrate, lobate, and kink band boundaries</td>
<td>3) Lack of internal deformation in strain-free grains in microcrystalline matrix</td>
</tr>
</tbody>
</table>
excluded from modal analysis estimates during microstructural studies.

Domain 1 is defined by the onset of a weak foliation. This foliation is often difficult to discern in the field as gravel-size clasts are virtually undeformed (Fig. 52). In thin section the foliation is manifested as a weak alignment of quartz, plagioclase and rock fragments which delineate a plane of flattening or S-fabric (Fig. 53). These sand-size components constitute approximately 85% of the matrix (Q Quartz = 40%; P Plagioclase = 30%; F Rock Fragments = 15%). Quartz and plagioclase grains are subangular in shape with mean grain size measuring 442/249 µm for quartz and 471/310 µm for plagioclase. Aspect ratios average 1.88 for quartz and 1.73 for plagioclase.

Evidence for dislocation creep is exhibited by undulose extinction which is the most common deformation structure observed in both quartz and plagioclase grains. Deformation bands are rare, as are recovery structures such as subgrains and polygonization. Primary twinning is well preserved in plagioclase, yet all grains have undergone some degree of sericitization.

The remaining 15% of the matrix consists of mud-size (<10 µm) microcrystalline material. Although difficult to identify, the matrix consists of rectangular to equant grains believed to be of quartzo-feldspathic composition. Platy chlorite grains observed in the matrix are similar in size to
Fig. 52. Weakly deformed conglomerate from Domain 1 (pencil parallel to S-fabric).

Fig. 53. Weak alignment of quartz, plagioclase and rock fragments which define S-fabric (field of view ≈ 7mm).
this finer fraction. Carbonate material is observed interstitial to plagioclase and quartz grains.

V.IIb DOMAIN 2 - PROTOMYLONITE

Domain 2 is recognized in the field by the presence of a moderate C-fabric in which gravel-size clasts become stretched in the direction of shear flow. The C-fabric is outlined in thin section by discrete shear planes (Fig. 54). Discontinuous lenses, approximately 60 μm thick, consisting of aggregates of white mica and lenses of quartzo-feldspathic material with chlorite parallel the C-fabric. This may represent compositional layering at the incipient stage. The existence of an S-fabric is identified in thin section by the orientation of the long axis of some quartz and plagioclase grains oblique to the C-fabric.

At this stage of deformation a porphyroclastic texture has begun to evolve. Porphyroclasts are estimated to comprise up to 80 % the conglomerate (Q=40;P=25;F=15). Quartz and plagioclase grains range from subangular to almost subrounded in shape. Mean grain size of quartz measures 438/219 μm and aspect ratios average 2.11. Plagioclase grains average 439/256 μm with mean aspect ratios of 2.39. Mean grain size decreases and mean aspect ratios increase compared to Domain 1. Primary twinning is still preserved in plagioclase grains but the amount of sericitization has increased. Rock fragments become difficult to distinguish from the individual grains and
Fig. 54. Discrete shear planes define a C-fabric in Domain 2 (field of view \(\approx 7\text{mm} \)).

Fig. 55. Subgrain formation, parallel to S-fabric, resulting in core and mantle structures (field of view \(\approx 7\text{mm} \)).
matrix.

An increase in dislocation creep structures is observed in Domain 2. Undulose extinction is abundant in most quartz and plagioclase grains. Deformation bands are present but rare. Subgrains and polygonization textures begin to develop (Fig. 55), especially in the margins of grains, producing core and mantle structures (Grifkins 1975). Subgrains may become misorientated in the mantle with an increase in dislocation density and form new strain-free grains during dynamic recrystallization according to White (1976). Serrated grain boundaries observed in quartz porphyroclasts represent a second mode of recrystallization known as strain induced grain boundary migration or bulging (Hutchinson 1974). During this process new grains are formed from bulges of an existing boundary into its less deformed neighboring grain. This bulging is driven by a strain gradient between adjacent grains (White 1976).

Subgrains and new strain-free grains range in size from half the dimensions of the host grain to very fine grains within porphyroclast tails estimated to be <10 µm. Grain size comminution is a continual process during cyclic dynamic recrystallization and is best exemplified in the production of recrystallized tails of quartz composition adjacent to quartz porphyroclasts and the production of tails of plagioclase and white mica composition adjacent to plagioclase porphyroclasts. These tails parallel the C-fabric in thin section, yet their
symmetric or inconsistent asymmetric orientation about the host porphyroclasts gives ambiguous or conflicting sense of shear (Simpson and Schmid 1983; Passchier and Simpson 1986) at this stage of mylonitization.

The microcrystalline matrix is similar in grain size (<10 μm) to Domain 1, but now occupies 20% of the rock. This slight increase is due to the development of white mica and additional quartzo-feldspathic material during syntectonic recrystallization of plagioclase and quartz porphyroclasts. Platy chlorite interleaved within the quartzo-feldspathic material parallels the C-fabric. The existence of a mylonitic foliation (C-fabric) and the presence of porphyroclasts in a very fine-grained matrix permits the classification of Domain 2 as a protomylonite.

V.IIc DOMAIN 3 - MYLONITE

A strong C-fabric dominates Domain 3 and corresponds with the C'-poor zone mapped in the field. Gravel-size clasts are stretched to the extent that some may not be discriminated from matrix on a mesoscopic scale (Fig. 56). Discontinuous layers, similar in composition to Domain 2, display a crude compositional layering on a microscopic scale. These layers average 100 μm in thickness and parallel the C-fabric. Recrystallized porphyroclast tails also parallel the C-fabric, thus enhancing the foliation. An irregular, widely spaced C'-fabric is observed sporadically in thin section, but
Fig. 56. Conglomeratic clasts stretched and rotated into parallelism with C-fabric in Domain 3 (pencil parallel to C-fabric; note clast and fringe asymmetry).

Fig. 57. Early development of recrystallized porphyroclast tails in Domain 3 (field of view ≈ 7mm).
is not detected in the field.

The proportion of porphyroclasts to matrix is approximately 50:50 at this stage of mylonitization. Porphyroclast composition is estimated to consist of Q=25;P=15;F=10. Quartz and feldspar grains are rounded and more elliptical than Domain 2. Average quartz grain size decreases to 408/192 μm, as does mean plagioclase grain size to 410/203 μm. Mean aspect ratios for quartz and plagioclase increase to 2.27 and 2.54, respectively. Primary twinning is not as well preserved in plagioclase as sericitization increases. Rock fragments are most difficult to identify.

Dislocation creep remains the dominant deformation mechanism. Undulose extinction is still abundant in quartz and plagioclase porphyroclasts as in Domain 2, deformation bands are rare. However, core and mantle structures are more common and development of subgrains and production of new grains result in well defined recrystallized porphyroclast tails (Fig. 57). A dextral shear sense can be deduced from the asymmetry of these tails as syntectonic recrystallization parallels the direction of flow along the C-fabric.

The matrix is similar in composition and grain size (<10 μm) to Domain 2, except that the proportion of matrix has increased to 50% of the conglomerate which designates mylonite classification. It is evident that an increase in total matrix content is a function of increased dynamic recrystallization resulting in grain size comminution and the production of
porphyroclast tails. Attenuation of these tails along the shear plane generates more matrix material. Recrystallized tails, because of their grain size (<10 μm), undergo further deformation by diffusional processes (White 1976) which will be addressed in Domain 4. A progressive increase in white mica in the matrix may be positively correlated with an increase in sericitization of plagioclase grains.

V.IId DOMAIN 4 - α-ULTRAMYLONITE

The introduction of a regular, closely spaced, crosscutting C'-fabric distinguishes Domain 4 from previous strain domains. The conglomerate is intensely deformed and both the C'-fabric and the C-fabric are well developed (Fig. 58). Excepting clasts of granitoid and quartz composition, gravel-size clasts are difficult to discern from the matrix. C' foliation planes are discontinuous along strike measuring centimetres in length and spacing between C' foliation planes ranges from millimetres to centimetres. In thin section compositional layering, which parallels the C-fabric, is more pronounced than in Domain 3 as layering increases in thickness to approximately 150 μm. Recrystallized porphyroclast tails continue to enhance the mylonitic foliation (C-fabric) except when crosscut by the C'-fabric. Here the preexisting foliation is extended and displaced by the C'-fabric which behaves like a microshear with normal displacement in the same sense as the bulk shear direction. White mica and carbonate material is
Fig. 58. C'-fabric (pen) cross-cutting C-fabric (pencil) in Domain 4.

Fig. 59. C'-fabric behaves like a microshear displacing C-fabric in Domain 4 (field of view ≈ 7mm).
preferentially found along the C'-fabric.

Porphyroclasts of quartz and plagioclase are subrounded and represent an estimated 35% of the conglomerate (Q=25;P=10). Quartz mean grain size continues to decrease to 370/218 µm, but a slight increase in plagioclase mean grain size to 445/213 µm is observed. Mean aspect ratios measure 2.15 and 2.29 for quartz and plagioclase grains, respectively, which is marginally less than Domain 3. This slight decrease deviates from a gradual increase in average aspect ratios observed in previous strain domains with progressive deformation. From this data, supported by measurements of aspect ratios in Domains 3 and 5 (see Table 4), it is inferred that an equilibrium ratio has been attained. Primary twinning is only rarely detectable in plagioclase grains due to the degree of sericitization.

Dislocation creep structures are similar to those in Domain 3 with undulose extinction and core and mantle structures still abundant in porphyroclasts. Recrystallized porphyroclast tails are better developed and beautifully display asymmetric orientations which serve as indisputable kinematic indicators for a dextral sense of shear. However, where C'-fabric crosscuts these tails they are deflected from parallelism with the C-fabric into the microshear plane of the C'-fabric (Fig. 59).

The matrix remains constant in grain size (<10 µm), but now represents an estimated 65% of the conglomerate.
Dislocation creep is important during deformation of porphyroclasts but at this stage of mylonitization is a less effective mechanism than diffusion creep during deformation of matrix and tails which comprise the bulk of conglomerate. This change in deformation mechanism accompanies a reduction in grain size and is clearly explained by White (1976, p. 79):

"Although dislocation creep processes are required to produce the recrystallization needed for grain refinement, they result in a strain rate which is independent of grain size and they cannot give the degree of strain softening required to preferentially concentrate deformation to the smallest grain zone. However, strain rate is inversely proportional to grain size if deformation is by diffusion creep and for this to become dominant would require a change in deformation mechanism with grain refinement."

Matrix composition is similar to that in Domain 3 except for a continued increase in white mica content at the expense of plagioclase. Mica fish (Lister and Snokes 1984) and foliation fish (Hanmer 1986) are present. An increase in carbonate material is observed due to the development of dilation zones along the C'-fabric. Although the porphyroclast:matrix ratio is estimated to be 35:65, this domain may be classified as a α-mylonite because of the intensity of deformation, matrix grain size and the presence of a well-developed C'-fabric.

V.IIe DOMAIN 5 - β-ULTRAMYLONITE

Domain 5 was recognized in the field by a decrease in the angle (β value) between the C-fabric and C'-fabric. This marks
the highest degree of deformation during mylonitization. Gravel-size clasts are difficult to identify but are present, and the original lithology may only be recognized as a metasedimentary rock by a pink, salmon colour observed in outcrop. The rock is much more fissile than previous domains. C- and C'-fabrics are similar to Domain 4 but for their angular relationships which decreases. Compositional layering continues to increase in thickness to approximately 500 µm.

Porphyroclasts make up only an estimated 10% of the conglomerate (Q=6;P=4). Quartz and plagioclase grains decrease in mean size to 280/141 µm and 145/068 µm, respectively. Aspect ratios appear to maintain an equilibrium value. The average quartz grain aspect ratio is 2.35 and plagioclase measures 2.45. Sericitization engulfs most of the few remaining plagioclase grains.

Evidence still exists for syntectonic recrystallization but porphyroclasts are few. Undulose extinction, subgrains and new grains in recrystallized porphyroclast tails are present but not as abundant as previous domains. These tails are not as well defined and certainly less spectacular. Most of the deformation appears to be confined to the microcrystalline matrix where absence of dislocation microstructures implies that diffusion creep operates as the dominant deformation mechanism.

Approximately 90% of the mylonitized conglomerate is now matrix. The content of white mica is highest at this stage and
Fig. 60. Foliation fish present in Domain 5 (field of view $\approx 7 \text{mm}$).
is estimated to comprise approximately half of the total matrix as platy aggregates which form layers. The remainder of the matrix is similar to Domain 4 excepting the addition of foliation fish (Fig. 60). This domain is classified as a \( \beta \)-ultramylonite.

Mylonitization of the conglomerate in the PLDZ is summarized in Table 4.
Table 4. Summary of mylonitization of conglomerate.

<table>
<thead>
<tr>
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<td>mylonite</td>
<td>α-ultramylonite</td>
<td>β-ultramylonite</td>
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<td>strong C</td>
<td>intense C'</td>
<td>intense C'</td>
</tr>
<tr>
<td></td>
<td>weak S</td>
<td>incipient C'</td>
<td>intense C</td>
<td>intense C</td>
<td></td>
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<td>408/192</td>
<td>370/218</td>
<td>280/141</td>
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<td>439/256</td>
<td>410/203</td>
<td>445/213</td>
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<td>&lt;10</td>
<td>&lt;10</td>
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<td>subgrains</td>
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<td>foliation fish tails</td>
<td></td>
</tr>
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<td>deform. bands</td>
<td>undulose ext. tails</td>
<td>core &amp; mantle foliation fish</td>
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<td>polygonization serrate grains</td>
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<td>(rare)</td>
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<td>dislocation</td>
<td>diffusion</td>
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V.III MYLONITIZATION OF A VOLCANIC

V.IIIa DOMAIN 1 - WEAKLY DEFORMED

A typical example of the intermediate volcanic is represented by a weakly deformed pyroclastic breccia. Fragments comprise up to 50% of the volcanic, range in size from 2 cm to 25 cm and are more felsic in composition than matrix material (Fig. 12). Fragments will be excluded from modal analysis estimates during microstructural studies. The fine-grained portion of the tuff breccia is probably of andesitic composition and exhibits a porphyritic texture. Plagioclase phenocrysts, with minor quartz phenocrysts, constitute approximately 25% of the rock which sit in a matrix of plagioclase with minor actinolite, chlorite +/- epidote.

Domain 1 is defined in the field by a weak S-fabric foliation. The long axes of fragments are generally parallel to the foliation. In thin section plagioclase phenocrysts are randomly oriented. Mean grain size measures 932/508 μm and aspect ratios average 1.90. Plagioclase grains are subangular to subrounded in shape and sericitization often masks primary twinning. When preserved, twins remain continuous, sharp and are usually bent.

Cataclasis is the dominant deformation mechanism exhibited in plagioclase phenocrysts which are commonly fractured, but not mechanically separated. Dislocation creep structures, including undulatory extinction and subgrains,
give phenocrysts a poorly developed core and mantle structure. Weakly developed recrystallized tails adjoining some of the phenocrysts implies incipient recrystallization. Quartz phenocrysts are rare, less than 5% of phenocrysts, and exhibit similar microstructures as plagioclase grains with the addition of serrated grain boundaries and polygonization texture indicative of grain boundary bulging.

The matrix, approximately 75% of the volcanic, consists of predominantly subangular to lath-shaped plagioclase grains which exhibit dislocation creep microstructures similar to plagioclase phenocrysts. They have a mean grain size of 86/37 μm and a mean aspect ratio of 2.28. Matrix plagioclase grains are weakly aligned to define the foliation (Fig. 61). Fine-grained (<10 μm) platy chlorite is dispersed throughout the matrix.

V.IIIb DOMAIN 2 - PROTOMYTONITE

A moderate foliation is recognized in the field and interpreted in thin section to be a C-fabric. This is manifested by the preferred orientation of chlorite and plagioclase grains parallel to slip planes in the matrix. Some of the fractured plagioclase phenocrysts become mechanically separated due to extension of the grain along the shear plane. They now behave as porphyroclasts and constitute approximately 20% of the volcanic. The obliquity of the long axes of the larger plagioclase phenocrysts observed in thin section
Fig. 61. Matrix plagioclase grains in volcanic are weakly aligned to define an S-fabric in Domain 1 (field of view $\approx 7\text{mm}$).

Fig. 62. Long axes of plagioclase phenocrysts aligned parallel to S-fabric which lies oblique to C-fabric in Domain 2 (field of view $\approx 7\text{mm}$).
defines an S-fabric, while smaller grains have rotated into the C-fabric (Fig. 62).

Plagioclase porphyroclasts are subangular in shape, sericitization is common, and primary twinning, when visible, may be bent and displaced. Mean grain size of the porphyroclasts has decreased to 381/212 μm and the mean aspect ratio has decreased to 1.93. Undulose extinction and core and mantle structures are common. Deformation bands are present but rare. Recrystallized porphyroclast tails are better developed than those in Domain 1, but remain relatively weak. The tails are of the same composition as the parent porphyroclast, like those observed in the conglomerate, and parallel the shear plane. Serrate and lobate grain boundaries are common in the few isolated quartz porphyroclasts that are present.

Shear sense inferred from the asymmetry of the tails is ambiguous at this stage. However, the obliquity of the long axes of the majority of plagioclase porphyroclasts and the sense of displacement between broken grains indicate dextral shear.

The amount of matrix has increased to approximately 80% of the rock. The preferred orientation of matrix plagioclase grains delineate the C-fabric. Mean grain size of matrix plagioclase decreases to 28/17 μm and mean aspect ratios decrease to 1.93. Very fine-grained (<10 μm) quartzofeldspathic material constitutes a minor portion of matrix
which was produced during dynamic recrystallization. Platy chlorite is scattered throughout, producing a matrix comparable in composition and grain size to the conglomerate. Dislocation creep microstructures similar to those observed in the plagioclase porphyroclasts are present in the matrix.

Aggregates of white mica may form discontinuous lenses averaging 30 μm thick. Long axes of chlorite and white mica grains parallel the C-fabric. Carbonate material is found infilling along the fractured plagioclase porphyroclasts and may develop as pressure shadows (Spry 1969; Simpson and Schmid 1984). Domain 2 represents a protomylonite.

V.IIIc DOMAIN 3 - MYLONITE

Domain 3 corresponds with the C'-poor zone mapped in the field and is shown in Figure 63 adjacent to a C'-rich zone. A strong C-fabric is defined in thin section by discrete shear planes and a compositional layering which is approximately 30 μm thick (Fig. 64). In the microcrystalline matrix of alternating quartzo-feldspathic material and, platy chlorite and white mica grains defines the layering which parallels the C-fabric. The preferred orientation of plagioclase porphyroclasts also parallels the C-fabric. An incipient C'-fabric displays an irregular, widely spaced, cross-cutting foliation in thin section. Sporadic spacing and lengths measuring but a few millimetres makes detection of this foliation in the field virtually impossible.
Fig. 63. C-fabric in Domain 3 (pencil with eraser) adjacent to C'-fabric in Domain 4 (pencil without eraser).

Fig. 64. C-fabric in Domain 3 defined by discrete shear planes and a compositional layering (field of view ≈ 7mm).
Plagioclase porphyroclasts rarely exhibit primary twinning which is usually obscured by sericitization. Porphyroclast shape varies from subangular to subrounded. Undulose extinction and subgrains remain common, however, there is no longer any evidence of fracturing. Well developed recrystallized tails are the dominant microstructure and their asymmetry now serves as a reliable kinematic indicator. The porphyroclast:matrix ratio decreases to approximately 15:85, as does the mean porphyroclast grain size (200/103 µm). The mean aspect ratio measures 1.98 which is a slight increase from Domain 2.

The amount of matrix has increased to approximately 85% of the rock, coinciding with an increase in dynamic recrystallization. Mean grain size of the matrix has decreased to microcrystalline size, approximately <10 µm. Although dislocation creep continues to operate on porphyroclasts, the small grain size of the matrix implies deformation by diffusion creep. Matrix composition is dominated by quartzo-feldspathic material interleaved with platy chlorite and white mica. Carbonate material continues to infill as pressure shadows around porphyroclasts. Domain 3 is classified as a mylonite.

V.IIIId DOMAIN 4 - α-ULTRAMYLONITE

A regular, closely spaced, crosscutting C'-fabric dominates Domain 4. This corresponds with the C'-rich zone in
the field (Fig. 28). Spacing between the \( C' \) foliation planes and the length of these discontinuous planes is equal to those observed in the conglomerate. Compositional layering, dominated by layers rich in white mica, increases to 40 \( \mu \text{m} \) in thickness and parallels the \( C \)-fabric. Both the \( C \)- and \( C' \)-fabrics are well developed.

Plagioclase porphyroclasts constitute an estimated 8\% of the volcanic and their mean grain size has decreased to approximately 93/50 \( \mu \text{m} \), with a mean aspect ratio of 1.97. Porphyroclasts parallel the shear plane and have become more rounded in shape. Although porphyroclasts are small, they still exhibit microstructures similar to Domain 3. Nonetheless, dislocation creep processes are less important than diffusion creep processes operating in the matrix at this stage of mylonitization.

Normal displacement of the \( C \)-fabric by the \( C' \)-fabric produces foliation fish (Fig. 65). An increase in white mica accounts for approximately 20\% of the total matrix content. Preferential infilling of carbonate material along the \( C' \)-fabric may reach up to 2 mm in thickness. The long axes of carbonate grains generally parallel the \( C \)-fabric and may be subsequently displaced along the \( C' \)-fabric (Fig. 66). Recrystallized quartz grains are common along the margins of carbonate infilling and in the centre of carbonate material which has been sheared. Domain 4 is labelled an \( \alpha \)-ultramylonite.
Fig. 65. Normal displacement of C-fabric by C'-fabric, which is infilled with carbonate material in Domain 4 (field of view ≈ 7mm).

Fig. 66. Carbonate material infilling along C'-fabric is subsequently sheared (field of view ≈ 4.5mm)
V.IIIe  DOMAIN 5 - β-ULTRAMYLONITE

The β-ultramytonite in Domain 5 contrasts with the α-mylonite in Domain 4 as a result of a decrease in the angular relationship (β value) between the C-fabric and the C'-fabric measured in the field. Domain 5 represents the most advanced stage of mylonitization and is accompanied by an increase in fissility and a state of deformation which leaves the original lithology almost totally obscured (Fig. 67). A green coloration enables the rock to be identified as a volcanic.

Subrounded plagioclase porphyroclasts exhibit dislocation creep microstructures similar to Domain 4, but diffusion creep mechanisms continue to dominate deformation. Mean grain size of plagioclase porphyroclasts increases to 176/101 μm. Mean aspect ratios measure 1.81, which is marginally less than previous domains.

The porphyroclast:matrix ratio remains relatively constant at an estimated 8:92. The matrix is similar in grain size and composition to Domain 4, excepting an increase in thickness of compositional layering (150 μm) and an increase in white mica content to approximately 50% of the total matrix in parts and manifested as foliation fish (Fig. 68).

A summary of mylonitization of an intermediate volcanic is provided in Table 5.
Fig. 67. $C'$-fabric (pen) cross-cutting $C$-fabric (pencil) in Domain 5.

Fig. 68. Foliation fish of white mica composition in Domain 5 (field of view $\approx 7\text{mm}$).
Table 5. Summary of mylonitization of an intermediate volcanic.

<table>
<thead>
<tr>
<th>DOMAIN</th>
<th>1</th>
<th>2</th>
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V.IV COMPARISON OF MYLONITIZATION IN CONGLOMERATE AND VOLCANIC

The northern portion of the PLDZ lies within a predominantly volcanic unit of intermediate composition, whereas the southern segment lies within a primarily conglomeratic unit. Some mafic volcanics lie to the north and south of the PLF in the western half of the study area. Five domains have been identified in both the conglomerate and the volcanic, suggesting that both lithologies responded in a similar manner to the same deformation process. C'-rich and C'-poor zones, Domains 3 and 4, were observed in the mafic volcanics, however, an insufficient amount of outcrop inhibited a detailed study of mylonitization of this type of lithology.

Each lithology has unique properties such as texture, mineralogy and grain size which may dictate the type of microstructures produced during deformation. Textural differences between the volcanic and the conglomerate may explain differences in the trends of mean aspect ratios with increasing strain. Plagioclase and quartz grains display a poor equigranular texture in the conglomerate and undergo grain elongation during the initial stages of mylonitization. The development of a porphyroclast/matrix system with increasing strain shows that mean aspect ratios of porphyroclasts reach an equilibrium value as new, smaller strain-free grains are produced. These new grains are formed in the mantle of porphyroclasts and are subjected to further
dynamic recrystallization. Thus, comminution of porphyroclasts
is achieved without further grain elongation at this stage of
mylonitization. This is observed in the statistical trend of
mean aspect ratios for porphyroclasts in the conglomerate.

An original porphyritic texture is present in the
intermediate volcanic. Brittle plagioclase phenocrysts deform
initially by microfracturing. A more ductile matrix,
consisting of a pilotaxitic arrangement of predominantly
plagioclase composition, develops a preferred orientation
defined by the alignment of the long axes of plagioclase
grains. Grain elongation is observed in phenocrysts, but it is
not statistically significant during the initial stages of
mylonitization. Like the conglomerate, aspect ratios appear to
attain an equilibrium value with increasing strain. The
original lath shape of the relatively undeformed plagioclase
grains in the matrix possesses an initial high aspect ratio.
This may explain the absence of an increase in aspect ratios
observed in the matrix of the volcanic with increasing strain.
Increases in mean aspect ratios during mylonitization are most
common in lithologies, such as granites or other plutonic
rocks, with an original equigranular texture (Berthé et al.
1979, Brodie 1980).

In addition to differences between lithologies,
consideration must be given to any primary, lithological
variation which may exist within in each lithology. In the
conglomerate unit gravel-sized clasts decrease in size to the
south, away from the shear zone centre. Detailed work by Devaney (1987) interprets this as a proximal-distal relationship in which the sequence coarsens upward to the north. Although no studies of the matrix from such a sequence have been found in the literature, an increase or at least constant matrix grain size may be expected to accompany an increase in the size of the clasts. Microstructural analysis reveals that mean grain size of the matrix decreases toward the PLF with increasing strain. Such a trend is unlikely due to a primary grain size variation. Evidence for similar arguments can not be made for the volcanic unit as more detailed mapping is required north of the PLDZ.

Field mapping and thin section studies have shown that consistent trends observed during mylonitization of both the conglomerate and the volcanic with increasing strain can be best explained by a tectonic origin. Primary variations may explain local inconsistencies such as a slight increase in mean grain size of plagioclase porphyroclasts in Domain 4 of the conglomerate and in Domain 5 of the volcanic, however, both of these samples give large standard deviations for mean grain size measurements.

Table 6 lists consistent microstructural trends, common to the conglomerate and the volcanic, which were observed with increasing strain during the mylonitization process.
Table 6. Consistent microstructural trends observed during mylonitization of a conglomerate and a volcanic with increasing strain.

<table>
<thead>
<tr>
<th>Property</th>
<th>Trend</th>
</tr>
</thead>
<tbody>
<tr>
<td>porphyroclast mean grain size</td>
<td>decrease</td>
</tr>
<tr>
<td>matrix mean grain size</td>
<td>decrease</td>
</tr>
<tr>
<td>porphyroclast:matrix ratio</td>
<td>decrease</td>
</tr>
<tr>
<td>compositional layering</td>
<td>increase</td>
</tr>
<tr>
<td>mean aspect ratio</td>
<td>increase + constant</td>
</tr>
<tr>
<td>degree of sericitization</td>
<td>increase</td>
</tr>
<tr>
<td>white mica content</td>
<td>increase</td>
</tr>
<tr>
<td>plagioclase primary twins</td>
<td>decrease</td>
</tr>
</tbody>
</table>
V.V MICROSTRUCTURAL DEFORMATION MECHANISMS

Deformation mechanisms active during mylonitization play variable roles with progressive deformation. White (1976) has shown that this is a function of grain size. Grain size governed the deformation mechanisms operating on the porphyroclasts and the matrix in both lithologies. Porphyroclasts typically exhibit microstructures indicative of dislocation creep, while the matrix and porphyroclast tails, upon reaching microcrystalline size of approximately \( <10 \mu m \), deformed by diffusion creep. Thus, the ratio of dislocation:diffusion appears to be positively correlated with the clast:matrix ratio. Identification of diffusion creep microstructures involving grain boundary sliding and/or grain boundary migration was not made possible by examination of thin sections with an optical microscope capable of a maximum 40 x 10 magnification. S.E.M or T.E.M examination is required to properly document deformation mechanisms at this scale. Such a study is beyond the scope of this thesis.

Nevertheless, a set of deformation maps from White (1976) clearly illustrates the significance of grain size on the deformation mechanism operating during mylonitization (Fig. 69). White (1976) is quick to comment that these grain sizes represent only ball-park accuracy but the maps are not limited to quartzites and may be extended to polycrystalline rocks. Figure 70 illustrates the interpretation of deformation mechanism from grain size reduction during mylonitization.
Fig. 69. Deformation mechanism map (differential stress vs. temperature) for quartz of differing grain size, with strain rate contours (from White 1976).
Fig. 70. Attenuation of recrystallized porphyroclast tail
(field of view ≈ 4.5mm)
Grain size reduction begins with the development of small subgrains by dislocation processes in the mantle of a plagioclase porphyroclast. Recrystallization results in the production of new strain-free grains which undergo continual recrystallization during comminution. Attenuation of these grains with increasing strain until dislocation microstructures were no longer observed, typically less than 10 μm, is presumed to have been a result of diffusional processes.

Mean grain size data obtained from both the volcanic and conglomerate, supported by microstructural analysis, suggests that dislocation creep processes dominated the early stages of mylonitization (Domains 1-3), whereas diffusion creep processes prevailed in the late stages (Domains 4-5). C’-fabric development is temporally associated with the dominance of diffusional processes. Microfracturing of plagioclase phenocrysts in Domain 1 of the volcanics demonstrates the competency contrast between brittle plagioclase and a more ductile matrix. This is the only domain in which cataclasis was the dominant deformation mechanism and is excluded from the diagram below. Figure 71 is a qualitative plot of dislocation creep versus diffusion creep which illustrates the change in the dominant deformation mechanism accompanying a reduction in grain size during mylonitization of a conglomerate and volcanic.
Fig. 71. Qualitative change in deformation mechanism with increasing strain.
CHAPTER VI - METAMORPHISM

VI.1 REGIONAL METAMORPHISM

Rocks of early Precambrian age in the map area have been metamorphosed to lower greenschist facies during a regional metamorphic event. Equilibrium mineral assemblages in the main lithologies consist of the following:

1) volcanics: albite-actinolite-chlorite-epidote-sericite-quartz-carbonate
2) conglomerate: albite-chlorite-sericite-quartz-carbonate
3) trondhjemite: albite-chlorite-epidote-sericite-quartz
4) gabbro: albite-actinolite-chlorite-epidote/zoisite-sericite-quartz-carbonate

Felsic porphyry, lamprophyre and diabase dikes exhibit equilibrium mineral assemblages which have suffered minimal metamorphism, hence their emplacement is believed to post date peak metamorphism.

The inner part or middle of gabbroic bodies/coarse-grained volcanic flows may retain original primary mineral assemblages. Coarse-grained clinopyroxene may show little, if any, sign of metamorphism. From this relationship it is evident that the centres of coarse-grained lithologies behave as resistant and relatively impermeable bodies during low grade metamorphism.

A slight metamorphic gradient has been documented by Robinson (1986) in the BGB from lower greenschist facies in
the north of the belt to upper greenschist facies in the south. Narrow contact metamorphic aureoles mapped by Mackasey and Wallace (1978) around trondhjemitic stocks north of the PLDZ are superimposed on the regional metamorphism and subject surrounding volcanics to silicification, recrystallization and/or feldspathization. Kresz and Zayachivsky (1986) report amphibolitized volcanics adjacent to one of the stocks west of the study area in the OTT. Metamorphic aureoles have been mapped by Pirie and Mackasey (1978) in the OTT and south of the BGB. Gabbroic bodies have no metamorphic effect on host lithologies, in agreement with studies by Lawson (1986) just south of the map area. No apparent change in metamorphic grade was observed across the PLDZ.

Equation 3 represents a possible metamorphic reaction taking place during regional metamorphism of mafic volcanics under low grade greenschist facies. Due to the infinite number of possible reactions, the absence of geochemical data, the simplification of mineral formulas to end members and the presence of an open system, equation 3 and equation 4 are qualitative. Both equations are balanced as ionic equilibria to emphasize the importance of fluid-rock interaction.

The extent of muscovite, carbonate, and quartz production in the regionally metamorphosed rocks outside the PLDZ is minimal compared to that observed within the PLDZ. According to Winkler (1979) the absence of biotite in rocks of appropriate composition gives a temperature range from 350 -
15CaAl$_2$Si$_2$O$_6$ + 5Ca(Mg,Fe)Si$_2$O$_6$ + 2NaCa$_2$(Mg,Fe)$_5$Si$_8$O$_{22}$(OH)$_2$  Eqn. 3
Ca-rich plag  clinopyroxene  hornblende
+ 4H$_2$O + CO$_2$ + K$^+$ =

2NaAlSi$_3$O$_8$ + 2Ca$_2$(Mg,Fe)$_5$Si$_8$O$_{22}$(OH)$_2$ + Ca$_2$Al$_3$Si$_3$O$_{12}$(OH)
albite  actinolite  epidote
+ (Mg,Fe)$_5$Al$_2$Si$_3$O$_{10}$(OH)$_6$ + KAl$_2$(AlSi$_3$O$_{10}$)(OH)$_2$ + 6CaCO$_3$
chlorite  muscovite  carbonate
+ 5SiO$_2$ + 3H$^+$
quartz

2NaAlSi$_3$O$_8$ + Ca$_2$(Mg,Fe)$_5$Si$_8$O$_{22}$(OH)$_2$ + 2Ca$_2$Al$_3$Si$_3$O$_{12}$(OH)  Eqn. 4
albite  actinolite  epidote
+ 4H$_2$O + 6CO$_2$ + 2K$^+$ =
(Mg,Fe)$_5$Al$_2$Si$_3$O$_{10}$(OH)$_6$ + 2KAl$_2$(AlSi$_3$O$_{10}$)(OH)$_2$ + 6CaCO$_3$
chlorite  muscovite  carbonate
+ 11SiO$_2$ + 2Na$^+$
450 degrees C and estimated pressures of 2 - 8 kb at depths less than 25 km.

VI.II RETROGRADE METAMORPHISM

Retrograde metamorphism of the regionally metamorphosed rocks is restricted to the confines of the PLDZ. Retrograde metamorphism of conglomerate, intermediate and mafic volcanics within the PLDZ has produced an equilibrium assemblage consisting of chlorite-sericite-carbonate-quartz. Equation 4 represents a possible metamorphic reaction taking place during retrograde metamorphism of intermediate and mafic volcanics.

The retrograde reactions are hydration reactions and will only proceed on introduction of water into the rocks (Beach 1980). Chloritization, sericitization and carbonatization is pervasive in conglomerate, intermediate and mafic volcanics lying within the centre of the PLDZ. The presence of carbonate material in dilation zones, such as along C'-fabric and less commonly in tension gashes, indicates that much of the Ca\textsuperscript{2+} precipitated as carbonate in the system. The paucity of silicification and quartz filled structures within the PLDZ suggests that most of the silica produced has left the system with the excess Na\textsuperscript{+}. Thus, the fluid phase pervading the shear zone during retrograde metamorphism was probably H\textsubscript{2}O- and CO\textsubscript{2}-rich. An increase in white mica content observed with increasing deformation suggests that K\textsuperscript{+} was also introduced into the system during retrograde metamorphism of the PLDZ.
Primary mineral assemblages may contribute some K\(^+\) through reaction softening of feldspars (Dixon and Williams 1983), but are not considered to supply enough K\(^+\) to account for the amount of white mica in highly strained rocks.

Field mapping supported by thin section analysis demonstrates that retrograde metamorphism within the PLDZ post-dates a regional metamorphic event of low grade greenschist facies. The timing of retrograde metamorphism relative to deformation cannot be distinguished. Whether the presence of a fluid phase is a result or cause of deformation cannot be resolved. An increase in deformation is accompanied by an increase in retrograde minerals as these processes are intimately associated with the development of the PLDZ. Therefore, retrograde metamorphism is syn-deformational, each process enhances the other. This relationship between metamorphism and deformation in shear zones has been examined in the field and experimentally by workers over the years including Duchille and Roy (1964), Beach (1976), White and Knipe (1978), White et al. (1980). Beach (1980) believes that metamorphic reactions occur during the earliest stages of shear zone formation, however, they may not always be an essential part of shear zone formation. The source of the fluid phase in shear zones is controversial and still in doubt (Sibson et al. 1975; Beach 1976; Beach 1980).
A kinematic indicator exhibits the sense of flow in a shear zone. It has been defined by Hanmer (1986) as "a structure, resulting from flow, whose geometry is indicative of the progressive rotation of the principal axes of finite strain with respect to the principal axes (stretches) of the kinematic framework and/or the shear plane of deformation".

Careful interpretation requires that it be viewed in a plane oriented perpendicular to the foliation or shear plane, and parallel to the stretching lineation or shear direction.

Within ductile deforming materials, kinematic inference using a single structure must be based on numerous observations which give a consistent sense of shear. An assemblage of different kinematic indicators provides a more reliable interpretation. It is important to note that interpretations are limited to the scale of observation. Extrapolation to larger scales must be tentative and supported by a large data set.

The following kinematic indicators observed in the PLDZ are listed in order of decreasing reliability. Most types indicate a consistent dextral sense of shear.

1) S- and C-fabrics - Macroscopic analysis of foliations mapped within the PLDZ, illustrated by the construction of foliation trajectories (Fig. 72a & 27), show a rotation of a S-fabric into parallelism with the C-fabric indicative of dextral shear (Berthé et al. 1979).
Fig. 72A. Kinematic indicators (see text for descriptions).
Fig. 72B. Kinematic indicators (see text for description).
2) **C’-fabric** - A single set of C’-fabric has a consistent clockwise orientation with respect to the shear plane (C-fabric) on a macroscopic scale (Fig. 28 & 59). This angular relationship indicates dextral shear. In Figure 72b the sense of shear along the C’-fabric observed on a macroscopic and microscopic scale is the same as that of the main shear zone and can be used to deduce the overall sense of shear (Platt 1978; Weijermars and Rondeel 1984).

3) **Asymmetric Augen Structures** - Microscopic study reveals large and relatively flow resistant grains of quartz and plagioclase, referred to as porphyroclasts or augen, within a more ductile, fine-grained matrix. Fine-grained recrystallized tails, generally of the same composition as the parent porphyroclasts, are asymmetrically distributed around the augen structure extending along the foliation plane in the direction of shear (Fig. 72c & 70). Such microstructures are analogous to Passchier and Simpson's (1986) sigma-type porphyroclast systems.

4) **Asymmetric Pressure Shadows** - A high ductility contrast may also produce regions in the rock which are protected from deformation by the presence of relatively rigid porphyroclasts. Shadows occur where a foliation wraps around the porphyroclast with carbonate material and white mica concentrating in the protected region (Fig. 72d & 73). The asymmetry of the pressure shadow wings shows the sense of shear (Simpson and Schmid 1983; Spry 1983).
Fig. 73. Carbonate pressure shadow developed in protected region around quartz grain (field of view ≈ 4.5mm).

Fig. 74. Fibrous quartz growth forms pressure fringe around opaque grain (field of view ≈ 4.5mm).
5) **Asymmetric Pressure Fringes** - Opaque crystals, probably magnetite and/or pyrite, may develop pressure fringes around them due to inhomogeneous strain (Spry 1983). The fringes are typically composed of fibrous quartz growths with long axes curved to give the total structure an asymmetric shape from which shear sense can be interpreted (Fig. 72e & 74). Fringes differ from shadows and tails because the fringe boundary is generally discordant with the foliation in the enclosing matrix (Spry 1983).

6) **Asymmetric Fish Structures** - Variations of this type of structure include the following:

(a) **foliation fish** - Microscopic study reveals fish-shaped ellipsoidal volumes of matrix which form as lithons between C' cleavage zones. The internal foliation and/or layering of the fish in Figure 72f & g lies oblique to the external foliation (C-fabric). The fish have undergone back rotation (Fig. 60 & 68) in response to displacement along flanking C' foliation planes similar to Hanmer's (1986) Type 2B structures. The asymmetry of the foliation fish is antithetic (anti-clockwise) to the bulk sense (clockwise) of shear.

(b) **boulder fish** - Boulder-size conglomeratic clasts may display a fish geometry. Back-rotation of the clasts is opposite in sense to the shear direction and the clasts now lie oblique to the external foliation parallel to the S-fabric (Fig. 72h & 24).

7) **Asymmetrical Pull-Aparts** - Competent, once continuous
quartz veins exhibit pinch-and-swell structures on a mesoscopic scale as a result of heterogeneous layer-parallel extension (Fig. 72i & 75). The swells correspond to the fish structures and lie oblique to the external foliation exhibiting back rotation similar to Hanmer’s (1986) Type 2A structure.

8) **Displaced Broken Grains** – Fragmented rigid grains, typically plagioclase, possess microfractures oriented oblique to the foliation plane. Slip between the fragments (Fig. 72j & 76), due to extension in the ductile matrix, is analogous with a sheared card deck (Etchecopar 1977) and opposite to the overall sense of shear in the rock (Simpson and Schmid 1983).

9) **Asymmetric Active Folds** – Local inhomogeneities in the flow field cause deviant orientations of small sectors of the layering which become amplified into minor asymmetric folds by continuing flow. The axial trace of the folds generally lies in the plane of flattening (S-fabric) and may rotate into parallelism with the C-fabric (Berthé and Brun 1980). These folds are termed active because they were produced during shearing and their asymmetry demonstrates the sense of shear (Fig. 33). Figure 72k & 34 exhibit nucleation of an asymmetric fold around a granitoid clast in a conglomerate. Folding initiates down-flow of the clast and the amplitude dies out along the axial trace.

10) **Tear-Drop Shaped Quartz Grains** – Quartz porphyroclasts display a tear-drop shape in which their tails parallel the
Fig. 75. Quartz veins exhibiting asymmetric pull-apart structure.

Fig. 76. Displaced broken plagioclase grain exhibits slip between fractured fragments (field of view ≈ 2mm).
shear direction. These microstructures resemble the tear-drop shaped quartz grains of Simpson (1983) except they are not restricted to shear zone boundaries. Tear-drop shaped quartz grains preferentially develop where two grains, quartz and/or plagioclase, are separated by a minor amount of matrix, averaging <50 μm wide, suggesting that their spatial relationship obstructs the flow field around them (Fig. 72 l & 77). The tails may become separated and drawn out from the porphyroclasts due to fracturing or dynamic recrystallization eventually forming augen.

11) **Asymmetric Kinkbands** - Experimental studies have shown that antithetic asymmetric kinkbands are produced during simple shear (Hoeppener and Schwarz 1980). Statistical analysis reveals that antithetic kink bands dominate synthetic kink bands by more than two to one. A corresponding ratio was recorded by Wallace and Clifford (1983) and Buck (1986a) in shear zones in the Superior Province. Consequently, collection of a large data set of asymmetric kinkbands permits the deduction of shear sense. The sense of displacement along the dominant kinkband is antithetic to the bulk shear sense (Fig. 72m).

12) **Obliquity of Long Axis** - Oblong porphyroclasts, typically plagioclase, may be oriented with their long axes oblique to the shear plane (C-fabric). The long axis, according to Ramsay and Graham (1970), marks the plane of flattening or S-fabric which rotates into the C-fabric with progressive shear (Fig.
Fig. 77. Tear-drop shaped quartz grain develops as adjacent plagioclase grain obstructs flow field (field of view ≈ 2mm).

Fig. 78. Obliquity of subgrains parallel S-fabric and rotate into C-fabric (field of view ≈ 2mm).
13) Obliquity of Subgrains and Recrystallized Grains - The long axes of subgrains and dynamically recrystallized grains, when not obscured during the development of porphyroclast tails, may lie oblique to the C-fabric, thus defining an S-fabric (Fig. 72o & 78). Simpson and Schmid (1983) state that the orientation of the new grains is approximately perpendicular to the maximum compression direction of the last increment of deformation. Like many of the preceding structures the long axes may rotate into parallelism with the C-fabric.

14) Tension Gashes - En echelon tension gashes originate at an angle of approximately 45 degrees to the shear plane (C-fabric) in a direction perpendicular to the greatest incremental extension produced by simple shear (Ramsay 1967). With progressive shearing a characteristic sigmoidal shape develops from which kinematic sense can be derived (Fig. 72p & 40). Dextral and undeformed tension gashes are present, as well as some with a sinistral geometry. Sigmoidal tension gashes may also form without simple shear (Nicholson and Ejiofor 1987; Pollard et al. 1982) which further questions the reliability of these structures as kinematic indicators.

Field mapping supported by microstructural studies in the PLDZ provide an assemblage of kinematic indicators which exhibit predominantly dextral sense during movement along the PLSZ. The most reliable types give a consistent sense
throughout the PLDZ. Few ambiguous or sinistral kinematic indicators observed in less reliable types, may represent localized sinistral shear or shuffling during a bulk dextral sense event. Dextral displacement of late diabase dikes by the PLF is compatible with shear sense in the PLSZ.
CHAPTER VIII - TECTONIC SIGNIFICANCE OF PLDZ

The PLDZ demarcates a distinct structural and lithological break within the Superior Province of Canada. The PLDZ separates the Beardmore-Geraldton Belt (BGB) to the south from the Onaman-Tashota Terrane (OTT) to the north. The structure and lithology of the BGB and OTT is contrasted below.

VIII.I STRUCTURE OF THE BEARDMORE-GERALDTON BELT

Three east-west trending sedimentary belts separate three volcanic belts forming a relatively linear BGB (Mackasey 1976). The volcanics are mafic in composition, generally massive to pillowied, with tops to the north. Their environment of deposition was predominantly subaqueous. The overlying sediments constitute a coarsening-upward, northward younging sequence. Devaney and Fralick (1985) interpreted this sequence as a prograding clastic wedge whereby a northern source supplied clastic material to fluvial fans, braided rivers and deltas which eventually drained into an aquabasin to the south.

Deformation within the BGB is inhomogeneous. Williams (1986) recognized discrete shear zones and sheared contacts between the lower contact of volcanics and the upper contact of sediments demonstrated by the presence of foliations, lineations and other minor structures similar to those found
within the PLDZ indicating dominantly dextral sense with south-side-down. These layer-parallel brittle-ductile shear zones delineate the Watson-Oxaline Lake Fault and the Blackwater River Fault which parallel the PLDZ. Relatively undeformed rocks are preserved between these deformation zones. The strong linear trend of the homoclinal supracrustals accompanied by the absence of arcuate granitic intrusives, regional fold structures and felsic volcanics contrasts with typical greenstone belts.

VIII.II STRUCTURE OF ONAMAN-TASHOTA TERRANE

The OTT consists of intermediate to felsic volcanics conformably overlying a subordinate mafic sequence (Mackasey and Wallace 1978; Kresz and Zayachivsky 1986). The abundance of pyroclastics in the intermediate sequence indicates a dominantly subaerial environment of deposition. These intermediate volcanics are of calc-alkaline affinity, while the mafic volcanics are tholeiitic (Mackasey and Wallace 1978). Massive to pillowed basalts adjacent to the PLDZ indicate tops to the south, in agreement with younging directions interpreted from volcanic layering by Kresz and Zayachivsky (1986) immediately to the northwest of the study area.

According to Mackasey and Wallace (1978) the volcanics form a broad E-W trending (synclinal) fold to the north of the map area. The orientation of foliation and layering of the
volcanics changes from ENE-WSW near the PLDZ (southern limb) to north at the nose and swings to NW-SE in the north (northern limb). This steep foliation, with an accompanying mineral elongation lineation plunging moderately to the west, is overprinted by foliations associated with the PLDZ and therefore predates the shear event.

The volcanics have been intruded by small subcircular massive stocks of trondhjemitic composition. The Coyle Lake Stock, the only one of the stocks located in the study area, possesses neither a foliation nor a lineation and metamorphism within the intrusive is moderate. Contacts mapped by Mackasey and Wallace (1978) are irregular and narrow where intrusions of the stock cut the volcanics or volcanic inclusions are found in the stock. Map 2 - Foliations shows that the stock often truncates a regional foliation believed to have been produced prior to PLDZ development. Detailed mapping is required north of the PLDZ to better understand the relationship between these plutons and surrounding foliation. The presence of a synclinal fold structure, intermediate to felsic volcanics and arcuate-shaped granitic intrusions is typical of a granite-greenstone terrane.

The structure and lithology of the BGB and OTT are summarized in Table 7.
Table 7. Contrasting structure and lithology in the Beardmore-Geraldton Belt and Onaman-Tashota Terrane.

<table>
<thead>
<tr>
<th>STRUCTURE</th>
<th>OTT</th>
<th>BSB</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regional Structure</td>
<td>E-W trending volcanic belts intruded by arcuate plutons</td>
<td>long, linear E-W trending volcanic and sedimentary belts</td>
</tr>
<tr>
<td>Folding</td>
<td>local, broad E-W trending syncline asymmetric folds restricted to shear zones</td>
<td></td>
</tr>
<tr>
<td>Foliations</td>
<td>pervasive; predominantly layer-parallel</td>
<td>present only along sheared contacts and in discrete layer-parallel</td>
</tr>
<tr>
<td>Lineations</td>
<td>pervasive; moderately plunging (elongation) to west</td>
<td>present only along sheared contacts; a) early = steep; b) late = horizontal (elongation) (stretching)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>LITHOLOGY</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volcanics</td>
</tr>
<tr>
<td>Sediments</td>
</tr>
<tr>
<td>Granitoid Plutons</td>
</tr>
<tr>
<td>S abdomic Intrusives</td>
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<tr>
<td>Younging Directions</td>
</tr>
</tbody>
</table>
VIII.III MOVEMENT ALONG THE PLDZ

Kinematic indicators within the PLDZ indicate dextral motion in a vertical plane with a subhorizontal stretching direction defining a transcurrent or wrench system. Although strike-slip movement dominates during the deformation history of the PLDZ, there exists field evidence for a component of early dip-slip motion. This is preserved by the relationship between C- and C'-fabrics. A northward dipping C'-fabric cuts the vertical C-fabric producing an oblique intersection lineation. According to Coward (1984), in pure strike-slip systems the C'-fabric intersects the C-fabric perpendicular to the mineral lineation and shear transport direction. To meet this requirement in the PLDZ, a vertical C'-fabric must intersect the vertical C-fabric to produce a vertical intersection lineation perpendicular to the subhorizontal mineral lineation. This is not the case. Hence, the PLDZ has a component of differential movement setting up a shear couple in the YZ-plane perpendicular to that of the XY-plane, or shear plane (Coward 1984). This relationship is illustrated in Figure 79.

C- and C'-fabric intersection lineations were rarely measurable in the field due to rare exposure of the XY-plane. However, where the YZ-plane was exposed with the XZ-plane, a plot of C- and C'-fabric orientations produced an intersection lineation parallel to that measured in the field, plunging 50 degrees to 70 degrees to the east. Ductile displacement of the
Differential movement interpreted from C- and C'-fabric relationships.
Fig. 80. Major component of strike-slip (dextral) motion in XZ-plane.

Fig. 81. Minor component of dip-slip (south-side-down) motion in YZ-plane.
C-fabric by the C'-fabric is greater in the XZ-plane (Fig. 80) than in the YZ-plane (Fig. 81). Thus, differential movement is comprised of a major component of strike-slip and a minor component of dip-slip. Dip-slip motion sense is south-side-down. The presence of subhorizontal stretching lineations and slickensides in the shear plane may be explained as a record of the last increment of the strain history which was strike-slip movement.

In summary, early ductile deformation along the PLDZ involved differential movement in a dextral sense with south-side-down. Strike-slip dextral movement prevailed with subsequent deformation and a transition from ductile shearing to brittle faulting followed during uplift. The PLDZ was active from the Archean to Proterozoic but movement was probably not continuous over this time span.

VIII.IV TECTONIC MODEL

Any of the classical models formulated to explain crustal evolution in the Archean (for example Talbot 1968, 1973; Anhaeusser et al. 1969; Glickson 1972; Goodwin 1977; Fyfe 1978; Young 1978) will be hard-pressed to explain the juxtaposition of two terranes which are structurally and lithologically very different. Structures associated with the development of the PLDZ indicate that a NW-SE compressional event was responsible for the deformation along this zone. Thus, a plate tectonic model should be considered an
alternative to explain an intensely deformed contact between two diverse terranes.

Movement along the PLDZ is in agreement with the tectonic model proposed by Williams (1986). This model involves oblique NW-SE collision of an island arc (OTT) to the north and a subducting oceanic crust which leads to the formation of an accretionary prism (BGB) to the south. The tectonic model presented below is based on Williams' model and plate tectonic models produced in Condie (1981, 1983) and Windley (1984). Table 8 summarizes progressive deformation stages during development of the PLDZ. Corresponding illustrations are provided in Figure 82.

Although this model is suitable to the study area, its lateral extent is problematic. To the west, the PLDZ can be traced to Lake Nipigon and the model remains intact. However, the eastern extent of the PLDZ has not been mapped in detail and correlation has been hindered by overburden. Detailed structural studies by Buck (1986a) in the Barton Bay Deformation Zone (BBDZ) to the southeast has revealed structures very similar to those observed in the PLDZ. Dextral shear sense is consistent and the two zones may represent the same structure or at least separate structures produced during the same tectonic event. Yet, the BBDZ does not demarcate the structural and lithological break which the PLDZ does. Thus, the presented model may not be applicable east of the study area.
Table 8. Tectonic model representing progressive deformation stages during development of PLDZ.

<table>
<thead>
<tr>
<th>Stage 1</th>
<th>Archean Oblique Plate Convergence</th>
</tr>
</thead>
<tbody>
<tr>
<td>NW-SE compression during plate motion resulting in oblique plate convergence between island arc (north) and oceanic crust (south)</td>
<td></td>
</tr>
<tr>
<td>Initiation of subduction with differential motion: dextral sense and south-side-down</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stage 2</th>
<th>Volcanism and Plutonism</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dehydration and partial melting of descending slab produces volcanics in an arc system above the slab</td>
<td></td>
</tr>
<tr>
<td>Early extrusives are tholeiitic basalts deposited in subaqueous environment</td>
<td></td>
</tr>
<tr>
<td>Later extrusives are calc-alkaline intermediate-felsic flows and pyroclastics deposited in subaerial environment</td>
<td></td>
</tr>
<tr>
<td>Partial melting of mafic rocks in upper mantle produces granitic melts resulting in trondhjemitic plutonism</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stage 3</th>
<th>Sedimentation and Accretionary Prism</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continued uplift exposes both greenstones and plutons which are eroded and detritus collects in fore-arc basin</td>
<td></td>
</tr>
<tr>
<td>Some sedimentation may be synchronous with volcanism</td>
<td></td>
</tr>
<tr>
<td>Sediment build-up evolves into accretionary prism with continual subduction</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stage 4</th>
<th>Thrust Imbrication</th>
</tr>
</thead>
<tbody>
<tr>
<td>Delamination of basaltic layers from subducting slab are thrust into contact with overlying sediments</td>
<td></td>
</tr>
<tr>
<td>Back-rotation of prism toward arc with continual subduction</td>
<td></td>
</tr>
<tr>
<td>Intrusion of concordant gabbroic lenses in arc and prism</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stage 5</th>
<th>Ductile-Brittle Transition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Termination of subduction process resulting in dominantly strike-slip movement along PLDZ, Watson-Oxaline Lake Fault and Blackwater River Fault</td>
<td></td>
</tr>
<tr>
<td>Transition from ductile to brittle deformation with continual uplift</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stage 6</th>
<th>Proterozoic Extension</th>
</tr>
</thead>
<tbody>
<tr>
<td>Extensional event during Proterozoic enables intrusion of late diabase dikes</td>
<td></td>
</tr>
<tr>
<td>Dextral, brittle displacement of diabase dikes indicative of discontinuous movement along PLDZ</td>
<td></td>
</tr>
</tbody>
</table>

Note: Stages 2, 3 and 4 may be synchronous.
Fig. 92A. Tectonic model (see text for descriptions).
Fig. 82B. Tectonic model (see text for descriptions).
To date, an island arc-accretionary prism model best accounts for the geology of the study area. Similar island arc models for the Superior Province have been proposed in the past by Langford and Morin (1976), Blackburn (1980) and Dimroth et al. (1983). Presumably, if accretionary prisms were developed during subduction in these examples, they were not preserved.

Mackasey (1976) included the study area in the Wabigoon Subprovince belt and defined the Wabigoon-Quetico boundary as the contact between the most southern mafic volcanic belt of the BGB and the sedimentary sequence to the south. Structural and lithological changes across the PLDZ presented in this study favour the PLDZ as a fundamental boundary. Devaney and Williams (manuscript) have proposed that the BGB represents a transitional zone between the Wabigoon Subprovince to the north and the Quetico Subprovince to the south.
CHAPTER IX - DISCUSSION

Detailed structural mapping in the study area has featured a varied assemblage of structures associated with development of the PLDZ. Some originated during brittle deformation, others during ductile deformation. These structures are believed to represent one regional deformation event during which movement along the zone commenced in the Archean and terminated in the Proterozoic. Displacement was probably not continuous over this time. An overall transition from ductile to brittle deformation may be explained by continual uplift, however, localized changes in behaviour can result from increases in fluid pressure which enhance brittle behaviour by reducing effective normal stress (Stott and Schneiders 1983; Henderson 1983). Thus, superimposed ductile and brittle structures may be generated by repeated increase and subsequent release of fluid pressure.

A summary of important structures associated with the PLDZ is given in Table 9. Parameters considered for each type of structure include: 1) mean orientation; 2) frequency of observations; 3) spatial relationship to centre of PLDZ; 4) temporal relationship during PLDZ formation; and 5) degree of development.
Table 9. Summary of important structures associated with the PLDZ.

<table>
<thead>
<tr>
<th>STRUCTURE</th>
<th>FOLIATIONS</th>
<th>LINEATIONS</th>
<th>FOLDING</th>
<th>KINK/BANDS</th>
<th>TENSION GASHES</th>
<th>FAULT/FRACTURES</th>
</tr>
</thead>
<tbody>
<tr>
<td>TYPE</td>
<td>1) S-fabric</td>
<td>1) stretching</td>
<td>1) asymmetric</td>
<td>1) dextral</td>
<td>1) dextral</td>
<td>1) wrench fault (PLF)</td>
</tr>
<tr>
<td></td>
<td>2) C-fabric</td>
<td>2) slickenside</td>
<td>a) Z-shape</td>
<td>2) sinistral</td>
<td>2) sinistral</td>
<td>2) step fractures</td>
</tr>
<tr>
<td></td>
<td>3) C*-fabric</td>
<td>3) intersection</td>
<td>b) S-shape</td>
<td>3) conjugate</td>
<td>3) undeformed</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4) conjugate</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MEAN ORIENTATION</td>
<td>1) ENE/90</td>
<td>1) 04/243</td>
<td>1a) 039?</td>
<td>1) 020/90</td>
<td>1) variable</td>
<td>1) EW-trending</td>
</tr>
<tr>
<td></td>
<td>2) EW/90</td>
<td>2) horizontal</td>
<td>b) -----</td>
<td>2) 124/90</td>
<td>2) variable</td>
<td>2) vertical</td>
</tr>
<tr>
<td></td>
<td>3) 279/70</td>
<td></td>
<td></td>
<td>3) 50-70/270</td>
<td>3) 021121/90</td>
<td>3) variable</td>
</tr>
<tr>
<td></td>
<td>4) 0506/97/90</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>FREQUENCY OF OBSERVATIONS</td>
<td>1) common</td>
<td>1) common</td>
<td>1a) rare</td>
<td>1) common</td>
<td>1) rare</td>
<td>1) abundant</td>
</tr>
<tr>
<td></td>
<td>2) abundant</td>
<td>2) common</td>
<td>b) absent</td>
<td>2) abundant</td>
<td>2) rare</td>
<td>2) rare</td>
</tr>
<tr>
<td></td>
<td>3) abundant</td>
<td>3) rare</td>
<td></td>
<td>3) rare</td>
<td>3) rare</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4) rare</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SPATIAL RELATIONSHIP TO CENTRE</td>
<td>1) distal</td>
<td>1) proximal</td>
<td>a) proximal</td>
<td>1) proximal</td>
<td>1) distal</td>
<td>1) proximal</td>
</tr>
<tr>
<td></td>
<td>2) proximal</td>
<td>2) proximal</td>
<td>b) -----</td>
<td>2) proximal</td>
<td>2) distal</td>
<td>2) proximal</td>
</tr>
<tr>
<td></td>
<td>3) proximal</td>
<td>3) proximal</td>
<td></td>
<td>3) proximal</td>
<td>3) distal</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4) distal?</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TEMPORAL RELATIONSHIP RELATIVE TO SHEARING</td>
<td>1) early</td>
<td>1) mid-late</td>
<td>a) mid-late</td>
<td>1) late-post</td>
<td>1) early-late</td>
<td>1) post</td>
</tr>
<tr>
<td></td>
<td>2) mid</td>
<td>2) late-post</td>
<td>b) -----</td>
<td>2) late-post</td>
<td>2) early-late</td>
<td>2) late-post</td>
</tr>
<tr>
<td></td>
<td>3) late</td>
<td>3) late</td>
<td></td>
<td>3) late-post</td>
<td>3) early-late</td>
<td></td>
</tr>
<tr>
<td>DEGREE OF DEVELOPMENT</td>
<td>1) good</td>
<td>1) poor</td>
<td>a) good</td>
<td>1) excellent</td>
<td>1) good</td>
<td>1) excellent</td>
</tr>
<tr>
<td></td>
<td>2) excellent</td>
<td>2) good</td>
<td>b) ----</td>
<td>2) excellent</td>
<td>2) good</td>
<td>2) good</td>
</tr>
<tr>
<td></td>
<td>3) excellent</td>
<td>3) good</td>
<td></td>
<td>3) excellent</td>
<td>3) good</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4) good</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
A proximal relationship to the centre of the PLDZ is defined as one where structures lie within the internal boundary of the zone, that is approximately 100 to 200 m from the centre. Distal structures lie between the internal and external boundaries of the PLDZ.

Documentation of these structures not only identifies the extent of deformation zones in the field, but also provides a wealth of information which facilitates a greater understanding of deformation parameters responsible for their development. Complemented with a detailed microstructural analysis, the above structures permit a good approximation of the following deformation parameters:

1) shear direction (strike slip/dip slip)
2) sense of displacement (dextral/sinistral)
3) type of deformation (ductile/brittle)
4) type of strain (pure shear/simple shear)
5) stress directions ($\sigma_1$, $\sigma_2$, $\sigma_3$)
6) amount of strain
7) total displacement

Approximation can only be made of the dominant type of deformation parameter as more than one type was probably active during the deformation history of a zone. For example, a minor amount of pure shear may accompany bulk simple shear, a predominantly strike-slip motion was preceded by an early dip-slip component, ductile structures give way to brittle structures and a bulk dextral displacement dominates with
possibly localized sinistral displacement. Because of the size of the PLDZ and the presence of a complex secondary strain gradient, calculation of the amount of strain and total displacement along the zone are speculative. Determination of a maximum stress direction can not be defined as a single line of directed pressure but can be restricted to roughly a NW-SE direction. Thus, the deformation parameters responsible for the PLDZ may not be constant over time and the relative timing of the parameters during the deformation history can only be resolved with caution.

Major faults have been recognized elsewhere in the Superior Province since the earliest mapping of the Canadian Shield and more recently examined in detail. Some examples are: Wabigoon Fault (Breaks et al. 1978); Quetico Fault (Poulsen et al. 1980); Bankfield-Tombill Fault (Buck 1986a); Sydney Lake Fault (Stone 1977); and Lake St. Joseph Fault (Stott 1985). These regional structures share many characteristics with the PLDZ including:

1) E-W trending lineament
2) up to and exceeding 100 km in length and up to a few kilometres in width
3) steep to vertical dip
4) stretching lineation is statistically horizontal
5) dominantly strike-slip movement
6) dominantly dextral sense displacement
7) mylonitic foliation subsequently faulted
8) occur along or near subprovince boundary
9) indicate a NW-SE regional compression

Another group of faults occur within subprovinces forming a NW and NE conjugate set and typically exhibit dextral and sinistral displacement respectively, as observed in the Pipestone-Cameron Fault (Buck 1986b) and the Manitou Straits Fault (Wallace and Clifford 1983). These possess structures which are similar to the E-W trending faults.

Recently some of these regional scale faults have been redefined as deformation zones because they consist of much more than a mere fault structure. These major structures must be examined in greater detail in an attempt not only to document their deformation history, but to determine their role in the evolution of the Archean in the Superior Province of Canada.

Many deformation zones, ranging in scale from kilometres to centimetres, have also served as channelways created by structurally induced permeability. These zones may be conducive to mineralization and represent good exploration targets. Detailed structural mapping in the PLDZ reveals the dominance of ductile structures, especially shear fabrics, within the central portion of the zone. Brittle structures spatially related to the shear fabrics are manifested as kinkbands, whereas quartz veining is rare and typically restricted to an area outside of the central zone of intense deformation. Quartz veins may have been obliterated due to
ductile deformation or more probably they were not produced as ductile deformation proceeded without rupture of deforming rocks, and open fissures may not have formed.

Retrograde mineral assemblages observed in conglomerate, and intermediate and mafic volcanics demonstrate that a fluid phase was introduced into the PLDZ to produce hydration reactions and that the fluid flow was focussed along the shear fabrics. According to Colvine et al. (1984) such fluids can generate widespread, disseminated alteration and mineralization in a ductile environment, rather than fissure filling vein systems. However, they conclude in presenting an integrated model for the origin of Archean lode gold deposits that (Colvine et al. 1984, p. 23)

"In most cases gold concentrations occur in direct association with quartz and sulphides where these minerals are present as vein constituents or zones of replacement. Otherwise, carbonate veins and carbonate, white mica-bearing rocks are not ore grade."

Thus, the absence of significant quartz and sulphide species and the abundance of carbonate, white mica-bearing rocks within the PLDZ may explain the observed absence of gold concentration. Nevertheless, the presence of ore grade gold concentration along the Brookbank Fault, a splay of the PLDZ to the west of the study area, and the unknown structure of the PLDZ at the bottom of Paint Lake may warrant further exploration. Hence, these zones of deformation are of both academic and economic interest.
This study has shown that the PLDZ consists of an assemblage of structures which upon examination augments our understanding of its deformation history and its tectonic significance within a regional framework. Conclusions drawn from this structural analysis are listed in Chapter X.
1) The PLDZ demarcates a major structural and lithological break between the OTT to the north and the BGB to the south, within the Superior Province of the Canadian Shield. 
2) The PLDZ is comprised of an early ductile component, termed the Paint Lake Shear Zone, and a late brittle component known as the Paint Lake Fault. 
3) Simple shear deformation dominated as a result of a regional NW-SE compressional event. There exists evidence for a minor component of pure shear. 
4) Movement along the PLDZ is in a dextral sense consisting of an early differential motion with south-side-down and a later strike-slip motion. PLDZ was active from Archean to mid Proterozoic times and movement was discontinuous during this time. 
5) Total displacement is difficult to accurately quantify, however, brittle displacement of diabase dikes equals approximately 5 kilometres and an estimate in the order of tens of kilometres is not unrealistic during ductile displacement. 
6) Strain profiling demonstrates the presence of an outer primary strain gradient and an inner secondary strain gradient within the PLDZ. 
7) Although the locus of the PLDZ may in part be lithologically controlled, mylonitization which accompanied
shear zone development is not dependent on lithological type. Conglomerate, intermediate and mafic volcanic units exhibit similar mesoscopic and microscopic deformation structures when transected by the PLDZ.

8) Microstructural analysis reveals a change in the dominant microstructure deformation mechanism from dislocation creep to diffusion creep with increasing strain during mylonitization.

9) Regionally metamorphosed lithologies of lower greenschist facies outside the PLDZ undergo retrograde metamorphic reactions during deformation within the PLDZ.


Bruce, E.L. 1937. The eastern part of the Sturgeon River Area (Jellicoe-Sturgeon River Section). Ontario Department of Mines, 45 Part.2, pp. 1-59.


Laird, H.D. 1937. The western part of the Sturgeon River area (Sturgeon River - Beardmore Section). Ontario Department of Mines, 45, part 2, pp. 61-117.


APPENDIX I

STRAIN PROFILES
PROFILE A–B

n = 49
displacement = 51

SHEAR STRAIN

DISTANCE

sz centre

1.20 km
PROFILE C-D

n = 45

displacement = 30

SHEAR STRAIN

DISTANCE

sz centre

0.96 km
PROFILE E-F

n = 38
displacement = 49

distance = 49

displacement = 49

n = 38

DISTANCE

1.40 km
APPENDIX II

FREQUENCY HISTOGRAMS
CONGLOMERATE - LENGTH (mm) QUARTZ GRAINS

DOMAIN 1
N = 49
MEAN = 0.442
VAR. = 0.050
S.D. = 0.223

DOMAIN 2
N = 48
MEAN = 0.438
VAR. = 0.031
S.D. = 0.176

DOMAIN 3
N = 49
MEAN = 0.408
VAR. = 0.027
S.D. = 0.166

DOMAIN 4
N = 48
MEAN = 0.370
VAR. = 0.049
S.D. = 0.221

DOMAIN 5
N = 49
MEAN = 0.280
VAR. = 0.010
S.D. = 0.100
CONGLOMERATE - WIDTH (mm) QUARTZ GRAINS

DOMAIN 1
N = 49
MEAN = 0.249
VAR. = 0.017
S.D. = 0.131

DOMAIN 2
N = 48
MEAN = 0.219
VAR. = 0.009
S.D. = 0.097

DOMAIN 3
N = 49
MEAN = 0.192
VAR. = 0.007
S.D. = 0.085

DOMAIN 4
N = 48
MEAN = 0.218
VAR. = 0.030
S.D. = 0.174

DOMAIN 5
N = 49
MEAN = 0.141
VAR. = 0.005
S.D. = 0.074
CONGLOMERATE - LENGTH/WIDTH RATIO QUARTZ GRAINS

Frequency Histogram

DOMAIN 1
N = 49
MEAN = 1.88
VAR. = 0.41
S.D. = 0.64

Frequency Histogram

DOMAIN 2
N = 48
MEAN = 2.11
VAR. = 0.41
S.D. = 0.64

Frequency Histogram

DOMAIN 3
N = 49
MEAN = 2.27
VAR. = 0.58
S.D. = 0.76

Frequency Histogram

DOMAIN 4
N = 48
MEAN = 2.15
VAR. = 0.90
S.D. = 0.95

Frequency Histogram

DOMAIN 5
N = 49
MEAN = 2.35
VAR. = 1.55
S.D. = 1.25
CONGLOMERATE - LENGTH (mm) PLAGIOCLASE GRAINS

**DOMAIN 1**

- N = 39
- MEAN = 0.471
- VAR. = 0.049
- S.D. = 0.221

**DOMAIN 2**

- N = 41
- MEAN = 0.439
- VAR. = 0.017
- S.D. = 0.129

**DOMAIN 3**

- N = 39
- MEAN = 0.410
- VAR. = 0.011
- S.D. = 0.103

**DOMAIN 4**

- N = 42
- MEAN = 0.445
- VAR. = 0.042
- S.D. = 0.206

**DOMAIN 5**

- N = 30
- MEAN = 0.145
- VAR. = 0.003
- S.D. = 0.051
CONGLOMERATE - WIDTH (mm) PLAGIOCLASE GRAINS

DOMAIN 1
N = 39
MEAN = 0.310
VAR. = 0.030
S.D. = 0.172

DOMAIN 2
N = 41
MEAN = 0.256
VAR. = 0.015
S.D. = 0.121

DOMAIN 3
N = 39
MEAN = 0.203
VAR. = 0.004
S.D. = 0.066

DOMAIN 4
N = 42
MEAN = 0.213
VAR. = 0.011
S.D. = 0.104

DOMAIN 5
N = 30
MEAN = 0.068
VAR. = 0.010
S.D. = 0.031
CONGLOMERATE - LENGTH/WIDTH RATIO PLAGIOCLASE GRAINS

**Frequency Histogram**

**DOMAIN 1**
- N = 39
- MEAN = 1.73
- VAR. = 0.70
- S.D. = 0.84

**DOMAIN 2**
- N = 41
- MEAN = 2.39
- VAR. = 8.49
- S.D. = 2.91

**DOMAIN 3**
- N = 39
- MEAN = 2.54
- VAR. = 5.59
- S.D. = 2.36

**DOMAIN 4**
- N = 41
- MEAN = 2.29
- VAR. = 0.94
- S.D. = 0.97

**DOMAIN 5**
- N = 30
- MEAN = 2.45
- VAR. = 1.56
- S.D. = 1.25
VOLCANIC - LENGTH (mm) PLAGIOCLASE CLASTS

**DOMAIN 1**
- N = 43
- MEAN = 0.932
- VAR. = 0.186
- S.D. = 0.431

**DOMAIN 2**
- N = 49
- MEAN = 0.381
- VAR. = 0.018
- S.D. = 0.134

**DOMAIN 3**
- N = 49
- MEAN = 0.200
- VAR. = 0.006
- S.D. = 0.074

**DOMAIN 4**
- N = 49
- MEAN = 0.093
- VAR. = 0.009
- S.D. = 0.030

**DOMAIN 5**
- N = 48
- MEAN = 0.176
- VAR. = 0.019
- S.D. = 0.138
VOLCANIC - WIDTH (mm) PLAGIOCLASE CLASTS

DOMAIN 1
N = 43
MEAN = 0.508
VAR. = 0.056
S.D. = 0.237

DOMAIN 2
N = 49
MEAN = 0.212
VAR. = 0.005
S.D. = 0.072

DOMAIN 3
N = 49
MEAN = 0.103
VAR. = 0.001
S.D. = 0.036

DOMAIN 4
N = 49
MEAN = 0.050
VAR. = 0.003
S.D. = 0.018

DOMAIN 5
N = 48
MEAN = 0.101
VAR. = 0.006
S.D. = 0.074
VOLCANIC - LENGTH/WIDTH RATIO PLAGIOCLASE CLASTS

**DOMAIN 1**
- $N = 43$
- $\text{MEAN} = 1.90$
- $\text{VAR.} = 0.35$
- $\text{S.D.} = 0.59$

**DOMAIN 2**
- $N = 49$
- $\text{MEAN} = 1.93$
- $\text{VAR.} = 0.36$
- $\text{S.D.} = 0.60$

**DOMAIN 3**
- $N = 49$
- $\text{MEAN} = 1.98$
- $\text{VAR.} = 0.26$
- $\text{S.D.} = 0.51$

**DOMAIN 4**
- $N = 49$
- $\text{MEAN} = 1.97$
- $\text{VAR.} = 0.23$
- $\text{S.D.} = 0.48$

**DOMAIN 5**
- $N = 48$
- $\text{MEAN} = 1.81$
- $\text{VAR.} = 0.14$
- $\text{S.D.} = 0.38$
VOLCANIC - LENGTH (mm) PLAGIOCLASE MATRIX

<table>
<thead>
<tr>
<th>Domain 1</th>
<th>Domain 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>N = 49</td>
<td>N = 63</td>
</tr>
<tr>
<td>MEAN = 0.086</td>
<td>MEAN = 0.028</td>
</tr>
<tr>
<td>VAR. = 0.001</td>
<td>VAR. = 0.006</td>
</tr>
<tr>
<td>S.D. = 0.034</td>
<td>S.D. = 0.008</td>
</tr>
</tbody>
</table>

VOLCANIC - WIDTH (mm) PLAGIOCLASE MATRIX

<table>
<thead>
<tr>
<th>Domain 1</th>
<th>Domain 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>N = 49</td>
<td>N = 63</td>
</tr>
<tr>
<td>MEAN = 0.037</td>
<td>MEAN = 0.017</td>
</tr>
<tr>
<td>VAR. = 0.001</td>
<td>VAR. = 0.001</td>
</tr>
<tr>
<td>S.D. = 0.009</td>
<td>S.D. = 0.015</td>
</tr>
</tbody>
</table>
VOLCANIC - LENGTH/WIDTH RATIO PLAGIOCLASE MATRIX

Frequency Histogram

<table>
<thead>
<tr>
<th>Frequency</th>
<th>1.2</th>
<th>2.0</th>
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<td>12</td>
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<tr>
<td>8</td>
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<tr>
<td>4</td>
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</table>

LAMATRIXD1

Frequency Histogram

<table>
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LAMATRIXD2

DOMAIN 1
N = 49
MEAN = 2.28
VAR. = 0.60
S.D. = 0.77

DOMAIN 2
N = 63
MEAN = 1.93
VAR. = 0.38
S.D. = 0.62