STRUCTURE, METAMORPHISM AND STRATIGRAPHY OF ALLOCHTHONOUS UNITS OF THE SOUTHERN EXMOUTH ANTIFORM, WOPMAY OROGEN, NORTHWEST TERRITORIES.

by

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ABSTRACT

The area of study is a small area of allochthonous rocks structurally located immediately above an Archean antiform in the Wopmay Orogen of the Northwest Territories. The rocks of the study area represent various units of the Akaitcho Group epicontinental rift fill deposits. All rocks are allochthonous, lying structurally above basement and autochthonous cover.

Within the allochthonous rocks, three phases of deformation and a metamorphic culimnation are evident. The first episode (D1) involved thrusting and folding of the allochthon over the Slave Craton, coinciding with peak metamorphic conditions. This phase of deformation was followed by an episode of coaxial, thick-skinned folding (D2), producing the large scale folds of basement and cover witnessed in the area. A late cross-folding event (D3), has provided up to 6 km of structural relief in the study area. The deformational history thus recorded in the Wopmay Orogen is similar to that documented in other orogenic belts such as the eastern Alps and Cape Smith Belt in Quebec.

Metamorphism in the study area is hot side up Buchan type. A progression is observed in pelitic assemblages from muscovite-sillimanite grade to sillimanite-K feldspar grade, from east to west across the study area. A retrogression has produced sillimanite quartz pods (faserkiesel), its

occurrence has permitted a relocation of the pre-existing prograde isograd.

These finds are consistent with the current tectonic model for the area.

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CHAPTER 1 : INTRODUCTION

1.1 Purpose of Study

Within the Metamorphic Internal Zone of the early Proterozoic Wopmay Orogen in north-western North West Territories are areas of complexly deformed allochthonous rocks which have been thrust over the Slave craton [McGlynn, 1970; Fraser et al., 1972; Hoffman, 1980]. While four consecutive, though temporally discrete stages of deformation have been recognized in this zone, polydeformed areas exist within the allochthon, where the structural and metamorphic nature need to be documented in greater detail and reconciled with the regional tectonic model for the orogen. In 1986, officers of the Geological Survey of Canada, as part of ongoing research in the orogen, examined the contrasting styles of basement deformation in the Metamorphic Internal Zone. As part of this work, the basement core of the Exmouth Antiform was examined in detail, as well as the overlying autochthon and allochthon. The study area is located in the allochthonous rocks here.

This thesis study, based on field observations made during several weeks of the 1986 field season, and subsequent laboratory investigations, focusses on structural style and metamorphic grade of an area approximately 10 x 3 km, made up of allochthonous rift deposits located south of the

basement core, just above a sole thrust. The objective of this study is to determine the nature of the polyphase deformation pattern and associated metamorphism first recognized by a G.S.C. field party in 1983.
Maps produced show the allochthon as well as the basement and autochthon. While not acquired by the author, details for the basement and autochthon are essential for the documentation of the study area, and have therefore been included.

1.2 Location and Access

Wopmay Orogen is located in the North West Territories. The orogen covers an area roughly 550 km long and 280 km wide between Coronation Gulf, Great Bear Lake and Great Slave Lake. The Coronation Supergroup rift deposits are located in the eastern part of the exposed orogen. This study concentrates on these rift deposits as found between 65 deg. 01 min. N and 65 deg. 16 min. N latitude and 115 deg. 44 min. E and 115 deg. 51 min. E longitude {FIG 1}.

The field work was done during the summer of 1986, when the author was employed as a Field Assistant by the G.S.C., under the direction of Dr. Janet King

The logistical base for the study was Yellowknife, N.W.T., to which road access is possible from Edmonton during winter and summer months. The study area, where a base camp was established, was reached by scheduled float plane service from Yellowknife. Daily traversing was by water access and by foot.

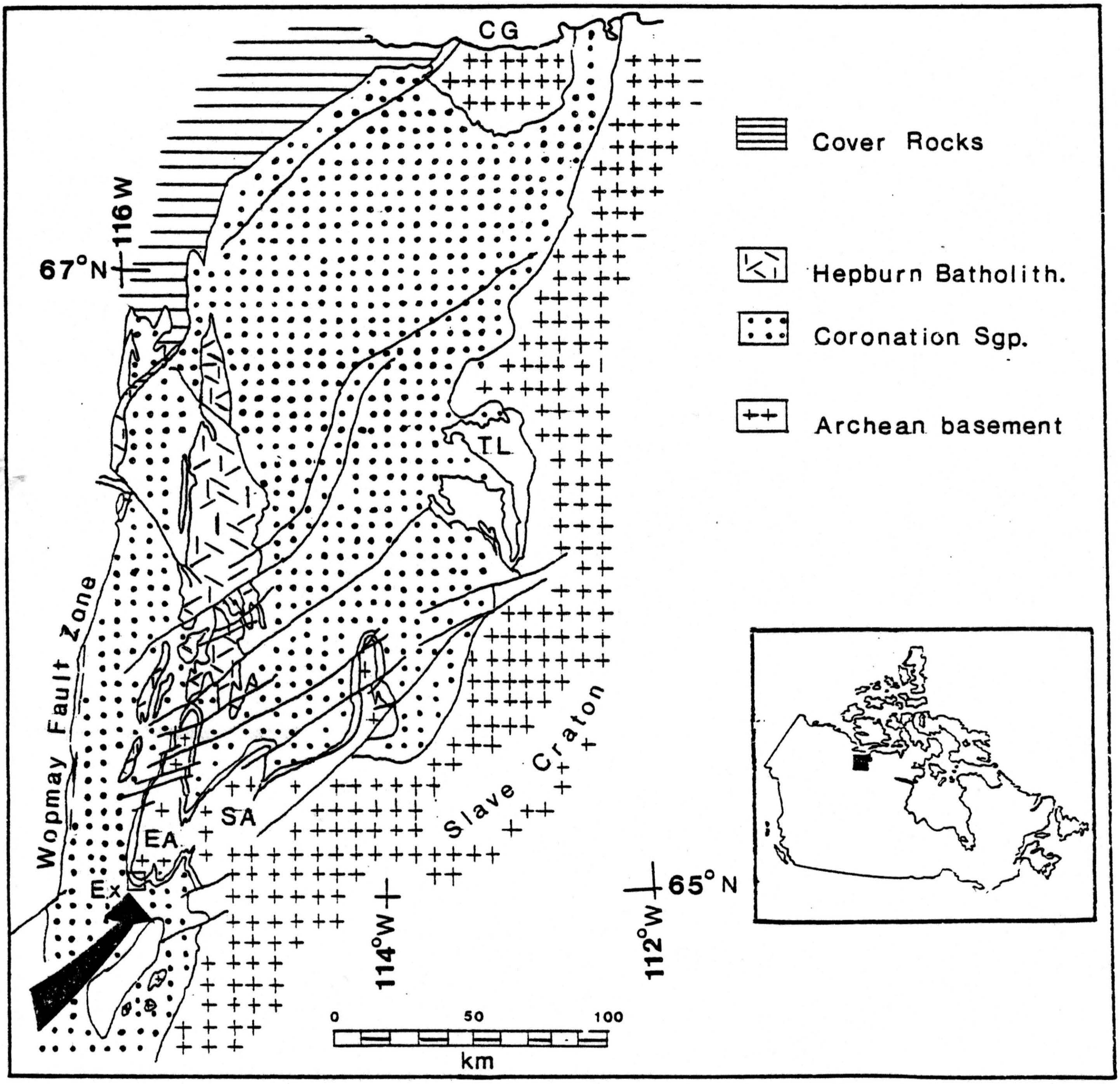
1.3 Previous Work

Until 1959, the Wopmay Orogen had never been mapped geologically at any scale. In 1965, the stratigraphical, structural and intrusive history of the area was worked out very generally by Fraser [Hoffman et al., 1986] in a follow up to reconnaissance mapping. Much work was carried out in the early 1970's by G.S.C. parties in various sections of the orogen. Since 1977, systematic mapping at 1:100,000 scale has produced well constrained geological maps and cross sections of the northern 300 km of the orogen, as well as comprehensive studies of structure, stratigraphy, igneous and metamorphic petrology and geochronology [Hoffman et al., 1986].

Regional mapping of the Redrock Lake map areas (86 G), containing the Exmouth Antiform and the study area was completed in 1983 at 1:100,000 scale. It was at this time that the basic stratigraphy, structure, and metamorphic nature of the area were worked out. Mapping in the 1986 field season was completed at 1:28,000 scale.

FIG 1.1: Location map

Simplified geological map of the northeastern part of the Wopmay Orogen. Ex, is in the Exmouth Lake area; CG - Coronation Gulf; EA - Exmouth Antiform; SA - Scotstoun Antiform; TL - Takujug Lake. Study area is enclosed in box indicated by large arrow. Modified from King [1986].



CG
Cover Rocks
Hepburn Batholith.
Coronation Sgp.
Archean basement
116W
67°N
TL
Wopmay Fault Zone
Slave Craton
SA
EA
Ex
114°W
112°W
65°N
0
50
100
km

CHAPTER 2 : GENERAL GEOLOGY

2.1 The Setting - Wopmay Orogen

Wopmay Orogen is a 1.95 to 1.85 Ga, early Proterozoic (Aphebian) belt [Hoffman and Bowring, 1984], located on the westernmost flank of the Archean Slave Craton {FIG 1.1}. It covers an identified, exposed area of 1.54 x 10^5 km^2.

The orogen is made up of three major tectonic zones which are, from east to west; (1) the Coronation Supergroup, (2) the Great Bear Magmatic Zone and (3) the Hottah Terrane. The present tectonic model for the Wopmay Orogen is that it represents the rapid opening and closing of an ensialic backarc (intracontinental) basin on the westernmost margin of the then juvenile Slave procraton [Hoffman et al., 1986].

The Coronation Supergroup {FIG 1.1}, located on the easternmost part of the Wopmay Orogen, is a west facing depositional prism, part of which has been shortened, thickened, and transported eastward over the Slave Craton, as a result of the 1885 Ma Calderian Orogeny [Tirrul, 1982; Hoffman et al., 1983; St. Onge et al., 1984; King, 1986]. In the easternmost part of this zone, the Coronation Supergroup is autochthonous, lying unconformably on the Slave Craton and comprising what is known as the Asiak Thrust-Fold Belt. West of the Asiak Thrust-Fold Belt in the Coronation Supergroup, and displaying a markedly different structural style is the polydeformed and intensely metamorphosed allochthonous Hepburn Metamorphic Internal Zone. The Hepburn Metamorphic Internal Zone is distributed through the westernmost Coronation Supergroup infolded with the underlying Slave Craton, and is truncated to the west by a steeply dipping mylonitic and brittle fault belt known as the Wopmay Fault Zone [Hoffman et al., 1978; Hoffman, 1980, 1984a; St. Onge et al., 1982, 1983, 1984; King, 1986]. Intruded into all allochthonous units of the Coronation Supergroup are syntectonic to late tectonic discordant plutons of the Hepburn and Wentzel Batholiths {FIG 1.1}. These are dominated by peraluminous biotite granite and tonalite, but also include some gabbroic bodies [Lalonde, 1984]. All units of the Coronation Supergroup were emplaced between 1.9 Ga and 1.85 Ga [Hoffman, 1986].

The westernmost exposed tectonic member of the orogen is the Hottah Terrane [McGlynn, 1976; Hildebrand, 1981]. At first interpreted to be an exotic terrane, it has recently been reinterpreted as an indigenous Andean-type continental magmatic arc built on crust adjacent to the Archean Slave Craton [Hildebrand and Roots, 1985]. A complex of amphibolite facies metasedimentary and intermediate metavolcanic rocks dominate the terrane and these have been intruded by 1914 to 1902 Ma calc-alkaline plutons [Hildebrand et al., 1986].

Between the Hottah Terrane to the west and the Coronation Supergroup to the east is the Great Bear Magmatic Zone. Interpreted as an 1875 to 1849 Ma volcano-plutonic complex [Bowring and Van Schmus, 1986], this zone unconformably overlies the Hottah Terrane, and partly overlies the Coronation Supergroup as well as being in fault contact with it along the Wopmay Fault Zone [Hoffman and McGlynn, 1977; Hildebrand, 1983; Hildebrand et al., 1983; Hildebrand and Bowring, 1984].

2.2 The Coronation Supergroup

As the study area lies entirely within the Coronation Supergroup [St. Onge et al., 1984], it is necessary to discuss in greater detail the nature of the Coronation Supergroup in order to understand the complex stratigraphy encountered within the study area. The stratigraphy of the study area will also be discussed in more detail in a later section (v. section 2.5).

The Coronation Supergroup is made up of three stratigraphically distinct assemblages; the Recluse Group, the Epworth Group and the Akaitcho Group. Each group has several subdivisions and the revised nomenclature for the formations and groups making up the Coronation Supergroup has been presented in Hoffman [1981].

The Akaitcho Group is the most important of the three groups for the purpose of this thesis; the rocks in the map area are entirely of this group. Dated, in its western part ca 1900 Ma [S.A. Bowring and W.R. Van Schmus, unpublished manuscript, 1986] the Akaitcho is characterized by metasedimentary and bimodal metavolcanic rocks [Easton, 1981c, 1982]. The rocks making up the Akaitcho include arkoses, feldspathic quartzite or pelites, with gabbroic sills, local basalt-rhyolite volcanic complexes (pillowed and tuffaceous basalt and subalkaline rhyolite flows, tuffs and sills) as well as local conglomerates bearing clasts of granite-gneiss [Easton 1980, 1981a,b,c, 1982, 1983]. The key to their interpretation as Akaitcho rocks is that all these rocks support the interpretation that they are epi-continental rift-fill deposits [Easton, 1980, 1981c] in a back arc setting [Hildebrand et al., 1986a,b]. Regional stratigraphic correlations are difficult in the Akaitcho group due to the complex structure of the area [King, 1986]. Correlation to any great extent was not possible in the map area due to its intensely deformed nature and therefore the stratigraphic sequence defined for the Akaitcho Group [Easton, 1981c, 1982] is not applicable here.

To the east of the Akaitcho Group is the Epworth Group, a stratigraphically distinct sequence in the Coronation Supergroup. It is a west-facing passive margin sequence of wave dominated shelf facies and contiguous slope rise facies

[Hoffman et al., 1983, 1984]. It is made up of two major groups; the lower Odjick Formation of mature siliciclastic rocks and the upper dolomitic Rocknest Formation. Shelf facies units are continuous for over 200 km in the Northern Wopmay Orogen [Hoffman et al., 1983].

The Recluse Group is the youngest stratigraphic unit of the Coronation Supergroup, and is an eastward-migrating synorogenic foredeep sequence [Hoffman et al., 1986]. The Recluse Group, conformably overlying the Epworth Group, is comprised of hemipelagic shales overlain by turbidite graywackes [Hoffman, 1980; Hoffman and Bowring, 1984]. This group implies a sudden sinking of the shelf and infilling by flysch deposits derived from a north-westerly plutonic-metamorphic source [Jeletsky, 1974; Hoffman, 1980].

The Coronation Supergroup provides a good depositional record of the opening and closing of a marginal basin [Hoffman et al., 1986]. The basin is thought to have evolved during a short period of stretching and contraction of continental crust behind a long-lived magmatic arc,(now the Hottah Terrane [Hildebrand et al., 1986]), which was generated above an east dipping subduction zone. Small Archean inliers of basement within the Akaitcho Group and the bimodal rift stage magmatism indicate that the Coronation Supergroup was deposited on a partially melted continental crust [Easton, 1981, 1983]. During closure of the marginal basin in the Calderian Orogeny, caused by the Hottah Terrane converging on the Slave Craton [Hoffman et al., 1986; Hildebrand et al., 1983]. The Coronation Supergroup was thickened and shortened against the craton. Additional results of the Calderian Orogeny were the collapse and infilling of the foredeep [Hoffman, 1980], the emplacement of syntectonic intrusions in the allochthon [St. Onge, 1981, 1984b,c] and the late folding of autochthonous basement [King, 1985].

2.3 <u>Regional Structure</u>

There were four deformation events in the Wopmay Orogen. The first two made up the Calderian Orogeny and are dated at 1.89 to 1.87 Ga; the third, known as the Tree River Deformation has been dated at 1.88 to 1.69 Ga; and the last deformation episode, which resulted in transcurrent faulting, has been dated at 1.84 to 1.69 Ga [Hoffman and Bowring, 1984; Bowring et al., 1984; Bowring and Van Schmus, unpublished manuscript].

The Calderian Orogeny was comprised of two separate deformations, D1 and D2, which were thin and thick-skinned deformations respectively [Hoffman et al., 1986]. The third episode of deformation, the Tree River Deformation, produced regional cross folds and the last deformation episode produced a system of conjugate transcurrent faults [Hoffman et al., 1986]. The manner in which these deformations were accommodated in the rocks varies considerably across the orogen. As the study area involves only part of the Akaitcho Group (subdivision of the Coronation Supergroup) thrust over the Slave Craton in the Hepburn Metamorphic Internal Zone, only structures related to this area will be discussed.

In general, D1 structures produced north-trending thrusts and folds in the Calderian allochthonous rocks above a sole thrust located 100 to 300 m stratigraphically above the Archean basement [Hoffman et al., 1986]. The allochthon steps up through 6 km of strata above a major ramp located west of the Exmouth antiform [Hoffman et al., 1986]. Isoclinal folds with a eastward vergence predominate in the Calderian allochthon at intermediate to deep structural levels [Hoffman et al., 1986]. This is the D1 folding style witnessed in the study area. In contrast are the D1 structures found in allochthonous strata structurally higher and to the east where upright open D1 folds dipping steeply to the west are found [Hoffman et al., 1986]. Of interest is the dramatic change in form between allochthonous rocks above the sole thrust and autochthonous rocks below. The autochthon seems remarkably unaffected by D1 as it lacks any D1 structures [Hoffman et al., 1986].

The structures typifying D2 are characteristically north-trending symmetrical to weakly east-vergent folds involving basement, its autochthonous cover and the overlying allochthon. It produced up to 13 km of structural relief

[Hoffman et al., 1986]. This deformation was responsible for the system of basement folds including the Exmouth, Scotstoun and Carousel antiforms. The map scale F2 folds are usually broad, open and coaxial with D1 thus hinting at a possible close temporal relationship with D1 [Hoffman et al., 1986]. D1 foliations, above the basement antiforms in the Akaitcho Group, are crenulated about steep, north-trending D2 axial surfaces while elsewhere D2 foliations are roughly subparallel to the D1 foliations [Hoffman et al., 1986].

The D3 deformation was responsible for producing a set of east-northeast trending cross folds superimposed on the Calderian structures [Hoffman et al., 1986]. The 1st order D3 folds produced regional scale culminations and saddles in the basement (three structural culminations, and two saddles) providing up to 30 km of structural relief [Hoffman et al., 1986]. These structures describe a regional scale type I interference pattern [Ramsay, 1967]. Higher order or mesoscopic D3 deformation structures have not been observed in the metamorphic internal zone [Hoffman et al., 1986].

The D4 event involved east west shortening in a pure shear regime. North-south extension was accomplished by block rotations about vertical axes in a system of conjugate transcurrent faults [Hoffman et al., 1986]. No significant displacements occurred in the study area as a result of D4.

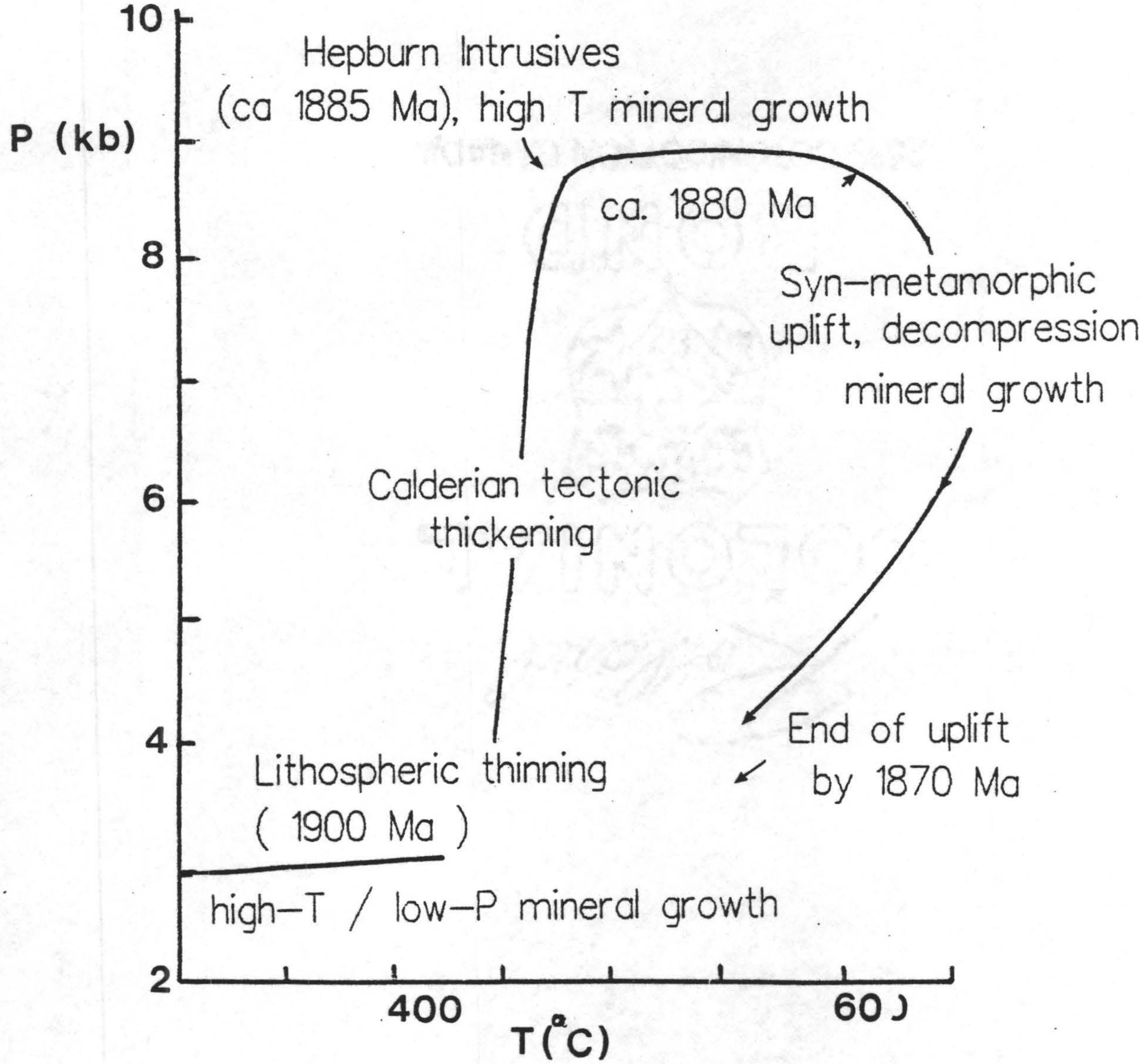
2.4 Regional Metamorphism

Several distinct mineral suites have been developed by progressive metamorphism in pelitic rocks in the Eastern Wopmay Orogen. Buchan type isograds have been defined on the ideal pelitic system K_2O-N_2O-FeO-MgO-Al_2O_3,SiO_2,H_2O. Two of these suites are related to the Hepburn Batholith and are locally inverted, depending on structural level with respect to the batholith, while the third mapped suite occurs on the east side of the Wentzel Batholith and is not inverted [Hoffman et al., 1986]. The mapped mineral isograds define a regional thermal culmination related to the emplacement of the batholiths and are elongate parallel to the structural grain of the orogen [St. Onge et al., 1986]. The batholiths have provided a heat source for the culmination after having been intruded into allochthonous units of the Coronation Supergroup. Hence metamorphic grade increases with approach to plutonic contacts, which, due to the funnel like shape of the Hepburn Batholith, takes place both structurally below and above the batholith [St. Onge, 1984b]. Thus the metamorphic inversion below the Hepburn Batholith in deep structural allochthonous rocks and underlying autochthon is consistent with the absence of syntectonic Hepburn intrusives in the autochthon.

Thrusting over a sole thrust during D1 occurred while the allochthon was approaching peak metamorphic temperature

FIG 2.1: Peizothermic Array : Eastern Wopmay Orogen

The three overlapping, thermally additive, tectonic regimes making up regional metamorphism (from Hoffman et al.,).



10
P (kb)
8
6
4
2
Hepburn Intrusives
(ca 1885 Ma), high T mineral growth
ca. 1880 Ma
Syn–metamorphic
uplift, decompression
mineral growth
Calderian tectonic
thickening
End of uplift
by 1870 Ma
Lithospheric thinning
(1900 Ma)
high–T / low–P mineral growth
400
60ɔ
T(°C)

as implied by isograds continuous on either side of the sole thrust [Hoffman et al., 1986]. Because of this, the batholiths must have been intruded as the basement closed during the Calderian Orogeny. The isograds appear to postdate the tight D2 basement synform between the Exmouth and Scotstoun antiform [Hoffman et al., 1986]. However, in the study area petrographical evidence suggests that the peak prograde mineral assemblages were achieved concurrently with D1 and pre-D2. This will be discussed in greater detail in a later section.

In summary, regional metamorphism of the eastern Wopmay Orogen consists of three overlapping, thermally additive, tectonic regimes {FIG 2.1}. These represent (i) the initial increase in heat flux resulting from lithospheric stretching, (ii) advection of heat during Hepburn plutonism and (iii) thermal relaxation concurrent with uplift following crustal thickening [Hoffman et al., 1986]. The entire orogeny took place very quickly (ca. 20 Ma), which had the effect of minimizing cooling, thus maintaining high temperatures during deformation.

2.5 Description of Allochthonous Units in Study Area

A variety of lithologies have been mapped in the study area, yet few could be easily distinguished as mappable units. All have been previously interpreted to be intercontinental rift basin deposits or as a back-arc rift basin deposits and therefore part of the Akaitcho Group.

The mandate of this thesis is documentation of the structural and metamorphic nature of the study area. As stratigraphical units cannot be traced for any great length (thereby rendering them useless for structural analysis) a detailed stratigraphical documentation will not be made nor will an attempt to establish them positively as Akaitcho Group deposits. Interpretations of previous workers showing that the allochthonous rock in the study area belong to the Akaitcho Group will be assumed to be correct for the purpose of this study.

A detailed stratigraphical column of the Akaitcho has been defined by Easton, [1981c] but predates recognition of its complex structure [King, 1986]. The Akaitcho Group should only be thought of as a tectonstratigraphic assemblage, the units of which can be interpreted to have originated in a rift environment.

A brief description of all lithologies will now be given supplemented by thin section observation, where applicable.

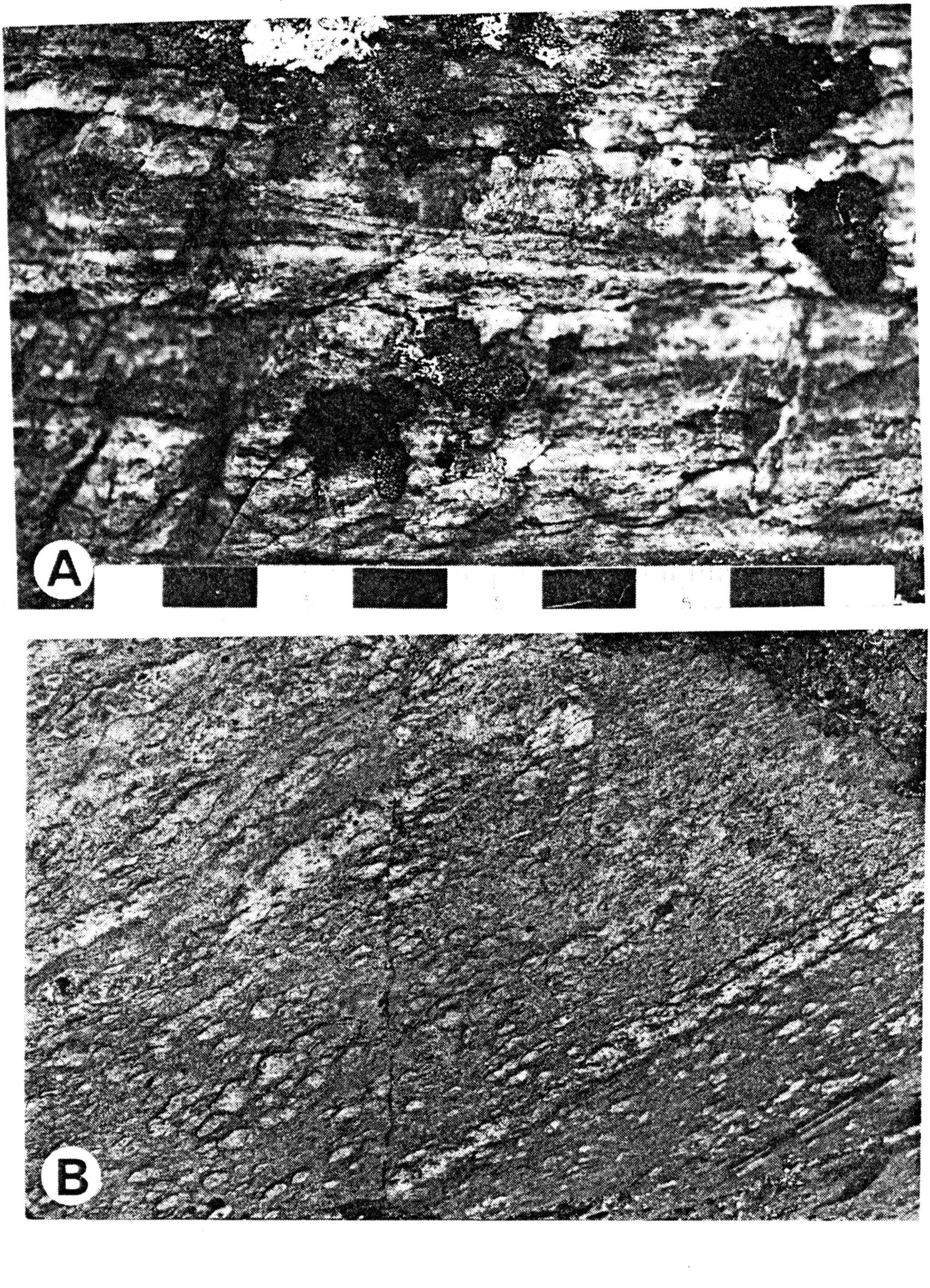
2.5.1 <u>Meta-orthoquartzite, Dirty metaquartzite</u> - Both types of quarzites are found sometimes occurring together in an interlayered fashion and sometimes as distinct units. Distinguished from each other by the abundance of quartz, the dirty quartzite is still very quartz rich when compared to

Plate 2-1A: Quartzite, Metaquartz sandstone

Crossbedded quartzite in allochthon. Scale is in centimeters.

Plate 2-1B: Pelitic, semipelite

Typical aluminous sediment found in central portion of study area. Porphyroblasts of pods of sillimanite and quartz termed faserkiesel and are a metamorphic feature not characteristic of all pelites and semipelites in the study area. Pen length is 15 cm.


A
B

psammitic rocks. The orthoquartzite is composed of quite pure recrystallized quartz with minor detrital muscovite, biotite and amphiboles. The quartz is fairly coarse grained and shows irregular subgrained boundaries which, along with their highly undulose nature, indicate extensive recrystallization. Crossbedding is sometimes preserved in the orthoquartzites.

2.5.2 Metapelites - Mica rich aluminous metasediments {Plate 2-1A} have been assigned the field nomenclature of metapelite. Sillimanite is usually abundant and garnet is often observed. Epidote is a common accessory mineral. Quartz and feldspars are always present yet in smaller proportions than in metasemipelites. In hand specimen, rocks termed metapelites are coarse grained enough to distinguish their micaceous nature. These field observations were confirmed by thin section examination. In these lithologies metamorphic growth of micas has defined a pervasive schistosity, identified as S1 {Plate 2-1B}.

2.5.3 Metasemipelites, Metasiltstones - This field term has been used to denote a fine grained detrital rock which is aluminous in nature, yet richer in quartz and feldspar than pelites. Garnet and sillimanite are sometimes present, but much less conspicuous than in pelites. Mafic minerals such as tremolite and olivine are sometimes present attesting to the rift basin origin interpretation. S1 is defined by aligned micas.

2.5.4 Meta-arenites, Biotite Sandstones - There is little compositional difference between this field subdivision and the semipelite, siltstone category. The distinguishing field difference is grain size, the biotite sandstones being coarser. Caution should be used in using this apparent grain size difference in making any metamorphic interpretations as the assessment of grain size was not systematic.

Many rocks identified originally as biotite sandstones contain faserkiesel pods. This feature indicates an aluminous rich rock (v. section 3.3) probably best classified as metasemipelitic or metapelitic.

All assemblages described above are commonly found interlayered with each other. Most typically this interlayering consists of pelitic rich sediments and quartz rich sediments {Plate 2-2B}. The layering varies from thickly layered to very irregular thin layers. The interlayered sequences cannot be put in any reasonable stratigraphic correlation with the surrounding lithologies due to their sporadic, non continuous nature.

2.5.5 Metaconglomerate - This lithology is not volumetrically substantial but occurs sporadically throughout the study area. Where seen, the conglomerates are quite coarse grained with clasts from several centimeters to tens of centimeters in length {Plate 2-2A}. The clasts are almost always elongate parallel to bedding with an approximate 4:1 length to width ratio. Clasts are dominantly composed of quartz rich and quartzo-feldspathic rocks with occasional calcareous clasts. The clast lithology confirms the depositional basin's proximity to a continental craton. The conglomerates are predominantly matrix supported, the matrix usually is pelitic to semipelitic in composition. As the clasts examined lacked a pervasive schistosity and any foliation observed in the matrix was deflected by the clasts, the formation of the conglomerate is interpreted to be pretectonic.

2.5.6 <u>Metabasalt</u> - This lithology is not common in the study area and occurs mostly in random association with the sedimentary lithologies. In hand sample it appears as a dark, fine grained rock with visible euhedral crystals. Thin sections are dominated by hornblende (subhedral to euhedral), diopside, plagioclase (undetermined composition) and minor amounts of quartz.

2.5.7 <u>Metagabbro</u> - This is the intrusive equivalent of basaltic lithologies. Actinolite is observed in the eastern end of the study area indicative of a grade lower than

Plate 2-2A: Metaconglomerate

Typical matrix supported metaconglomerate in allochthon. Clasts are coarse grained and quartz rich. Clasts are elongate parallel to bedding. Lens cap is 5 cm in diameter.

Plate 2-2B: Layered Metasandstone

Layered quartz rich (quartzite) metasandstone and psammite, metasiltstone in allochthon. The layering is regular and the quartz rich layers are generally thicker than the psammite layers. Fairly common "lithofacies" in allochthon. Scale is in centimeters.


A


B

amphibolite grade. Hornblende is the more common amphibole to the west, indicating amphibolite grade [Winkler, 1979].

The two preceding lithologies show very few signs of strain when compared to the sedimentary units, most likely a result of their homogeneity and isotropy.

2.5.8 Metavolcaniclastic - This rock is distinguished with difficulty from biotite sandstones by the proportions of mafic minerals such as hornblende and actinolite. While the biotite sandstones sometimes contain these minerals, they are not as prominent as in the volcaniclastics. This distinction is based only on estimated mineral proportions. The volcaniclastics show bedding and contain visibly abraded quartz grains unlike the metagabbros and metabasalts.

In air photographs there are prominent ridges in the very eastern part of the study area (in the muscovite-sillimanite metamorphic zone to be discussed shortly [v. section 3.1]) which appear more rigid and less deformed than the rocks to the west. As it turns out, these ridges are dominated by volcanic rocks; metagabbros, metabasalts and metavolcaniclastics. That these rocks do not exhibit a great deal of strain (i.e. higher order folds, strong schistosity, etc) is a reflection of their composition and relatively massive nature. Most of the study area, however is made up of finely layered metasedimentary units which show a great deal of deformational strain. The distribution of stratigraphical units in the study area are presented in FIG 2.2.

FIG 2.2: Simplified Stratigraphical Map of the Study Area

Lithological units as mapped in the basement, autochthon and allochthon are depicted. Note the general lack of stratigraphical continuity in the allochthonon. Continuous mappable units are enclosed in dashed lines and is of the lithology depicted inside.

Metaquartzite

Metapelite

Metasiltstone

Meta-arenite

Metaconglomerate

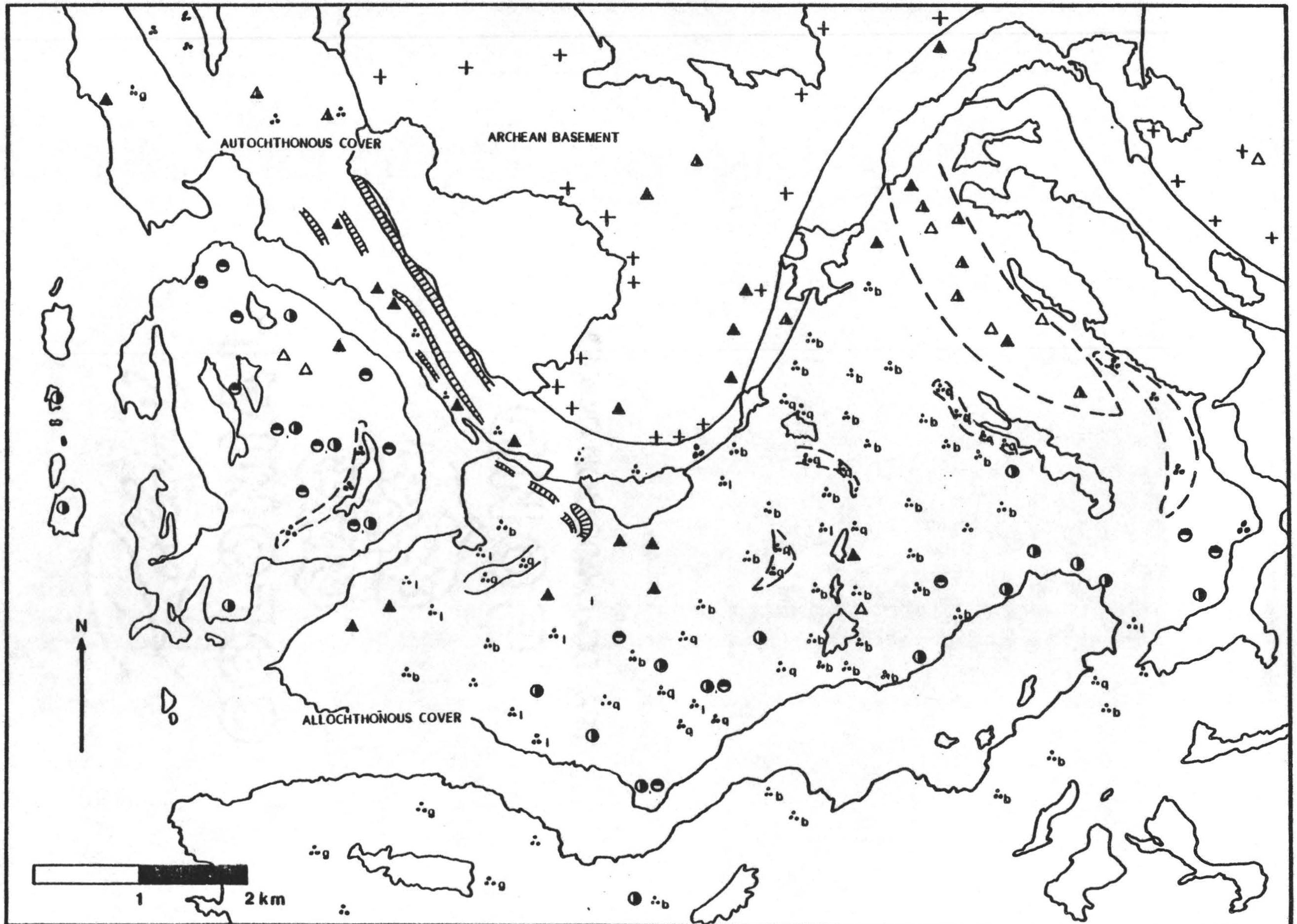
Metabasalt

Metagabbro

Metavolcaniclastic

Gabbroic sills

Granite, undifferentiated



AUTOCHTHONOUS COVER
ARCHEAN BASEMENT
ALLOCHTHONOUS COVER
N
1
2 km

CHAPTER 3 : METAMORPHISM

3.1 Introduction

The rocks of the study area located south of the Exmouth Antiform are made up entirely of Akaitcho group rift deposits, some of which are more useful than others in terms of systematic determination of regional metamorphic conditions. Most useful assemblages are the pelitic and semi-pelitic compositions. Psammitic compositions containing pelitic components in addition to quartz and feldspar but lacking in appreciable mafic constituents were also examined, as the metamorphic changes in these compositions are very like those for pelitic rocks. Outcrop exposure of pelitic, semipelitic and psammitic rocks is excellent allowing a good representation of samples from most of the study area. Determination of metamorphic grade is here based only on field and thin section evidence; more elaborate geochemical techniques are beyond the scope of this thesis.

Metamorphism of the Wopmay Orogen is characterized by a suite of syntectonic, low P (Buchan-type) mineral isograds associated with the emplacement of early Proterozoic plutons in the west part of a deformed and eastward transported continental margin prism [St Onge, 1984c]. The basement autochthonous cover beneath and allochthonous cover (containing the study area) above, the sole thrust around the Exmouth Massif (all represented in the study area) are "hot side up" or inverted [St. Onge et al., 1984]. This inversion of mineral isograds is the result of the emplacement of rocks of the Hepburn Intrusive Suite at relatively high structural/stratigraphic levels and the subsequent thrusting of the hot allochthon over a cold basement and autochthonous cover.

One isograd separating two grade zones has been identified by the author. This differs in position from that suggested by St. Onge et al. [1984] based on field identification of key index minerals. The isograd has been repositioned significantly to the east of where first positioned as a result of two lines of evidence;

(1) the thin section identification of the higher grade zone index minerals to the east in an area previously interpreted to be the lower grade zone

(2) recognition of a substantial zone of retrogression from the higher grade zone that has been inferred from the occurrence of faserkiesel.

3.2 <u>Metamorphic Grades of the Study Area</u>

Twenty one thin sections were examined, their mineral assemblages recorded {FIG 3.1} and their metamorphic textural relationships noted. On the basis of these thin sections two broad metamorphic grade zones, muscovite-sillimanite (desig-

FIG 3.1: Mineral Assemblages in Study Area

Localities of mineral assemblages from favourable pelitic and semipelitic rocks within the allochthon. Most assemblages are the result of direct thin section observation. Several assemblages are inferred from the field observation of faserkiesel. These are denoted by the abbreviation field obs.. Thin sections examined which contained faserkiesel (of varying development) are denoted with an F. Mineral abbreviations are summarized in Table 3.1.

Archean Basement

Autochthonous Cover

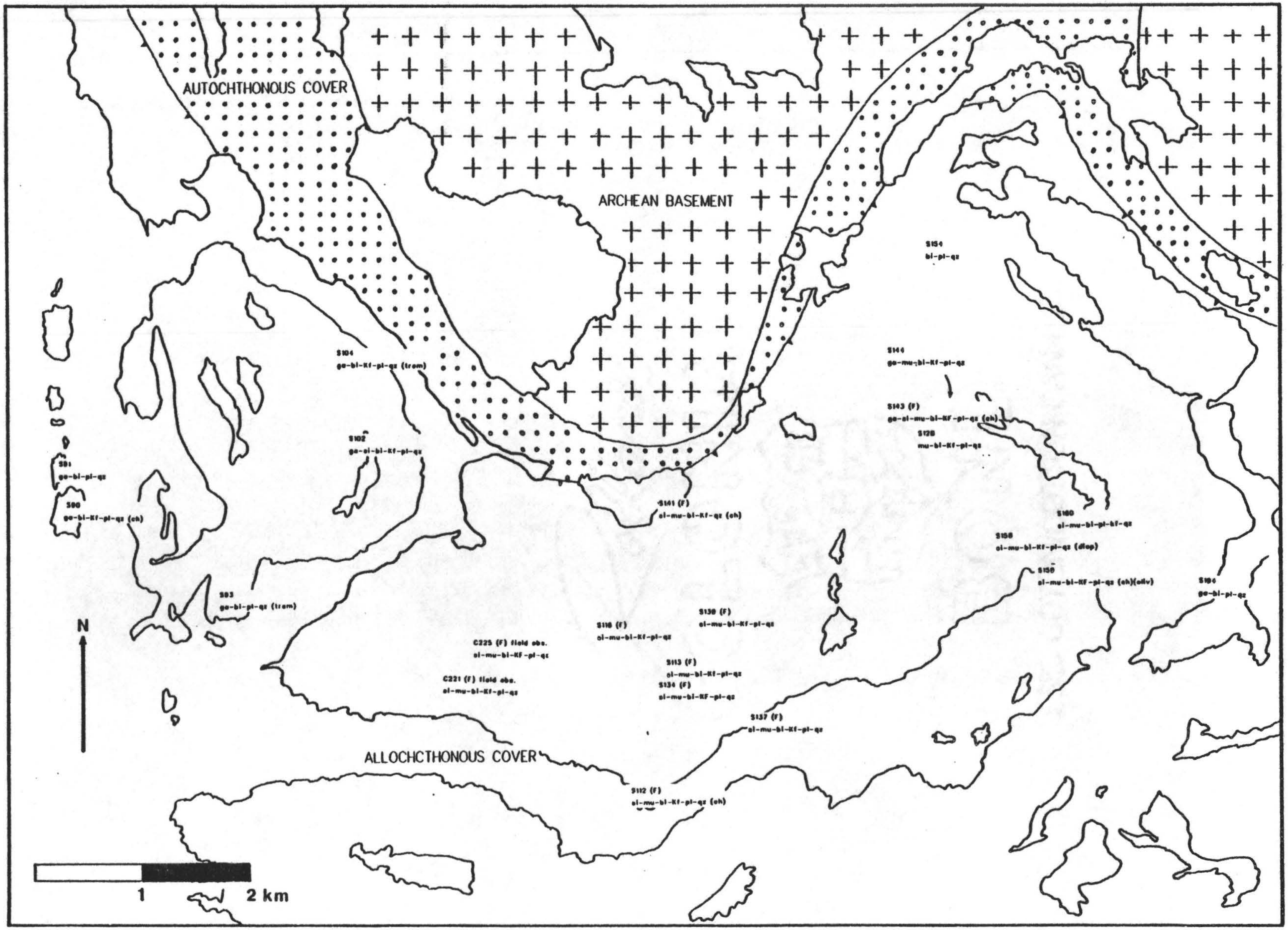

AUTOCHTHONOUS COVER
ARCHEAN BASEMENT
ALLOCHCTHONOUS COVER
S154
bi-pl-qz
S144
ga-mu-bi-Kf-pl-qz
S143 (F)
ga-al-mu-bi-KF-pl-qz (ch)
S128
mu-bi-Kf-pl-qz
S160
al-mu-bi-pl-kf-qz
S158
al-mu-bi-Kf-pl-qz (diop)
S150
al-mu-bi-KF-pl-qz (ch)(oliv)
S194
ga-bi-pl-qz
S104
ga-bi-Kf-pl-qz (trem)
S102
ga-al-bi-Kf-pl-qz
S81
ga-bi-pl-qz
S80
ga-bi-Kf-pl-qz (ch)
S93
ga-bi-pl-qz (trem)
S141 (F)
al-mu-bi-Kf-qz (ch)
S139 (F)
al-mu-bi-Kf-pl-qz
S118 (F)
al-mu-bi-Kf-pl-qz
S113 (F)
al-mu-bi-Kf-pl-qz
S134 (F)
al-mu-bi-KF-pl-qz
S137 (F)
al-mu-bi-Kf-pl-qz
S112 (F)
al-mu-bi-Kf-pl-qz (ch)
C225 (F) field obs.
al-mu-bi-KF-pl-qz
C221 (F) field obs.
al-mu-bi-Kf-pl-qz
N
1
2 km

FIG 3.2: Pelite - Semipelite Metamorphic Map

Approximate trace of the sillimanite-K feldspar isograd separating the muscovite-sillimanite (Zone A) grade zone, to the east and the sillimanite-K feldspar (Zone B) grade zone, to the west. Metamorphic grade increases to the west. Zone B is subdivided into three subzones based on differing metamorphic characteristics discussed in the text. Note; previous workers placed the sillimanite-K feldspar isograd slightly to the west of the large island in the western end of the map area.

Archean Basement ⊞

Autochthonous Cover ⁘

Sillimanite-K feldspar isograd
(semicircle in direction of higher grade)

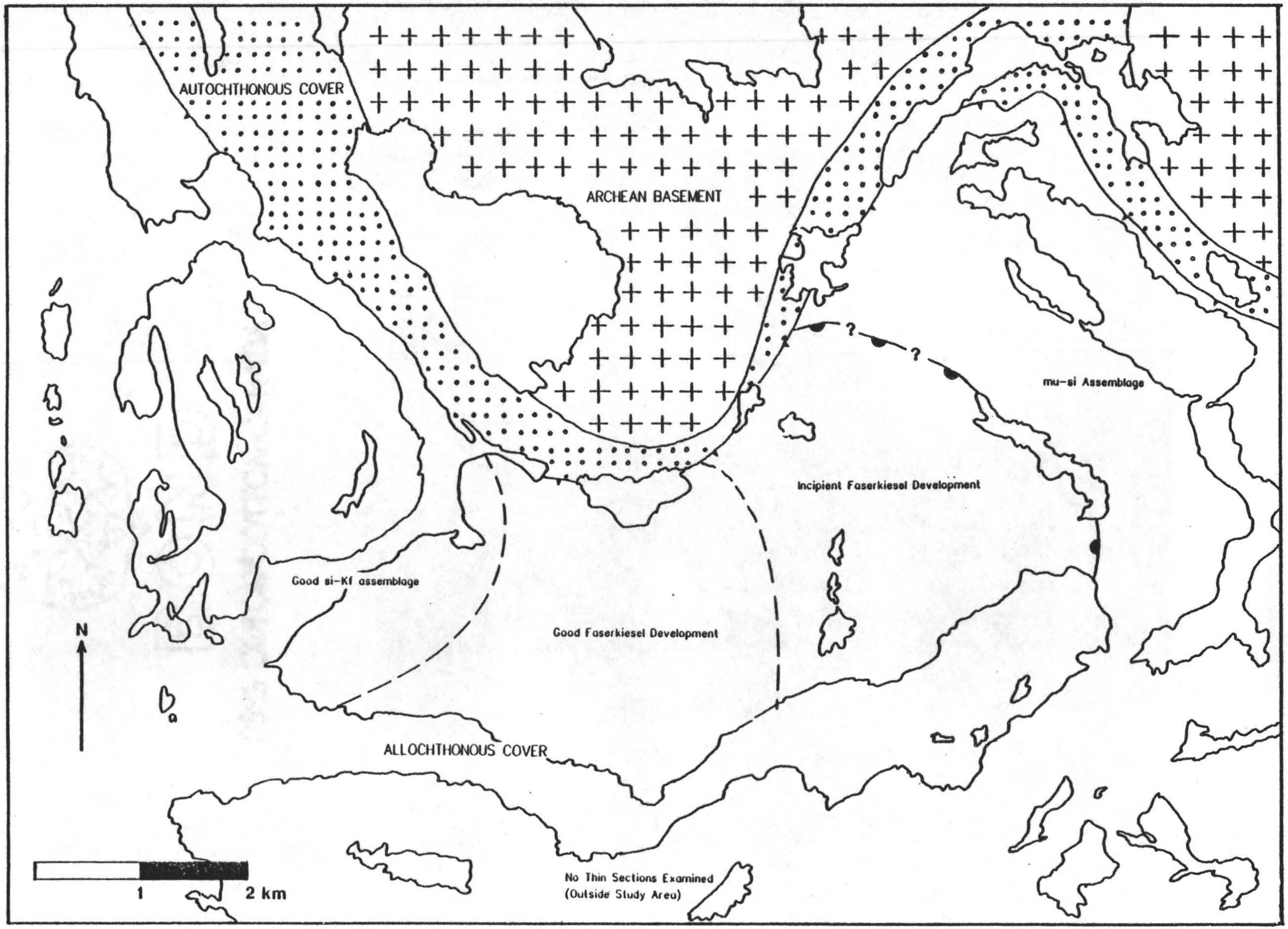

AUTOCHTHONOUS COVER
ARCHEAN BASEMENT
mu-si Assemblage
Incipient Faserkiesel Development
Good si-Kf assemblage
Good Faserkiesel Development
N
ALLOCHTHONOUS COVER
No Thin Sections Examined
(Outside Study Area)
1
2 km

TABLE 3-1: ABBREVIATIONS

ad	andalusite	si	sillimanite
als	Al_2SiS_{O5}	st	staurolite
ap	apatite	to	tourmaline
bi	biotite	V	vapour
ch	chlorite	zi	zircon
ga	garnet	P	total pressure
Kf	Kfeldspar	T	temperature
L	liquid		
mu	muscovite		
pl	plagioclase		
qz	quartz		

TABLE 3-2: Maximum phase assemlage for mineral zones in study area. Abbreviations are given in Table 1. Common accesory minerals include zi and to. (modified from St.Onge, 1984)

Mineral zone	ga	ad	si	st	mu	bi	ch	Kf	pl	qz
sillimanite (A)	-+		x		x	x			x	x
K-feldspar (B)	-+		x			x		x	x	x

nated Zone A) and sillimanite-K feldspar (designated Zone B) have been identified in the study area {Table 3.2}. Several subdivisions of Zone B have also been made based on differing metamorphic characteristics. The isograd separating Zone A and Zone B involves the breakdown, in the presence of quartz, of muscovite to form K feldspar and sillimanite in pelitic schists. It is denoted the sillimanite-K feldspar isograd. These zones are illustrated in FIG 3.2.

Zone A is defined by the assemblage <u>sillimanite + muscovite + biotite + plagioclase + quartz + graphite</u>. Although not observed in the study area, the transition to this zone from the lower temperature andalusite-muscovite zone can be explained by the following transition;

andalusite = sillimanite [St. Onge, 1984c]

Fibrolite is the common variety of sillimanite at and above this transition [Turner, 1980]. Coarser needles and laths of sillimanite are found more commonly in the upper muscovite-sillimanite and sillimanite-K feldspar zones (Zone B). Where present Zone A is well above the andulasite - sillimanite transition as no relict andalusite crystals have been observed in thin section.

Zone B is defined by the assemblage <u>sillimanite + biotite + K feldspar + plagioclase + quartz + graphite</u> and is at a higher grade than Zone A. The prograde transition from the muscovite-sillimanite zone to sillimanite-K feldspar zone can be described by the following reaction; muscovite +

plagioclase + quartz = K feldspar + sillimanite + biotite + vapour [St. Onge, 1984c].

The reaction assumes the absence of an in situ anetectic granitic melt phase. Zone B is characteristically marked by the first appearance of K feldspar poikiloblasts in contact with sillimanite grains [St. Onge, 1984c]. This has been observed in the study area also. Muscovite is observed over a wide geographical range (approximately 4 km) within this zone, partially as a result of the slow rate of the muscovite breakdown reaction and partially due to subsequent (post D1) retrograde metamorphism which in places is chlorite grade.

3.3 Faserkiesel : A Metamorphic Indicator

Field mapping in part of the sillimanite-K feldspar zone revealed the presence of abundant small pods of soft polymineralic material found mostly in pelites. These pods are composed of sillimanite and quartz with some porphyroblastic muscovite. Though not appreciated fully in the field, these pods have proved to be an invaluable tool for determining the location of the sillimanite-K feldspar isograd.

The pods, known as faserkiesel, occur exclusively in metasedimentary rocks of pelitic-psammitic composition. They are abundant enough locally to give a psuedoconglomeratic appearance to the rock. Concentrations, for obvious compositional reasons, are higher in pelitic than in psammitic

FIG 3.3: Flynn Diagram of Metamorphic Faserkiesel Pods.

A selection of the pods major axes were measured and plotted as dots. The value of the abscissa is;

$$R_{yz} = \frac{1 + e_2}{1 + e_3}$$

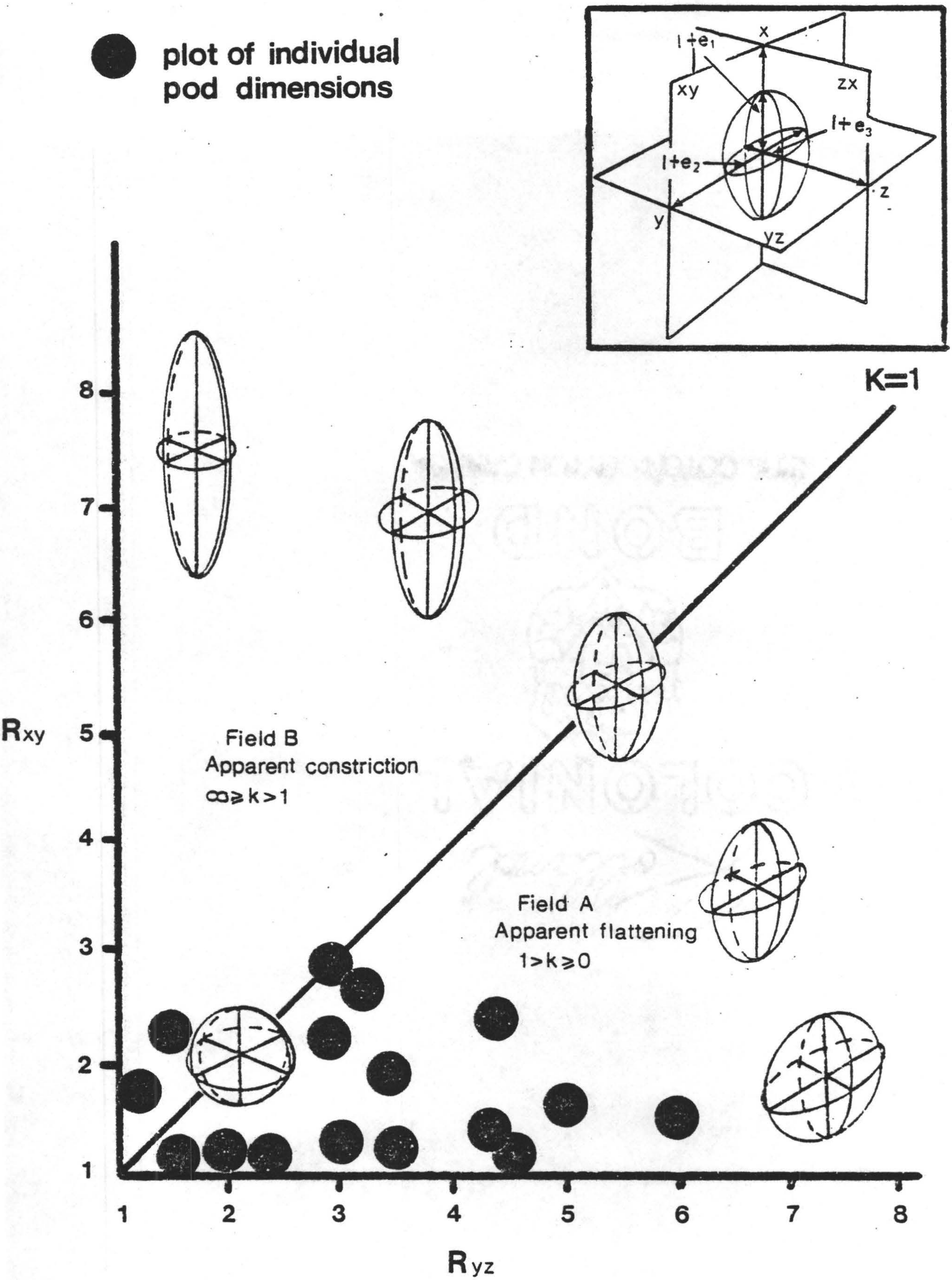
The value of the ordinate is;

$$R_{xy} = \frac{1 + e_1}{1 + e_2}$$

(where e is extension)
Real extensions are not known, and values should more accurately be considered only as relative lengths, $(1 + e_1 = X/2)$ etc.. The K value is defined as;

$$K = \frac{R_{xy} - 1}{R_{yz} - 1}$$

Most of the pods have values of K between 0 and unity and therefore, by convention are "pancake" in form (oblate ellipsoids) and lie in the field of apparent flattening [Ramsay, 1983]. Typical shapes for varying values are shown. Inset is a strain ellipsoid with rotation for principal strain, strain directions and planes.



plot of individual pod dimensions
x
$1+e_1$
xy
zx
$1+e_3$
$1+e_2$
z
y
yz
K=1
8
7
6
5
4
3
2
1
R_{xy}
Field B
Apparent constriction
$\infty \geq k > 1$
Field A
Apparent flattening
$1 > k \geq 0$
1
2
3
4
5
6
7
8
R_{yz}

Plate 3-1A: Faserkiesel pods (sillimanite-quartz porphyroblasts) and veins of folded sillimanite-quartz in semipelite. Folds represent D2. Assuming pods and veins are of the same origin, then pods are pre D2 in allochthonous rocks of the study area. Pen is 15 cm in length.

Plate 3-1B: Abundant faserkiesel pods in allochthonous pelite-semipelite. Pods are aligned with bedding and S1 on limb of a D1 fold in study area. Pen is 15 cm in length.

Plate 3-1C: Faserkiesel pods aligned with S1, both gently folding around a D2 hinge. From allochthonous rocks in study area. Pen is 15 cm in length.

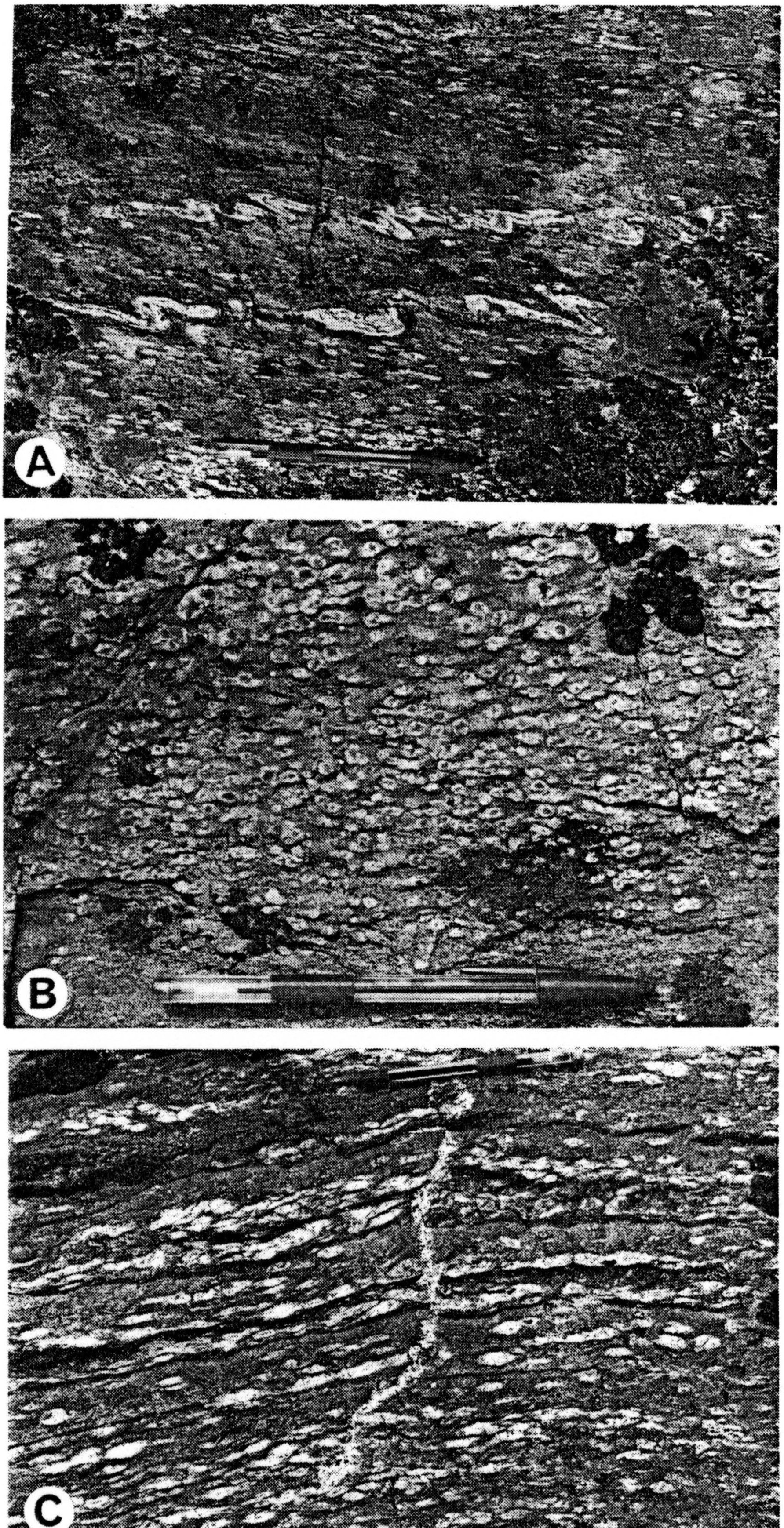
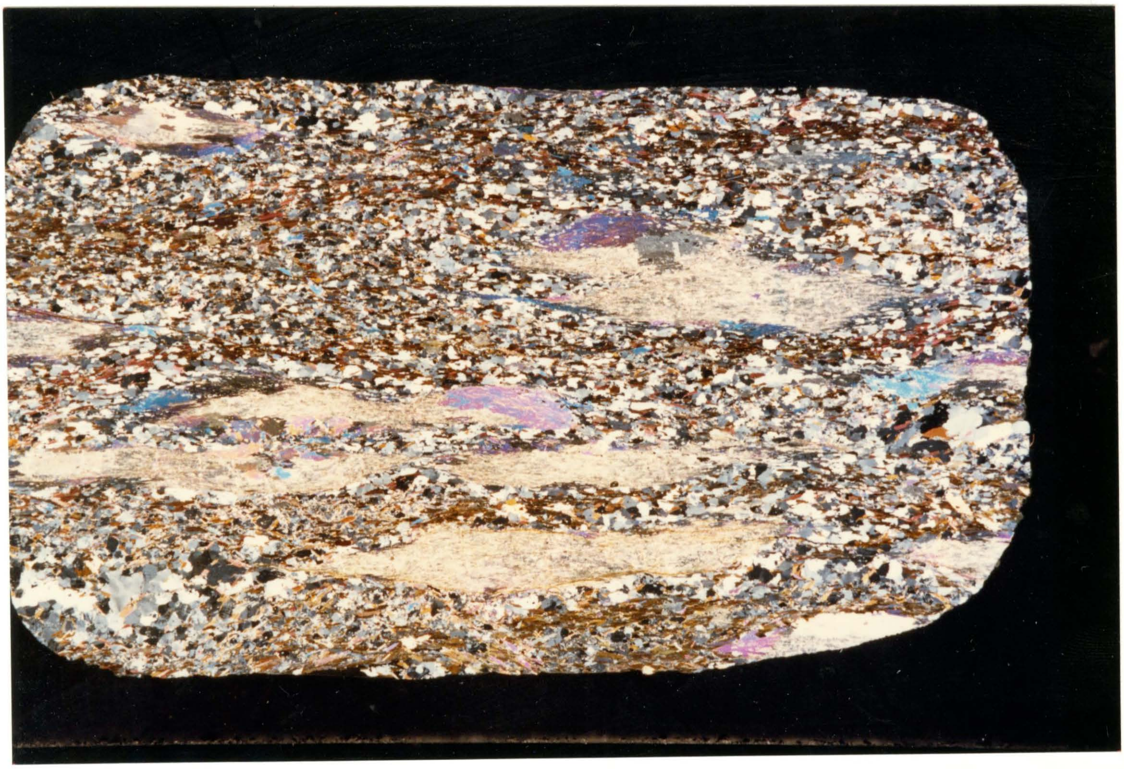
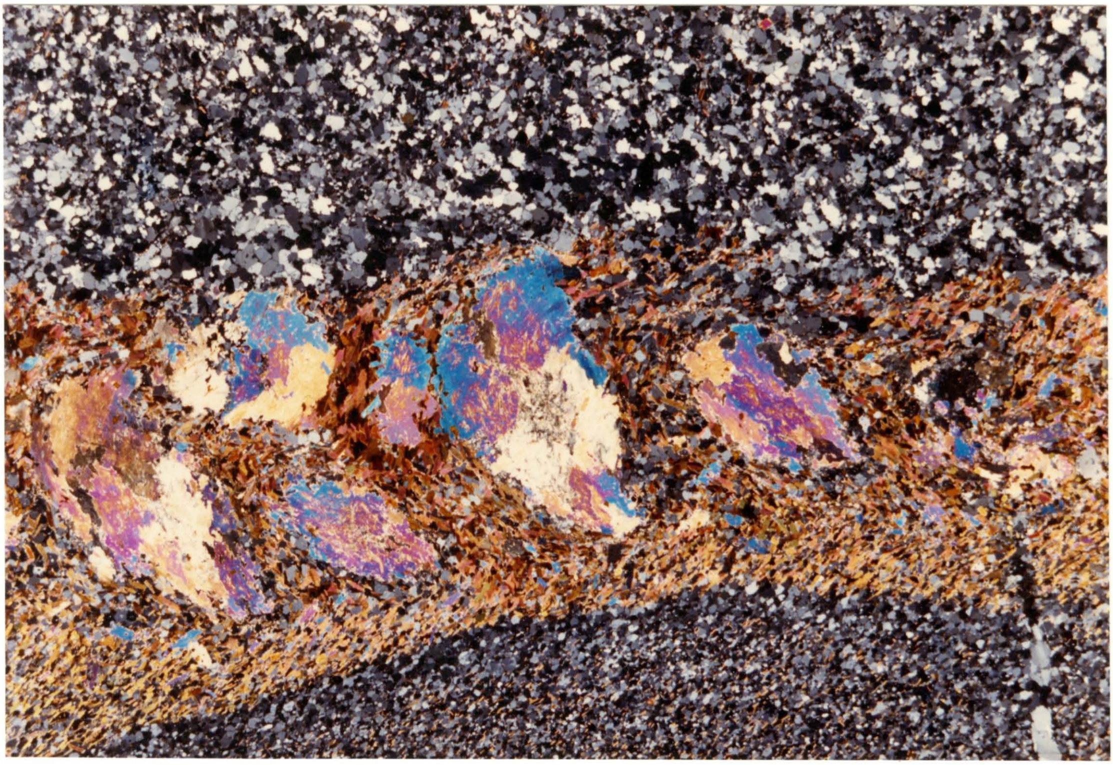

A
B
C

Plate 3-2A: Photomicrograph of thin section containing abundant faserkiesel in pelite lithology. Note segregation of sillimanite inside the pods and micas on the outside of the pods and matrix. Strong alignment with S1 exists. Tensile fracture cutting one pod is thought to a be a result of D2 stress field. Field of view is 21 mm x 36 mm, crossed polars.

Plate 3-2B: Layered metasediment with pelitic band (center). In the pelite band coarse grained muscovite is replacing coarse grained sillimanite. Other muscovite grains in band often have sillimanite inclusions (B). Thought to represent faserkiesel pods replaced by muscovite (A). Note lack of sillimanite in matrix. No similar faserkiesel textures found in adjoining quartz rich layers. Field of view is 9.5 mm x 14.3 mm, crossed polars.





rocks. Field and thin section evidence implies a correlation with D1 {Plate 3-1B}.

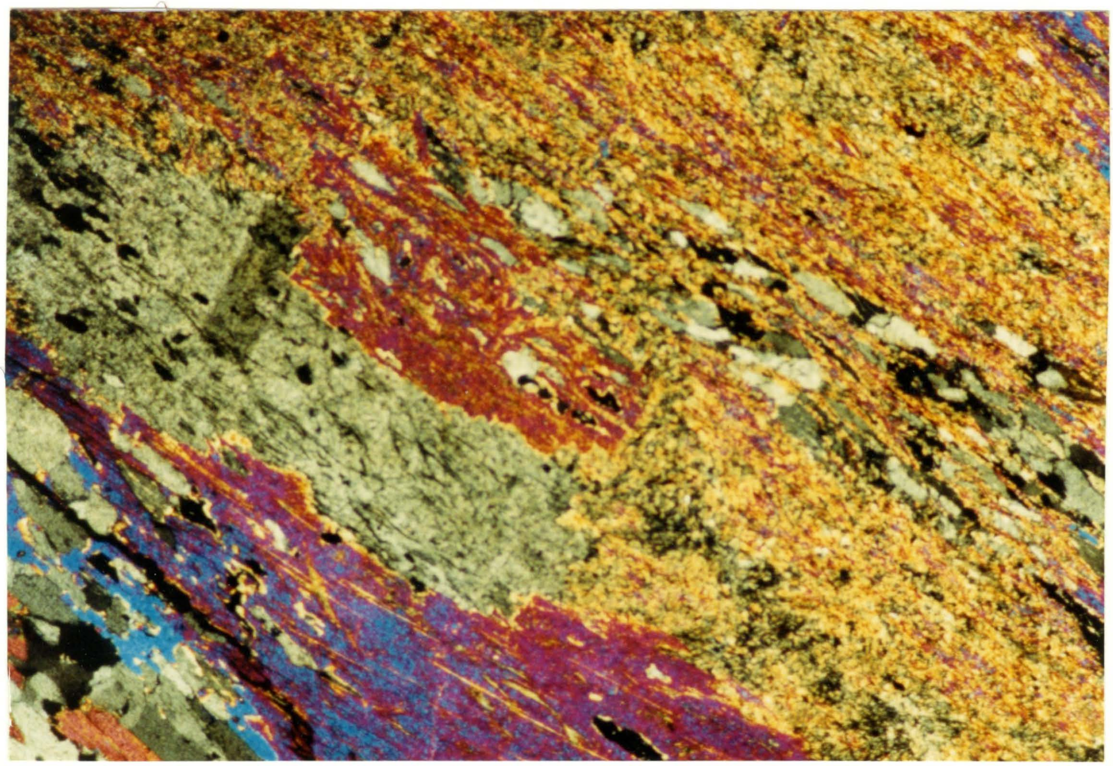
3.3.1 NATURE OF THE PODS

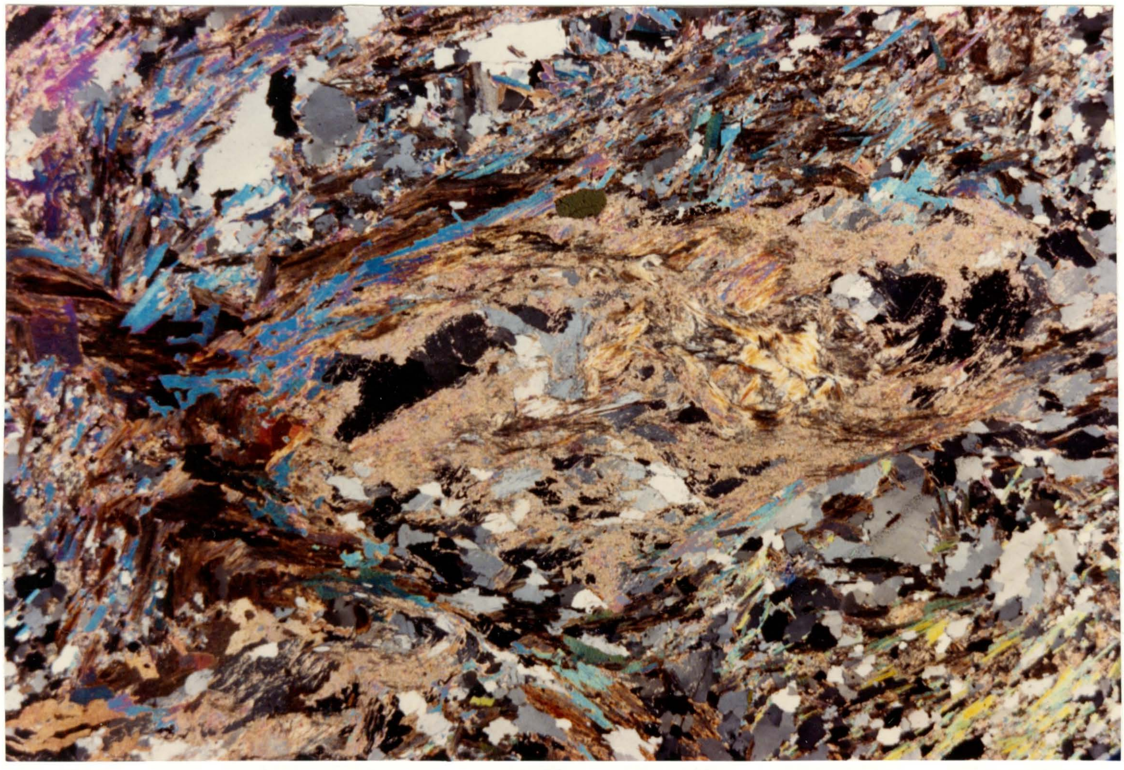
In hand specimen, the pods are ellipsoidal, bluish white, and pancake shaped. The rough average dimension of the pods is 0.9cm x 0.7cm x 0.2cm although considerable variation in size exists. FIG 3.3 shows little obvious differences in length between the long, [X] and medium, [Y] axes, respectively. The pods always align their short axis perpendicular to the prominent foliation, identified as S1. They are best described as oblate ellipsoids. Their shape seems non random and distinctly related to deformation.

In thin section, the pods are composed mainly of sillimanite and quartz with some porphyroblastic muscovite in certain areas. They always are found in aluminous sediments {Plate 3-2A,B}. Sillimanite is usually very fine grained and occurs as needles or as curving sheaves of fibrolite. It can be present interstitially to fine grained quartz or in irregular mats with little visible quartz. Many of the pods examined in thin section seemed to be surrounded and partially embayed by a zone of porphyroblastic biotite and muscovite showing textures suggesting replacement of the sillimanite {Plate 3-3A}. The orientation of the micas is almost always controlled by the shape of the faserkiesel. This feature and the replacement textures of sillimanite by

Plate 3-3A: Faserkiesel, centre of pod is top right margin of pod and encompassing matrix is bottom left,. Muscovite appears coarse, porphyroblastic and replacing sillimanite. Fine grained quartz is present interstitial to sillimanite. The coarse grain of K feldspar on outside pod (centre right) is probably prograde in origin. 1.16 mm x 1.72 mm field of view, crossed polars.

Plate 3-3B: Individual faserkiesel pod in pelite lithology. Note untwinned K feldspar on the inside of the pods being replaced by sillimanite. Partial engulfment/replacement of pod by muscovite and biotite. Field of view is 9 mm x 12 mm, crossed polars.





muscovite and biotite indicate growth of the porphyroblastic micas subsequent to the faserkiesel formation. The faserkiesel pods usually contain no feldspars (plagioclase or orthoclase), while the surrounding matrix commonly contains these minerals in various proportions. On occasion, K feldspar is found inside the pods, but is usually engulfed by sillimanite, which is thought to be replacing it {Plate 3-3B}. As well, quartz contained within the pods is of a markedly different grain size and form than quartz in the matrix, being finer grained and more equant. This is due to the fact that the quartz within the pods likely recrystallized at the time of faserkiesel formation, while the quartz in the matrix crystallized prior to that.

3.3.2 <u>TIMING OF THE PODS;</u>

The faserkiesel pods are controlled by D1 structures. The XY plane of the pods lies along S1 which is clearly distinguished from bedding in outcrops where S1 is oblique to bedding, as in the hinge area of D1 isoclinal folds. The pods are axial planar to these D1 folds indicating an origin related to that deformation structure. The pods are older than the D2 episode. They are seen in outcrop to be folded in areas of D2 folds, and the fold axes of such folded pods are consistently oriented with the range of other D2 folds axes.

3.3.3 The Origin of Faserkiesel

The origin of faserkiesel is perplexing. It is not very common in nature, suggesting an unusual set of condition. Yet in several metamorphic belts pods of sillimanite and quartz, "faserkiesel", encompassed in a matrix of K feldspar and quartz have been documented (see summary, in Tippett, 1980). The pods have been found previously in both metaplutonic environments [Breaks and Shaw, 1973] and metasedimentary environments [Tippett, 1980].

A critical feature of rocks containing faserkiesel pods is the segregation, either complete or partial, of sillimanite and quartz into the nodules, and muscovite and potassium feldspar in the matrix respectively. Chemically, the nodules are higher in alumina than the matrix while the matrix has a higher potassium and sodium content.

Previous models by Losert [1968], Eugster [1970] and others suggested that faserkiesel pods are related to dealkalization of the rocks around certain nuclei, brought about by a fluid passing parallel to schistosity, removing K and Na, and producing a residuum with high Si and Al concentrations. However, this model is not generally acceptable because it necessitates the existence of a fluid phase to act as an external sink for potassium. A correlation between faserkiesel pods and the presence of a melt phase does not always exist [Tippet, 1980].

A second feature found in metasedimentary pelites containing faserkiesel, is the presence of both products and reactants of the muscovite breakdown reaction [Tippet, 1980]. This reaction can be expressed most simplistically as;

$$KAl_2(AlSi_3O_10)(OH)_2 + SiO_2 = KAlSi_3O_8 + Al_2SiO_5 + H_2O$$

muscovite quartz Kspar sillimanite

Any explanation for the origin of faserkiesel should offer reasons for the coexistence of the assemblages listed beyond the isograd. Various writers have proposed models involving this theme [Guidotti, 1963; Evans and Guidotti, 1966], but none are totally satisfactory in explaining the extent of faserkiesel developed in high grade, metasedimentary rocks.

Tippet [1980] has proposed a model in which the development of faserkiesel may be a result of retrograde metamorphism given the right metamorphic and compositional conditions. He suggests that the persistence of products and reactants is a result of incomplete retrogression of prograde sillimanite-K feldspar assemblages. This approach has appeal in the current thesis area for the following reasons;

(1) regionally, the area has a complex structural history, involving 4 major deformational events, 3 of which effect the study area, directly. The deformations have been of varying intensity and scale; this, along with the intrusion of the Hepburn Batholith Suite coeval with D1

nearby would probably produce both spatial and temporal changes in metamorphic conditions.

(2) the correlation of D1 with the most pervasive schistosity suggests a peak in metamorphic grade at that time and the local spatial control of faserkiesel in the field by D1 structures suggests syn or post D1, but pre D2 generation.

(3) there is textural evidence in thin section for retrogression, especially replacement of sillimanite in the pods by muscovite and the replacement of some K feldspar in the pods by sillimanite.

Using the interpretation that metamorphic reactions can be considered in terms of combination of ionic equilibria summing to a total net nonionic combination (advanced by Carmichael, [1969]), the overall retrograde reaction of interest has been divided by Tippet into two ionic equilibria;

Total Retrograde Rx:

$2(K,Na)AlSi_3O_8 + 2Al_2SiO_5 + 2H_2O = (K,Na)_2Al_6Si_6O_{20}(OH)_4 + 2SiO_2$

Ionic Equilibria 1:

$2(K,Na)^+ + 3Al_2SiO_5 + 3SiO_2 + 3H_2O = 2H^+ + (K,NA)_2Al_6Si_6O_{20}(OH)_4$

Ionic Equilibria 2:

$2(K,Na)AlSi_3O_8 + 2H^+ = 2(K,Na)^+ + H_2O + Al_2SiO_5 + 5SiO_2$

In simple terms these reactions show that going downgrade across the K feldspar-sillimanite isograd, sillimanite reacts with sodium and potassium ions, silica, and water to form muscovite, while K feldspar reacts with hydrogen ions to form sillimanite. The products of the second reaction would then enter into equilibrium reaction 1, if retrogression is carried out to completion. These reactions would proceed at the edges of randomly distributed K feldspar, sillimanite and quartz grains, and as a result knots of sillimanite and quartz would develop from dealkalized K feldspar crystals while from the primary (prograde) sillimanite crystals and quartz, muscovite would be produced using the released potassium and sodium ions of reaction 2 [Tippet, 1980].

When this suggestion is applied to the present thesis area the following interpretations can be made. The pervasive schistosity, S1 was once defined by a mineral assemblage more characteristic of metamorphic conditions on and slightly above the sillimanite-K feldspar isograd (muscovite breakdown). Due to considerations to be discussed shortly there is evidence that some of the areas with faserkiesel pods were not completely prograded to sillimanite-K feldspar grade. This is not unusual as the breakdown of muscovite to sillimanite and K feldspar in many metamorphic belts is known to be sluggish. The partial retrogression to muscovite grade would then have occurred after D1, but prior to D2. The pods were formed along the preexisting D1 schistosity. This proposal meets both the strong field and the thin section evidence for the timing of the pods as well as the general documented metamorphic history of the area.

Textural relationships in thin section show most of the pods partially or totally engulfed by porphyroblastic muscovite, which, along with other evidence, strongly suggests replacement of sillimanite by muscovite. According to the proposed model, with continued retrogression, sillimanite and quartz formed from K feldspar would then proceed to muscovite, with the input of Na^+ and K^+ ions. The existence of some K feldspar alone in the matrix poses some problems, according to the model, it should have gone to sillimanite plus quartz. Tippett [1980] suggests that the initial growth of sillimanite and quartz from K feldspar promotes the clustering of subsequent sillimanite and quartz around those sites already having started the dealkalization process. Thus some K feldspar may be left unaffected. Tippett also indicates that the some faserkiesel may be the result of prograde metamorphism. This is unlikely in the study area as there are prominent overgrowths of muscovite on the sillimanite pods which effectively eliminates faserkiesel as a prograde phenomenon in the study area.

Faserkiesel pods are present over a fairly wide area (approximately 2 km) {FIG 3.2}. Accepting the retrogression hypothesis of their origin, the pods can be used as a rough estimator of the extent to which retrogression progressed in the study area. It then becomes possible to relocate the prograde K feldspar-sillimanite isograd substantially to the east of where it has been placed previously by St. Onge et al., [1984].

3.4 Metamorphic Interpretations

The reposition of the breakdown of muscovite to sillimanite and K feldspar with increasing grade as indicated on FIG 3.2 in the eastern end of the study area has been based partially on the hypothesis that faserkiesel pods are the retrograded result of pelitic assemblages which once were at K feldspar-sillimanite grade (Zone B).

The isograd separating Zones A and B is only an approximation as very few samples taken in the inferred muscovite-sillimanite zone were useful for metamorphic work. However, several non-ideal pelitic assemblages beneath the sillimanite-K feldspar isograd support this position of the isograd. These assemblages are quartz-biotite-plagioclase (oligoclase) plus orminus garnet and have previously been shown by Evans and Guidotti [1966] as being a characteristic assemblage below the sillimanite-K feldspar isograd. Further support for the placement of the sillimanite-K feldspar isograd is as follows;

(1) Just above the isograd is a zone, approximately 2 km wide, where muscovite is abundant and aligned with the

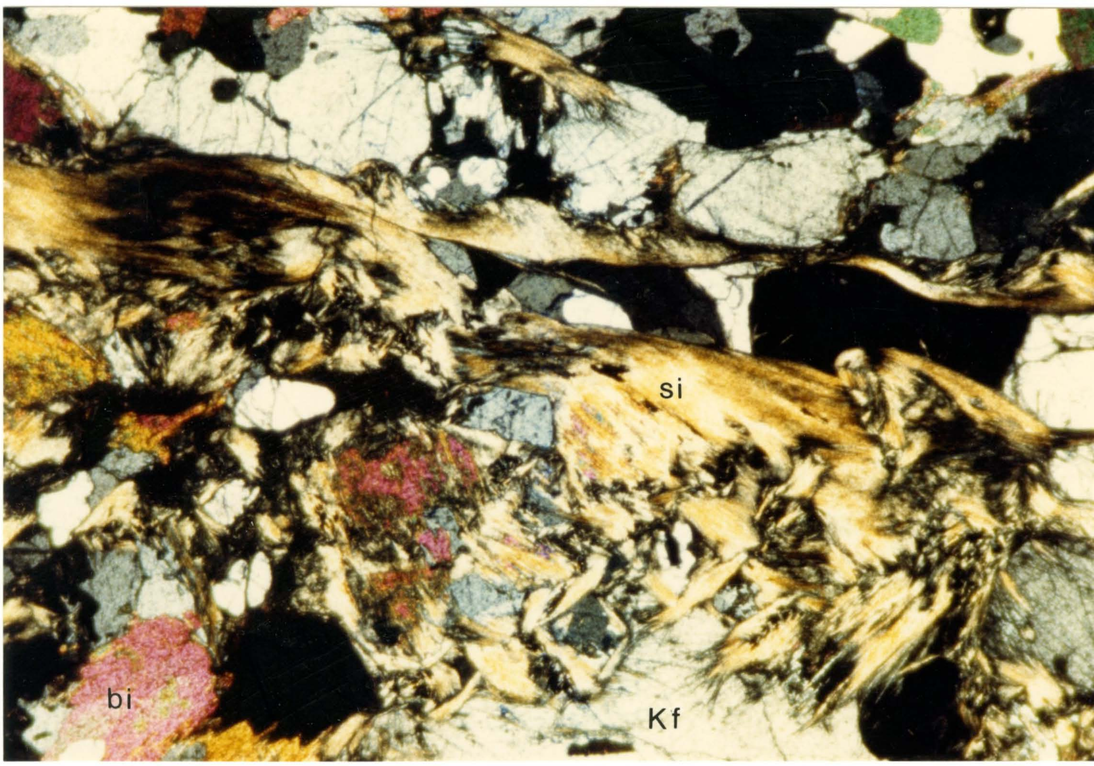
pervasive S1 schistosity. The muscovite appears fairly clean and free of replacement textures. In association are K feldspar and sillimanite.

(2) K feldspar while present in the zone just above the isograd, (marked Incipient Faserkiesel Development in FIG 3.2) is not as abundant nor as fresh as it is to the west in regions marked Good Faserkiesel Development and Good Si-Kf assemblage on FIG 3.2.

(3) Incipient faserkiesel pods were noted in thin section but not in hand samples in the region just above the proposed isograd. Since the pods centred around original K-feldspar grains, their relative scarceness in this region may merely indicate a lack of original prograde K feldspar.

As distance from the proposed isograd increased westward, maximum prograde temperature also increased. In the zone marked Good Faserkiesel Development on FIG 3.2 this results in abundant faserkiesel pods in most pelitic compositions while in a few samples, clean, abundant K- feldspar and associated sillimanite is present, with no evidence of faserkiesel. As previously indicated, faserkiesel pods are generated in rocks which have been metamorphosed above the sillimanite-K feldspar isograd. The variation in abundance of faserkiesel pods present in rocks of otherwise identical composition and undergoing the same retrograde metamorphism suggest that there is a correlation of higher prograde temperatures with the highest grade rocks containing the most

Plate 3-4: Well developed sillimanite-K feldspar assemblages from large island in western study area, Zone B. Untwinned K feldspar porphyroblasts are abundant and sillimanite grains are coarse grained and heavily matted. Some fibrolite sillimanite is present as well. Field of view is 2.5 mm x 4.0 mm, crossed polars.


si
bi
Kf

plentiful faserkiesel pods; i.e. more original metamorphic K feldspar has resulted in more faserkiesel pods. Another indication of increased prograde metamorphism towards the west is that sillimanite grains become coarser grained and more heavily matted westward {Plate 3-4} compared to the finer grained fibrolite associated with lower grades [St. Onge, 1984]. Lastly in the westernmost sections of the map area, illustrated as the Good Si-Kf assemblage in FIG 3.2, muscovite is totally lacking in pelitic assemblages.

On the basis of the various mineral assemblages found in the study area, four very generalized regions of differing metamorphic character have been set up from east to west {FIG 3.2}. The easternmost zone is of muscovite-sillimanite grade and is Zone A, as previously defined. The three western most regions lie above the prograde sillimanite-K feldspar isograd and make up Zone B. The two middle regions contain both products and reactants of the muscovite breakdown reaction; muscovite + plagioclase + quartz = K feldspar + sillimanite + biotite + vapour. Some muscovite present is thought to be retrograde in origin; but much is early, as it aligns with S1 particularly in the zone of incipient faserkiesel development just above the isograd. The coexistence of both reactants and products indicates a univariant assemblage. This reaction has been constrained in the ranges: 620 and 660^{o}C and 2.1 and 3.4 Kb [Carmichael, 1969], on the assumption of a similar ratio of Fe:Mg for biotite and muscovite, a constant value of X^{plag}_{an} and a fixed a_{H2O} [St. Onge,1984]. If we assume an average rock density of 2.7 this corresponds to a depth at time of metamorphism of between 7 and 12 km.

3.5 Tectonic Implications

If we accept the presence of faserkiesel pods as representing retrogression in areas presumably subjected to metamorphic conditions representative of the sillimanite-K feldspar grade, then the timing of the origin of these pods post dates the maximum prograde metamorphism. As indicated previously, evidence strongly suggests that the faserkiesel pods are pre D2. Therefore it is reasoned that the maximum prograde metamorphic assemblage is the result of D1 as is the pervasive schistosity seen in the area.

In discussing tectonic implications, several points must be emphasized; (1) As indicated by the regional geology map, the closest Hepburn intrusive is located approximately 10 to 15 km SE of the study area at the present exposed structural level and may have created a "hot side up" metamorphic pattern. (2) The normal trend in metamorphism is that metamorphic grade increases with the geothermal gradient and therefore depth. In the Wopmay Orogen, thrusting over the Slave Craton was from the west and therefore stratigraphic/structural depth in the allochthon increases to the west implying that higher metamorphic grades should lie west of lower metamorphic grades. This is found to be the case in the study area.

It is uncertain if the metamorphic characteristics describe a "hot side up" (or in inversion) situation or a normal geothermal metamorphic gradient. Two situations could either separately, or in combination, be responsible for the isograd derived; (i) the intrusion of the Hepburn Batholith suite at higher structural levels in which case the isograd would be expected to swing to the south, then to the east, and finally to the north, completely encompassing the lower isograd, or, (ii) a normal geothermal gradient which would result in an increase in metamorphic grade to the west. If this were the case then the isograd should continue on in a southerly direction. Unfortunately, due to the small area of study, the isograds were not able to be traced for their entirety and no positive conclusion can be revealed. Previous mapping has shown that higher metamorphic grades structurally overlie lower metamorphic grades in a wide area including the study area and therefore the "hot side up" theory is acceptable.

CHAPTER 4 : STRUCTURE

4.1 Introduction

Structural field data are shown in detail on a 1:28,000 scale map sheet, {FIG 4.1}, as mapped by King, Davies, Barrette, and Relf during the 1986 field season. Outcrop exposure is excellent, allowing for a near continuous acquisition of structural data across the study area. The stratigraphy in the study area is not conducive to the interpretation of structure, as the Akaitcho group rift deposits contain few, correlatable, mappable units [King,1986]. A second 1:28,000 scale map sheet, {FIG 4.2} is presented to show inferred macroscopic structures.

The high grade of metamorphism (amphibolite), several phases of coaxial and non coaxial deformation, as well as the lack of continuous mappable trace beds makes analysis of the gross structural characteristics of the study area difficult. While every phase of ductile deformation should have left a penetrative imprint on the rocks enabling a complete record of deformation to be made from small scale samples, in reality many complications exist. For example earlier fabrics are often obscured by later deformation, and as found by earlier researchers, small scale structures need not reflect precisely the orientation and character of large scale structures [Hobbs, Means and Williams ,1976]. Neverthe-

FIG 4.1: Structural Data Base Map

Simplified structural elements in the study area as mapped during the 1986 field season. Basement extent and age determined by geochronology from the 1983 field season [St. Onge et al., 1984]

Archean Basement

Autochthonous Cover

Allochthon

Gabbroic Sills

Quartzite Marker Beds

Sole Thrust

Bedding

Foliation/Schistosity

Gneissosity

Minor Fold Axes

Mineral Lineation (quartz, sillimanite, feldspar, hornblende) x

Crenulation Lineation c

Intersection Lineation x

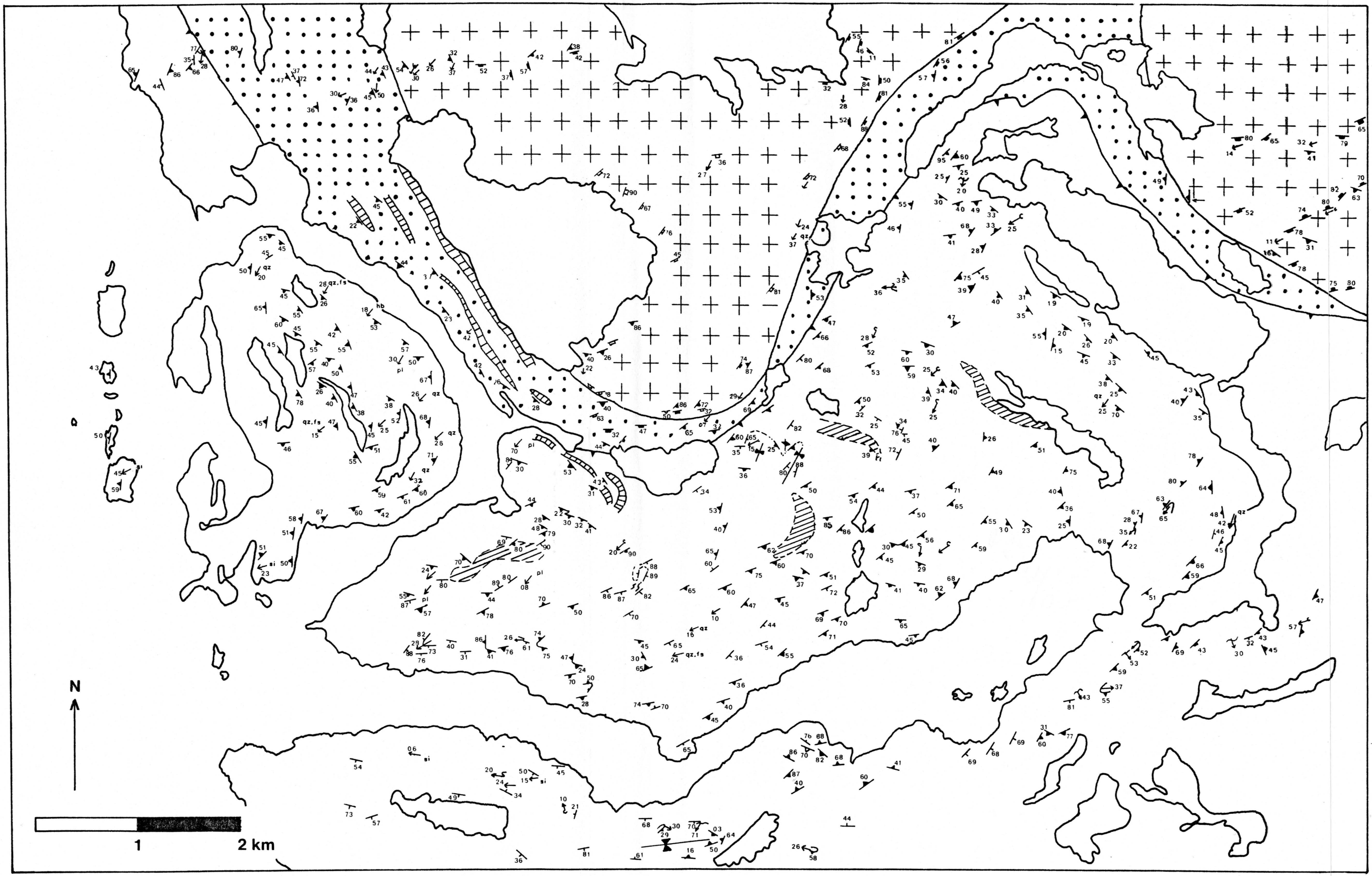

N
1
2 km

FIG 4.2: Structural Elements in the Study Area

Simplified geological map of the study area (south end of the Exmouth Antiform) emphasizing structural elements in the allochthon and to a lesser extent in the basement and autochthonous cover. A-B is line of section for FIG 4.4.

Archean Basement

Proterozoic Cover

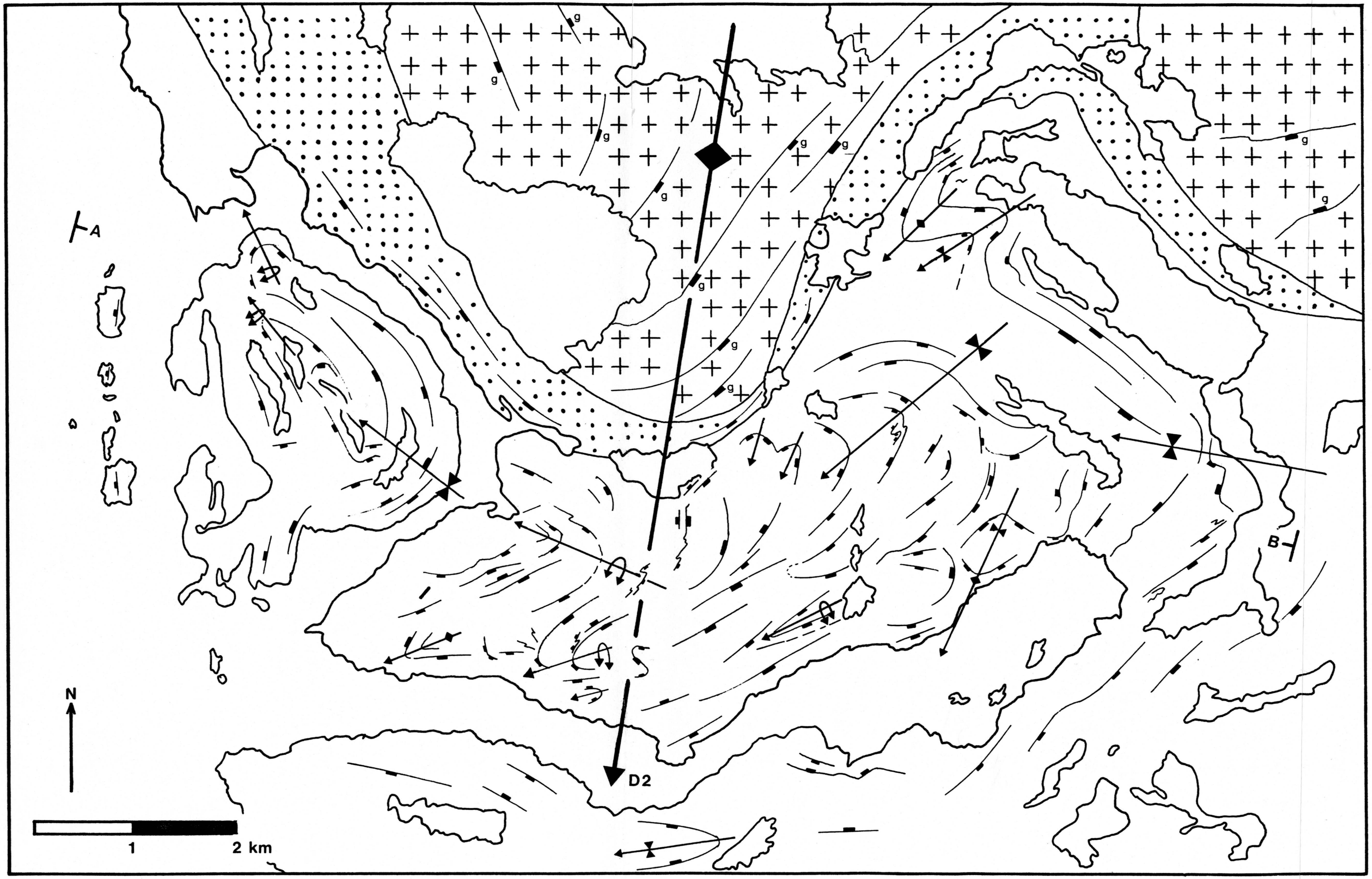
Allochthon

form lines (bedding, schistosity)

form lines in basement (gneissosity) g

Antiform

Synform


A
B
D2
N
1
2 km
g

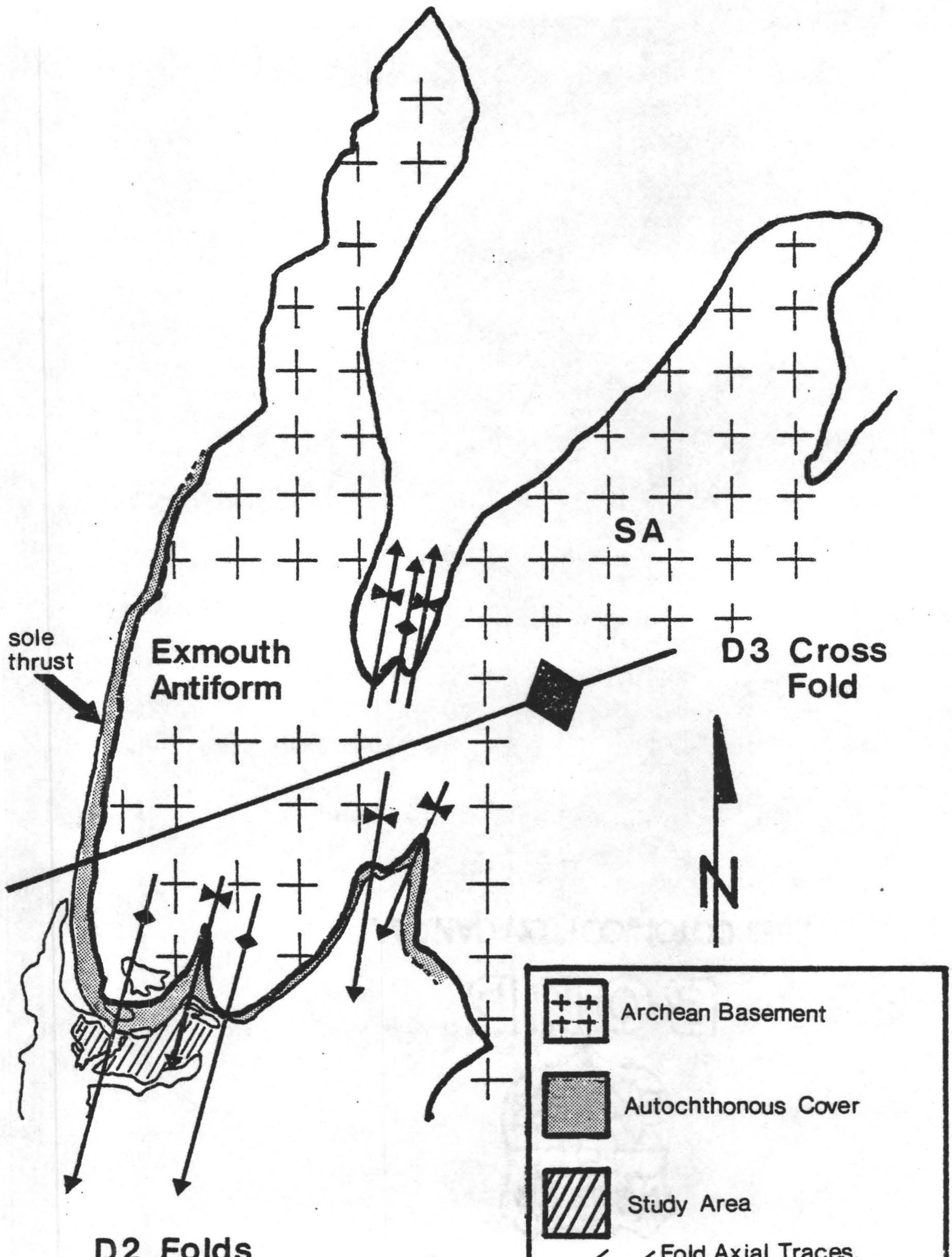
less extensive field mapping has involved examining overprinting relationships at all scales presuming that a characteristic structural style is generated by each phase of deformation. With these data, a generalized interpretation of the deformational history of the study area has been made.

4.2 General Structure of the Study Area

Mapping in the 1983 and 1986 field seasons has established that the study area is part of allochthonous Akaitcho Group rift deposits located above a basal sole thrust. Structurally below the sole thrust are basal Proterozoic units which unconformably overly Archean units. The position of the sole thrust is interpreted to be a narrow zone, now covered by 20 to 300 m of glacial drift separating two domains of drastically different structural style [King, 1986]. Nowhere is there first order evidence for the existence of the sole thrust. The basement units occur within the core of one of three large scale fold structures present in the Metamorphic Internal Zone {FIG 4.3}. Subsequent cross folding of these basement folds has produced a regional Type I interference pattern [Ramsay, 1967]. The southern half of the Calderian Exmouth Antiform is on the southern limb of the Tree River cross arch and plunges moderately south westward [King, 1986]. These D3 cross folds (Tree River Deformation) have provided 30 km of composite

FIG 4.3: Location of Basement Fold Structures

Location of two of the three large scale fold structures and regional decollement (sole thrust) as exposed by a major northeast-southwest D3 cross fold. The third fold structure, the Carousel Antiform, while not depicted, is continuous with the basement units and lay to the east of the map. SA - Scotstoun Antiform. The D2 folds in the southern half of the Exmouth Antiform have a southerly plunge while those to the north have a northerly plunge.


SA
sole
thrust
Exmouth
Antiform
D3 Cross
Fold
N
Archean Basement
Autochthonous Cover
Study Area
Fold Axial Traces,
Antiform, Synform respec.
D2 Folds

structural relief permitting the construction of useful down plunge cross sections of the basement cored antiform.

Three discrete episodes of deformation have affected the study area (D1, D2 and D3). The first episode of deformation, D1, involved the thrusting and folding of the Coronation Supergroup over the basement and cover. This produced recumbent isoclinal and east vergent folds in the allochthon but did not affect the basement. Coeval with D1 was the emplacement of the Hepburn Batholithic Suite responsible for the unique metamorphic character of the area. D2 involved shortening of the basement, autochthonous cover and allochthon, producing large scale north trending folds with no associated thrusting [Hoffman et al., 1986]. D3 folding which followed shortly after D2, trended in a east-northeast direction and produced large scale folds with no recognizable mesoscopic affects. D3 folding is thought to be the result of buckling of a relatively thin lithosphere [Hoffman et al., 1986].

The deformation history just presented is the result of several intensive seasons of field work in many different areas in the Metamorphic Internal Zone. It is within this framework that the author has interpreted the structure of the study area.

4.3 D1 Deformation

Thrusting and shortening of allochthonous Coronation Supergroup deposits above a basal sole thrust during D1 has resulted in two important characteristics within the study area.

(1) The creation of early recumbent, isoclinal folds of bedding with an associated axial planar schistosity. With continued compression this type of mesoscopic D1 fold folded both bedding and schistosity (S1) {plate 4-1A,B}.

(2) The rotation of early macroscopic thrust slices creating many over-rotated recumbent folds. Both bedding and schistosity are folded in these folds.

The over-rotated folds can be easily identified from FIG 4-2. This distinctive pattern of east vergent, over-rotated folds is present in the allochthon but is not found in autochthonous rocks. Over-rotated folds, associated with thrust slices are a common feature in many thrust fold belts, particularly in the early stages of deformation [Price et al., 1981].

Tight to isoclinal recumbent folds are common in the study area. They seem to represent a continuum in time as some are observed with axial planar cleavage (S1) while others involve folds of bedding and cleavage (S1) {plate 4-2A,B}.Occurring with small wavelengths and amplitudes (less than 0.5 m each), they often occur side by side over great

Plate 4-1A: Isoclinal D1 fold, folded about F2 in layered allochthonous metasandstones. Shows Type III interference pattern. Hammer is 40 cm in length.

Plate 4-1B: Typical D1 isoclinal fold of bedding with axial planar cleavage. In allochthonous layered metasediments. Lens cap is 5 cm in diameter.

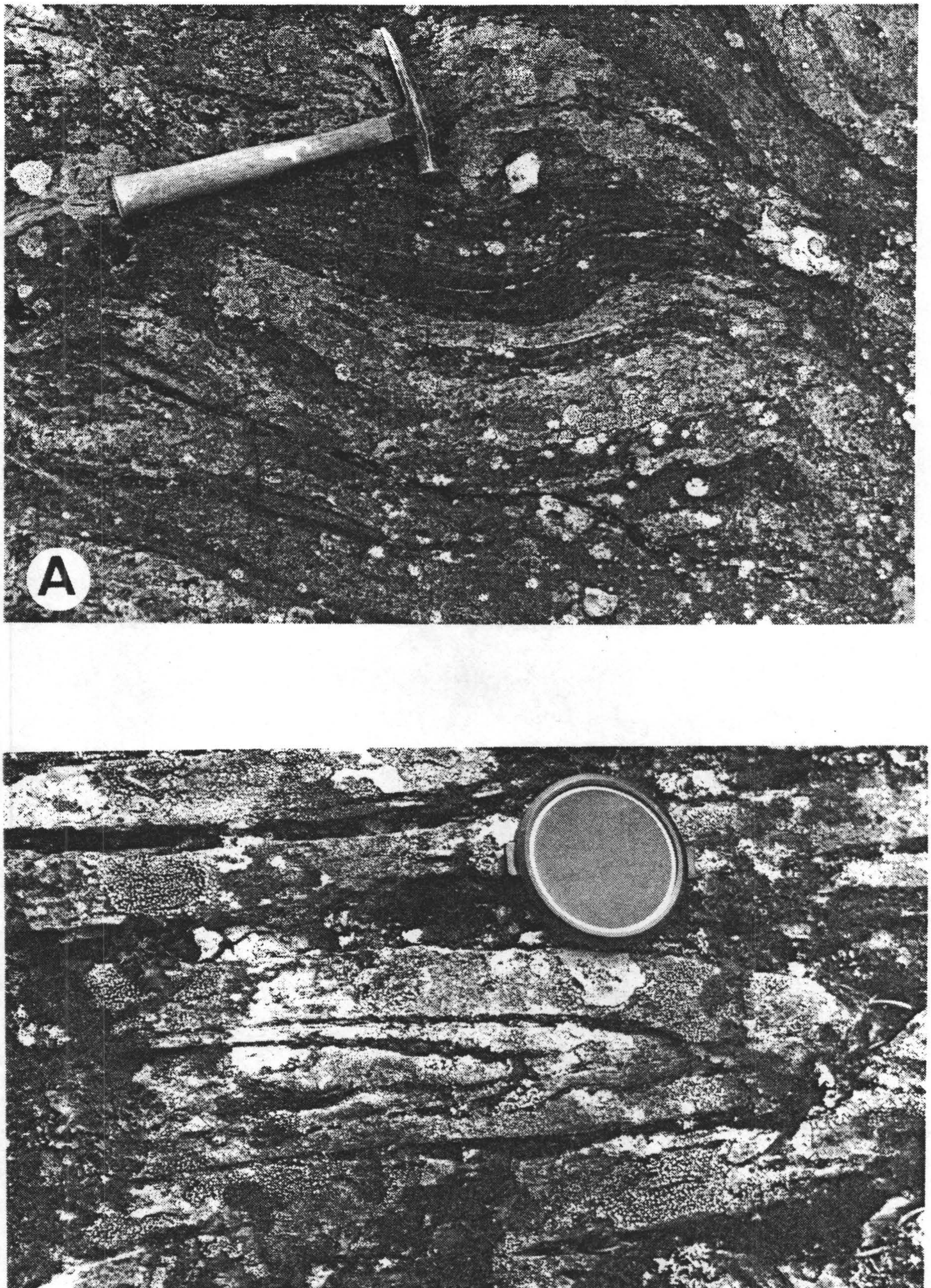
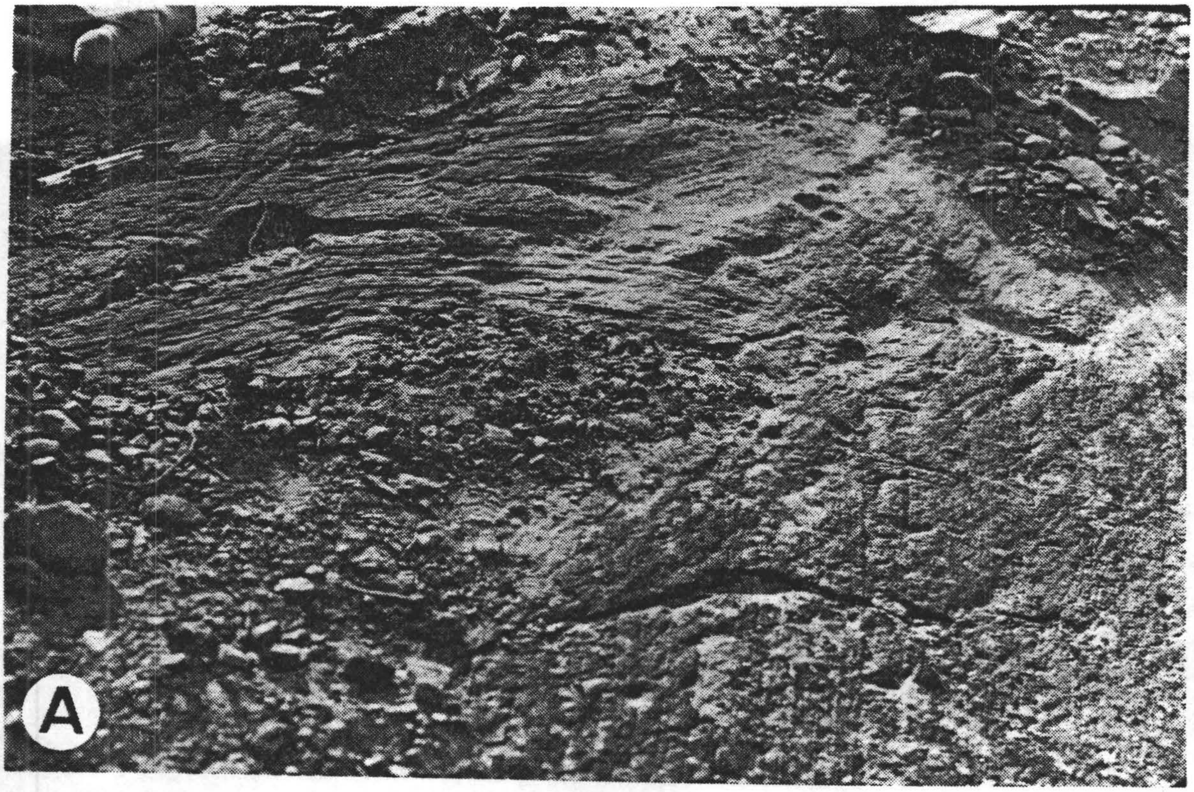

A
B

Plate 4-2A: Tight D1 fold closure fo limb of broad F2 in central allochthon. Folded bedding with barely discernable axial planar cleavage (S1). Pen is 15 cm in length.

Plate 4-2B: D1 fold closure in central allochthon showing folding of S1. Indicates a continuum of time in D1. Pen length is 15 cm.



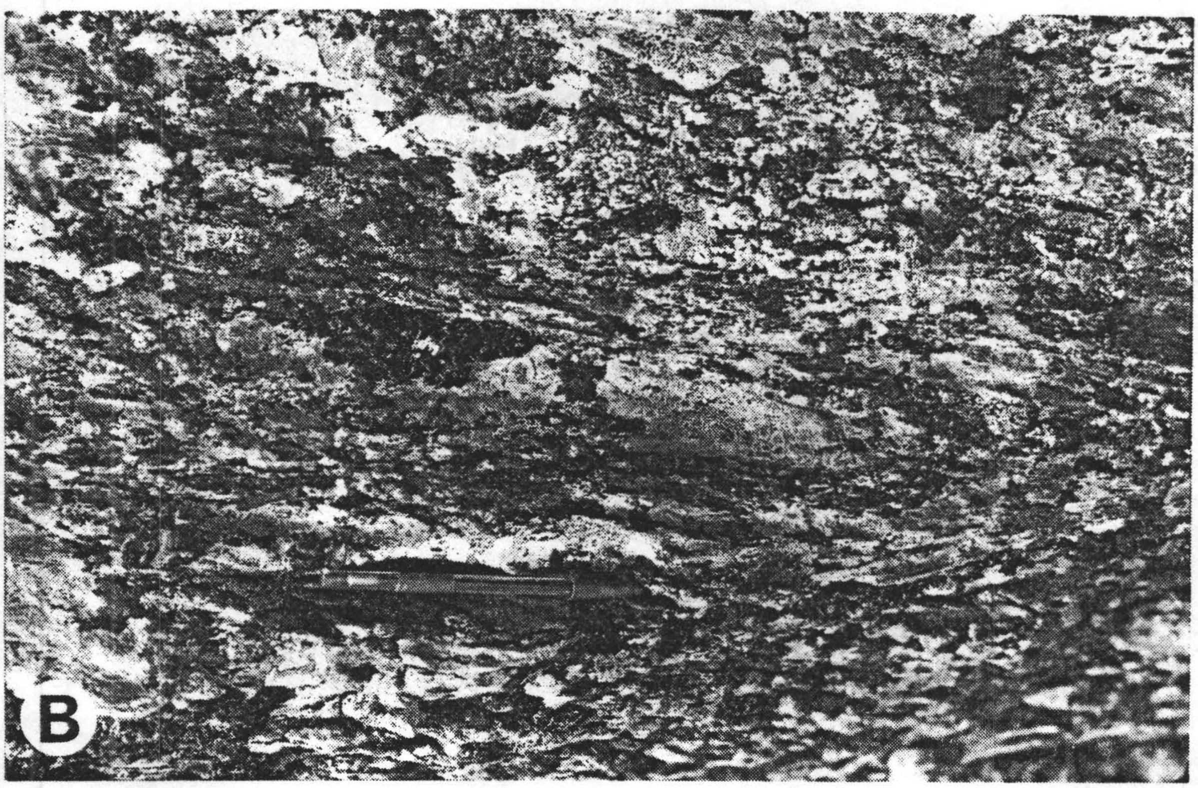
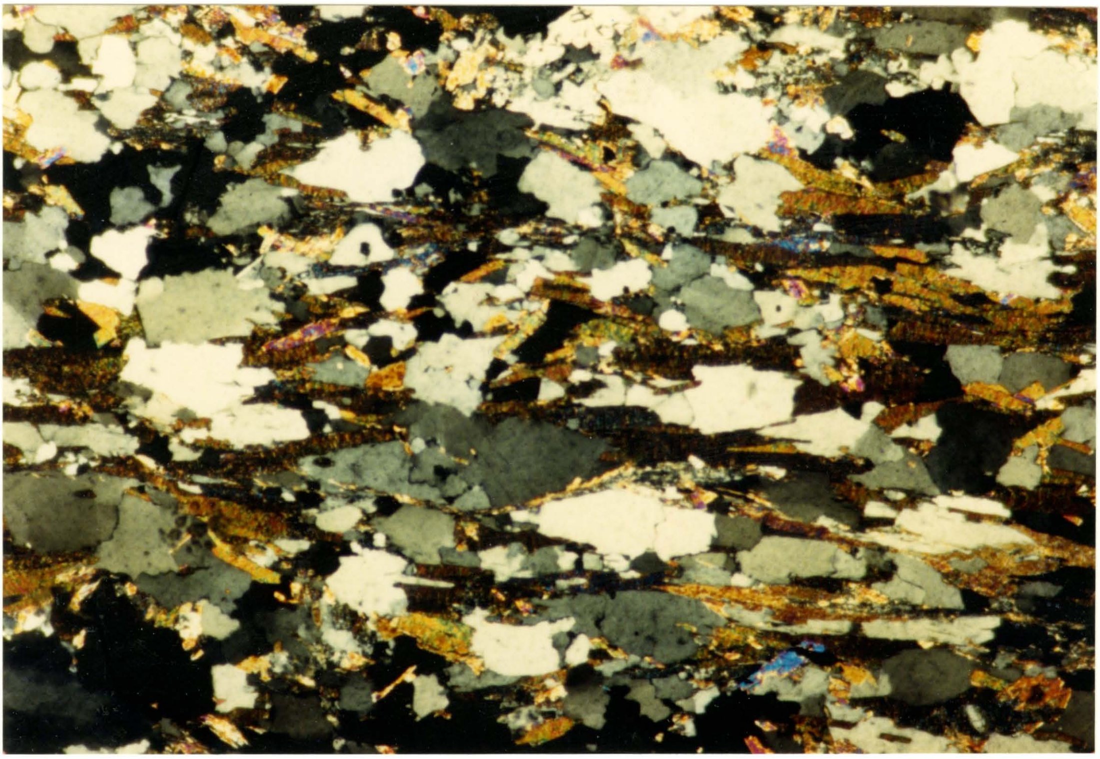


Plate 4-3A: S1 schistosity defined by alignment of micas and elongate recrystallized quartz and feldspars, semipelite to psammite composition. Field of view is 2.5 mm x 4.0 mm, crossed polars.

Plate 4-3B: Same photomicrograph as 4-1A but in plain light, emphasizing the schistosity. 2.5 mm x 4.0 mm, field of view.



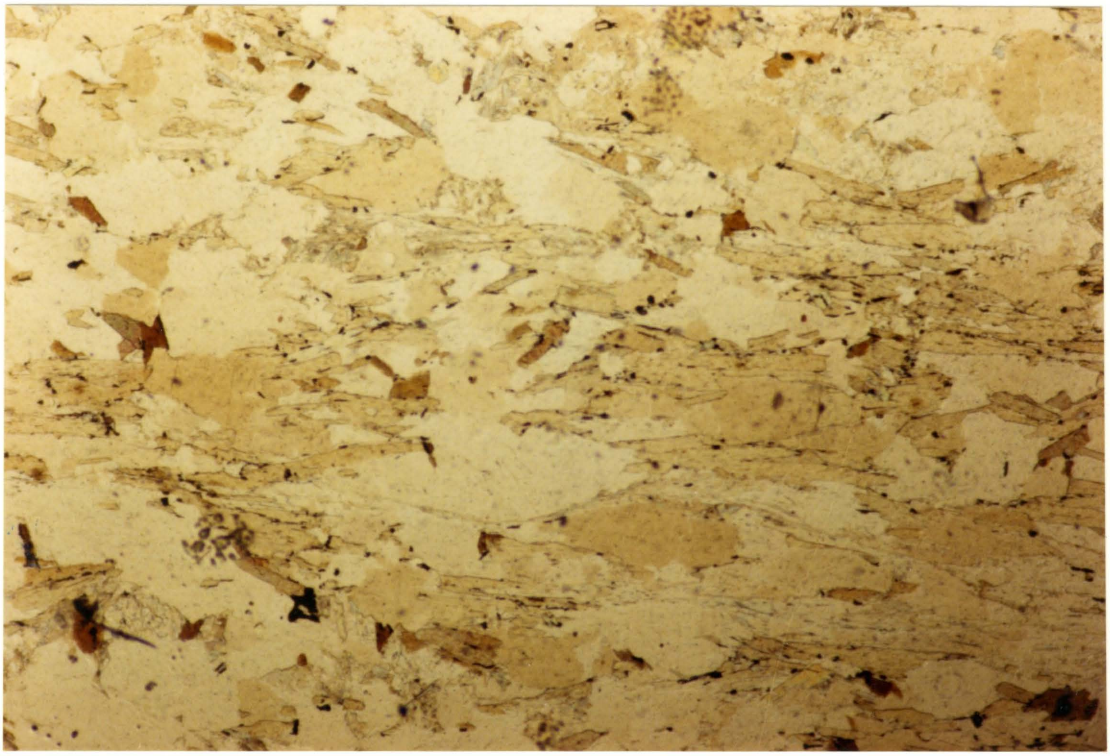
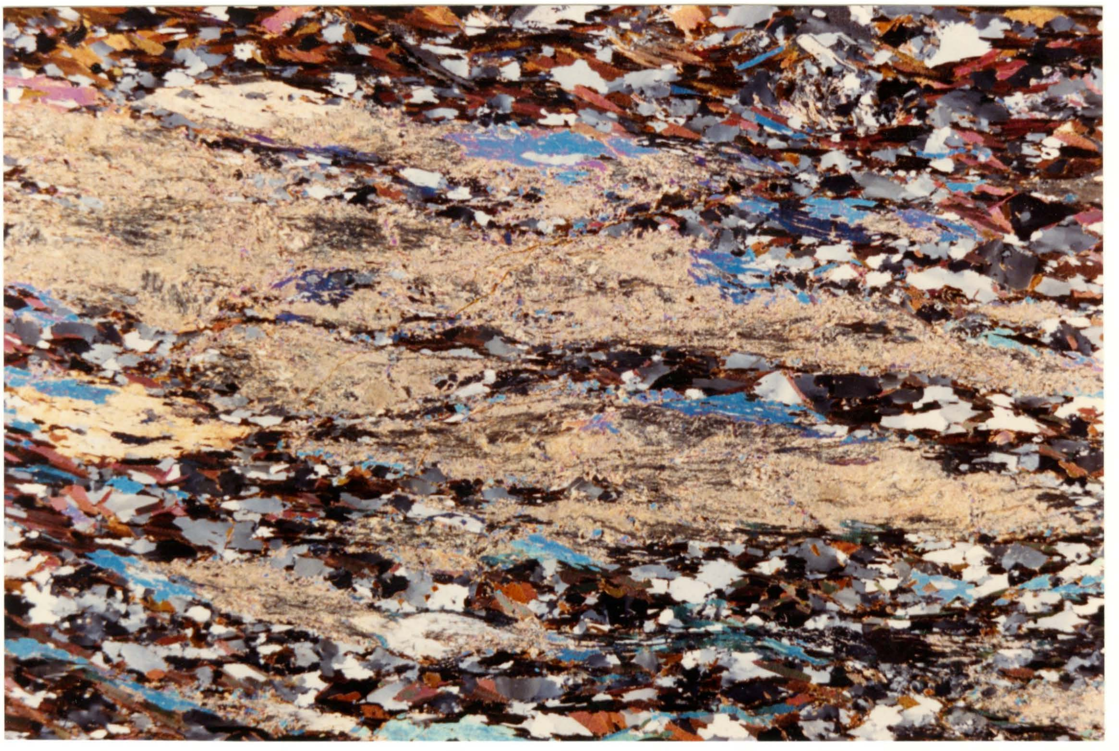


Plate 4-4A: S1 schistosity defined by alignment of muscovite and faserkiesel (quartz-sillimanite-K feldspar) pods in pelite lithology. Weak undulations of both micas and faserkiesel pods due to D2. Field of view is 7 mm x 12 mm, crossed polars.

Plate 4-4B: S1 schistosity in a pelite defined by a parallel alignment of micas, sillimanite and some quartz and feldspars. Weak undulations of S1 likely due to D2. Micas growing around garnets indicating pretectonic to syntectonic growth of garnets. Field of view is 7.3 mm x 10.9 mm, crossed polars.



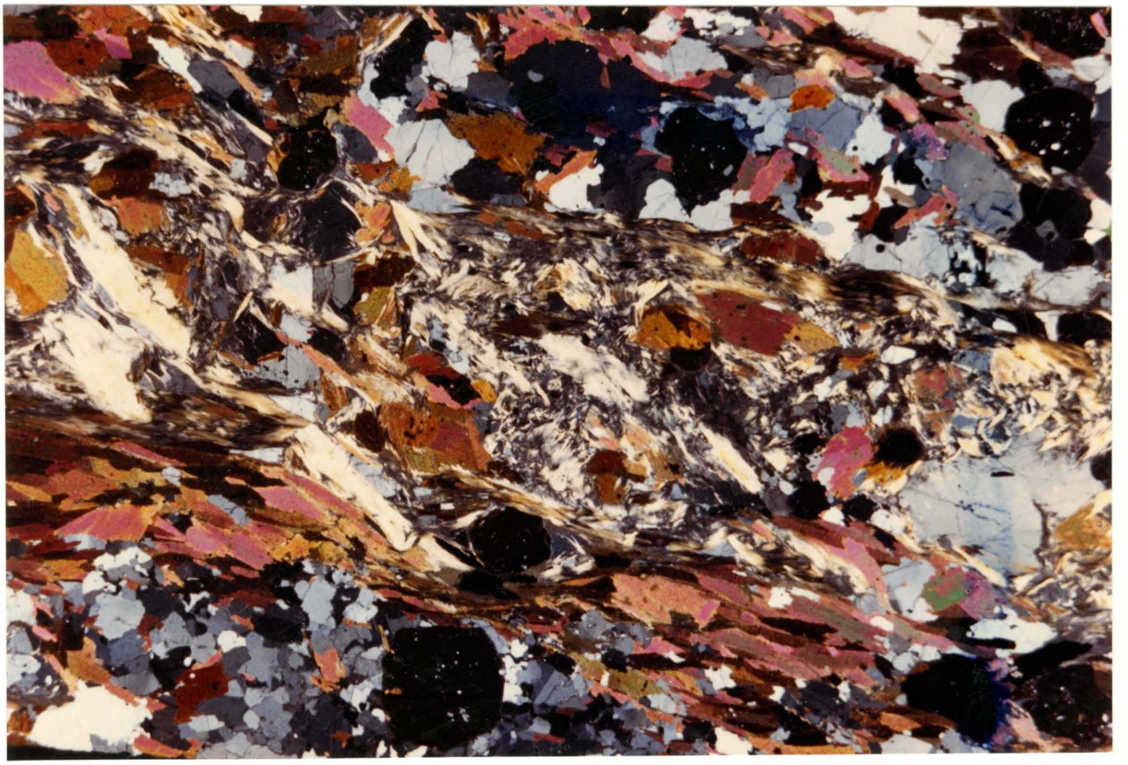
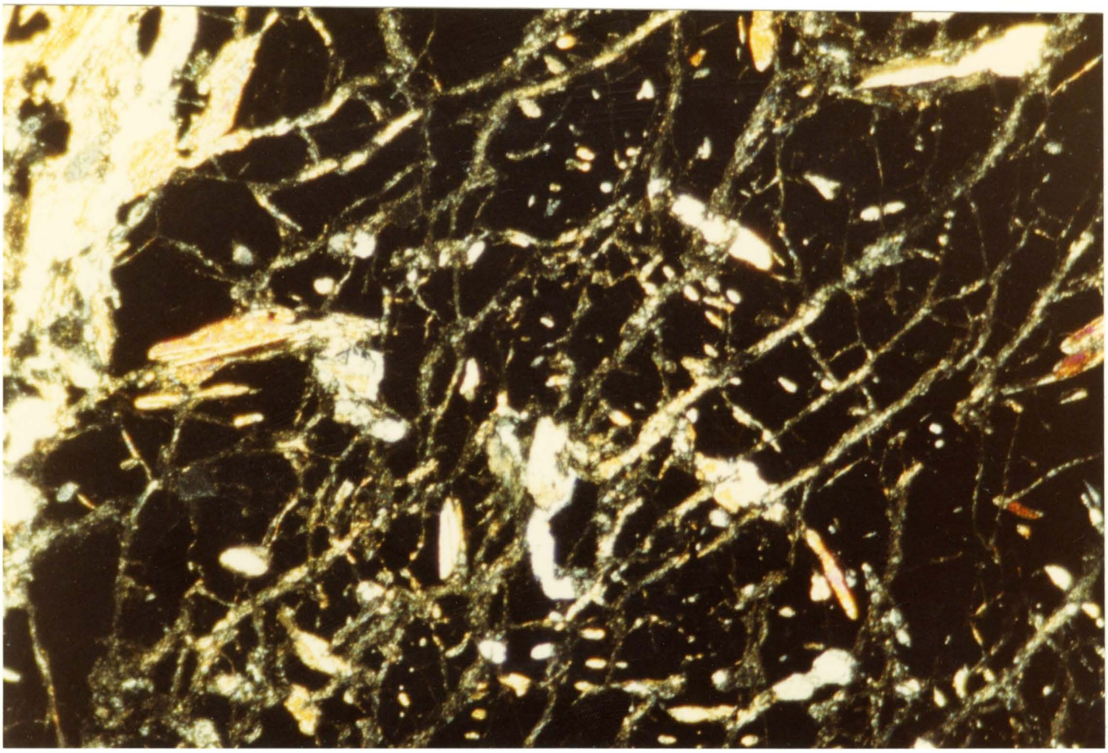


Plate 4-5: Coarse porphyritic, poikiloblastic garnet with quartz, muscovite, amphibole and opaque inclusions in a pelite. Inclusions are in spiral like orientations indicating syntectonic growth. Field of view is 2.5 mm x 4.9 mm, crossed polars.



lengths, giving a ribbed surface effect to the outcrop. Cleavage and bedding are found at oblique angles in the hinges of these folds; elsewhere they are parallel and indistinguishable.

Schistosity associated with D1, (S1) is pervasive throughout the study area and is most often defined by a parallel alignment of micas, sillimanite and hornblendes {plate 4-3A,B}. The micas are found growing into and around porphyroblasts such as garnets {plate 4-4A, 4-5 respectively}. Metamorphic faserkiesel pods, when present, are elongate parallel to S1 {plate 4-4B}. The schistosity is best developed in pelitic and semipelitic rocks.

A structural feature which may be attributable to D1 is pinch and swell structure. Although not common, when present their orientation always extends parallel to bedding and S1. They have not been observed in the hinge of a D2 fold and cannot be distinguished clearly from D2.

4.4 D2 Deformation

The second deformational event folded the earlier fabric and has resulted in the co-folding of basement and cover into an antiform which underlies the autochthon. As a result the allochthonous rocks in the study area are folded around a basement-cored antiform. Several deformation styles have

Plate 4-6A: D2 ptygmatic like folds of bedding and cleavage in allochthonous layered metasediments in study area. Shape controlled by the more competent quartzo-feldspathic layer (white). Lens cap diameter is 5 cm.

Plate 4-6B: Disharmonic D2 folds of bedding and S1 in allochthonous layered metasediments. Lens cap is 5 cm in diameter.

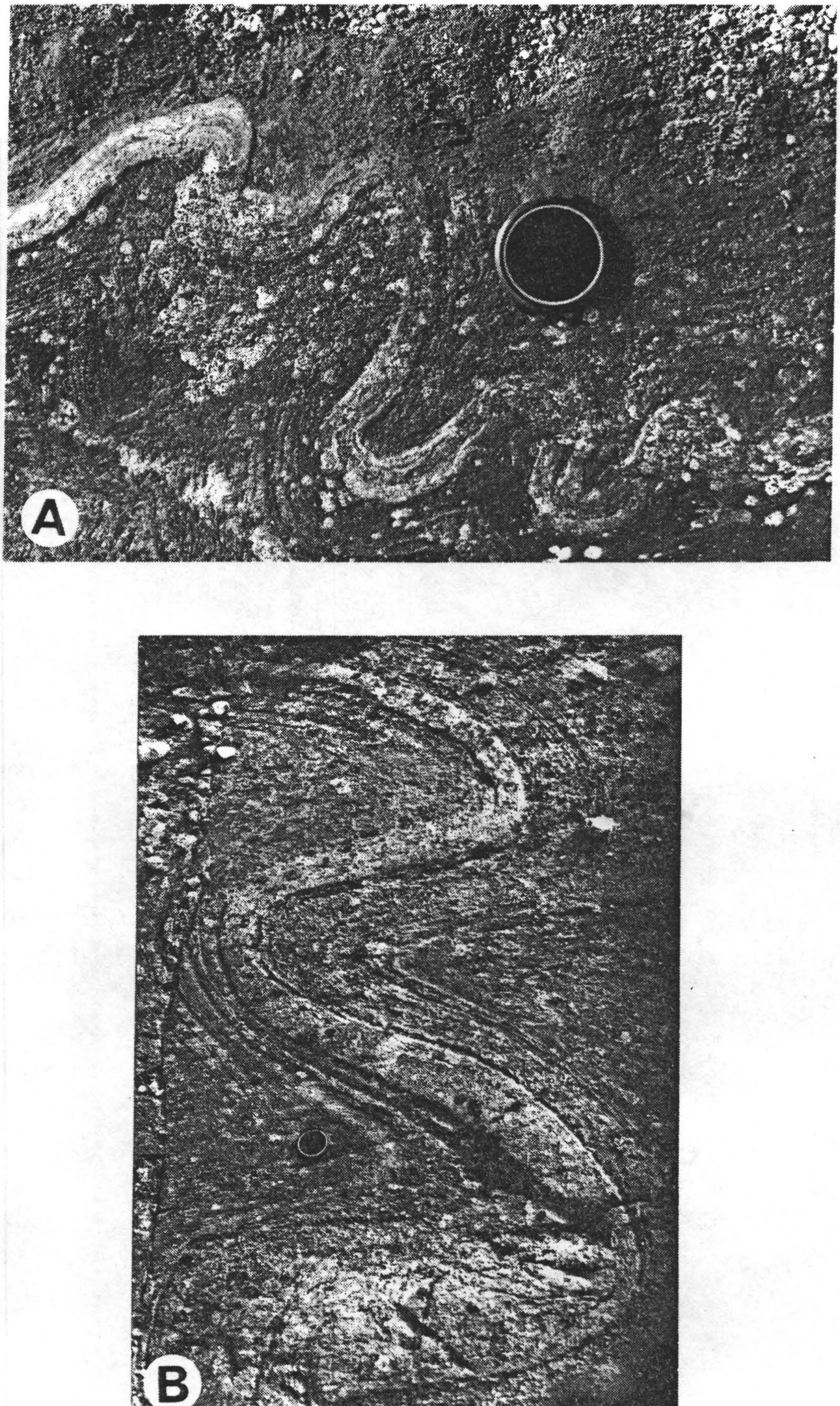
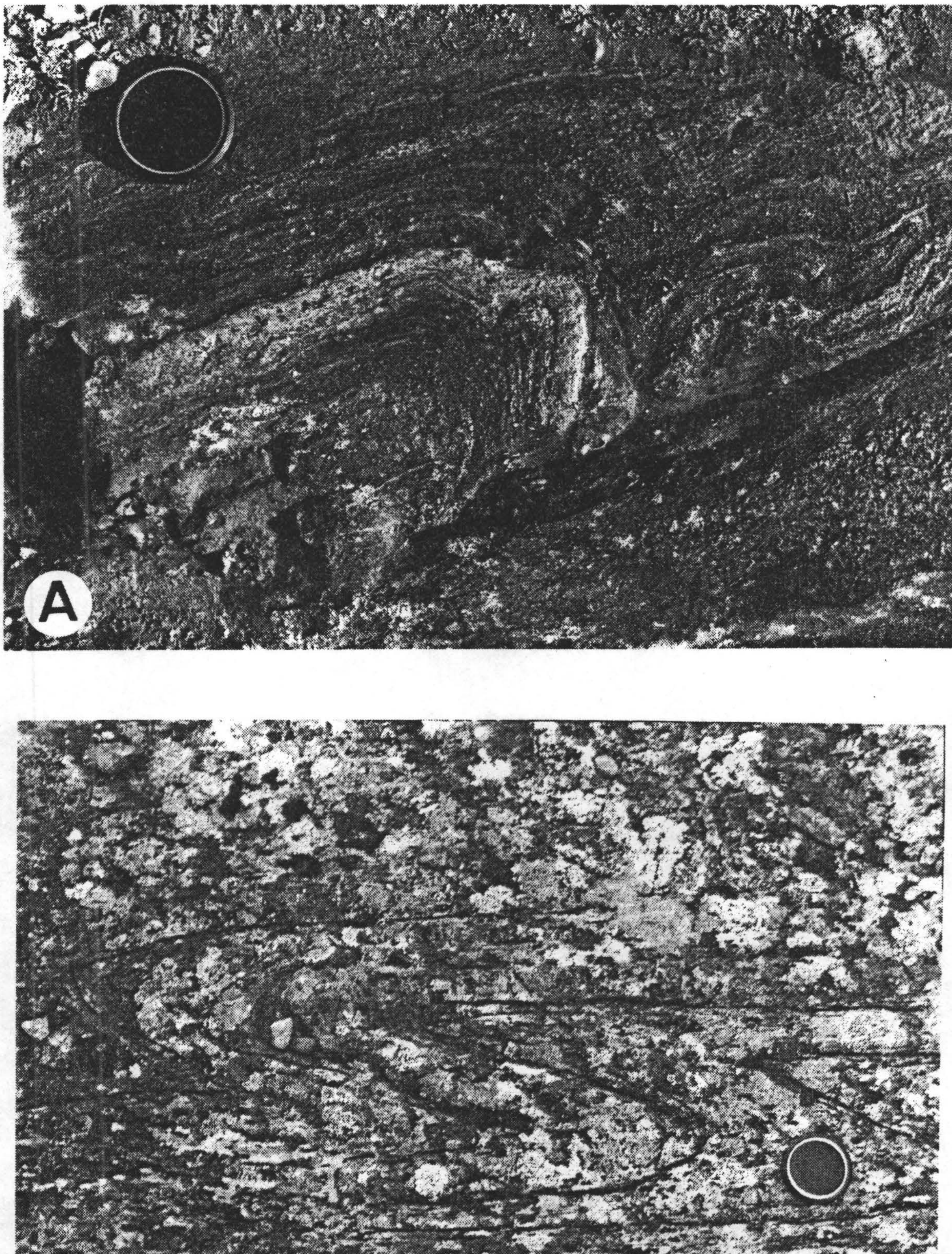

A
B

Plate 4-7A: Asymmetric D2 folds of bedding and folding immediately above the sole thrust in the central allochthon. The shape is controlled by the competent quartzo-feldspathic layer (white). Lens cap is 5 cm in diameter.

Plate 4-7B: Asymmetric "S" folding of bedding and cleavage in allochthon. Fold is typical of many asymmetric folds which are transitional between late D1 and early D2. Lens cap diameter is 5 cm.


A
B

been attributed to the second phase of deformation, partially a function of varying lithology.

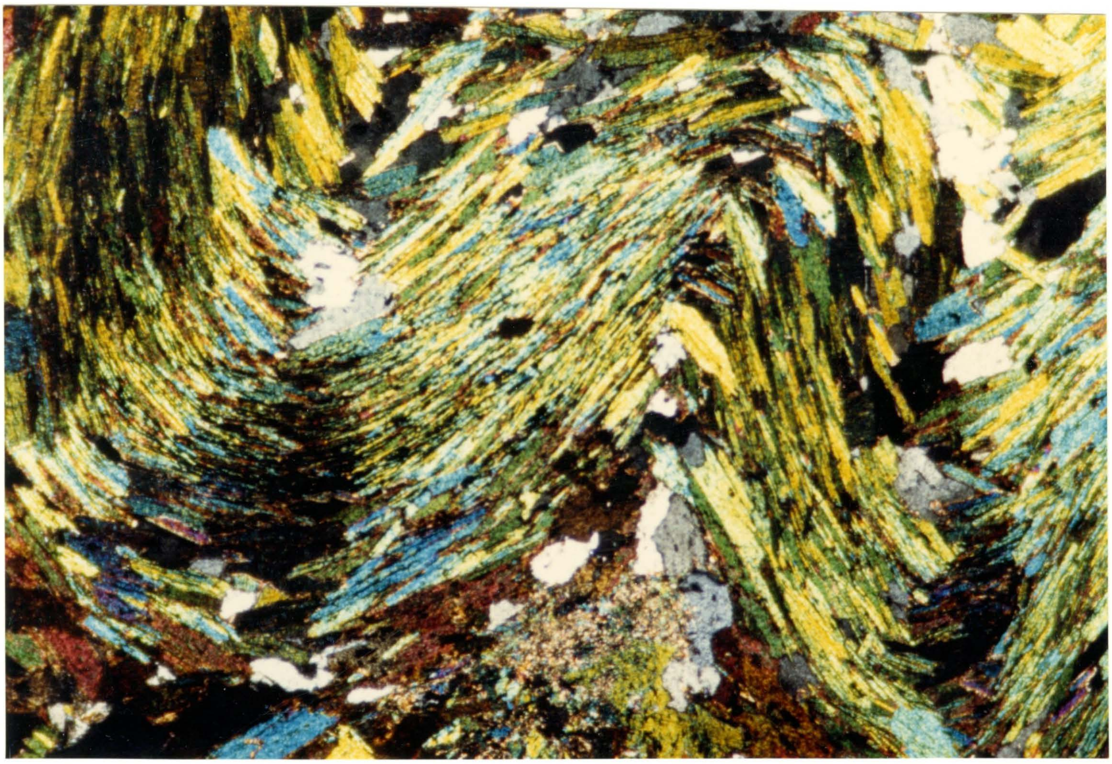
Ptygmatic style folding is found in psammitic sandstones with the more competent quartzo-feldspathic layers folded disharmonically, presumably as a result of buckling in a less viscous host rock {plate 4-6A,B}. Where observed, deformation involving bedding and cleavage (S1), has produced a suite of D2 folds plunging towards the southwest. In better layered sandstones, many layers are involved in disharmonic folding with the more rigid quartzo-feldspathic layers controlling the final shape. These folds characteristically deform bedding and S1 and have axes plunging consistently to the southwest.

Higher order asymmetrical folds are widely distributed, most commonly in the more micaceous lithologies {Plate 4-7A,B}. Their sense of asymmetry does not describe the large scale basement-cored antiform, but does describe, to some extent, the mesoscopic over-rotated D1 folds. However there are also folds which occur with no apparent relationship to any larger fold. All these folds always involve distortion of bedding and S1, and their folds axes almost always plunge southwest (v. section 4.7).

Gently undulating, open folds are also attributed to the second phase of deformation. They typically have an approximate amplitude of 1 m and wavelength of 5 m. They fold bedding, cleavage (S1) and in some places D1 isoclinal folds.

Plate 4-8A: Strongly developed asymmetric microfolds of S1 with associated zonal type in pelite. Note enrichment of quartz in nose relative to fold limb. Field of view is 2.5 mm x 4.0 mm, crossed polars.

Plate 4-8B: Incipiently developed microfold (crenulation) of biotite, muscovite and chlorite define schistosity (S1) in psammite. Chlorite is retrograde and post D1 in nature. Field of view is 2.5 mm x 4.0 mm, crossed polars.



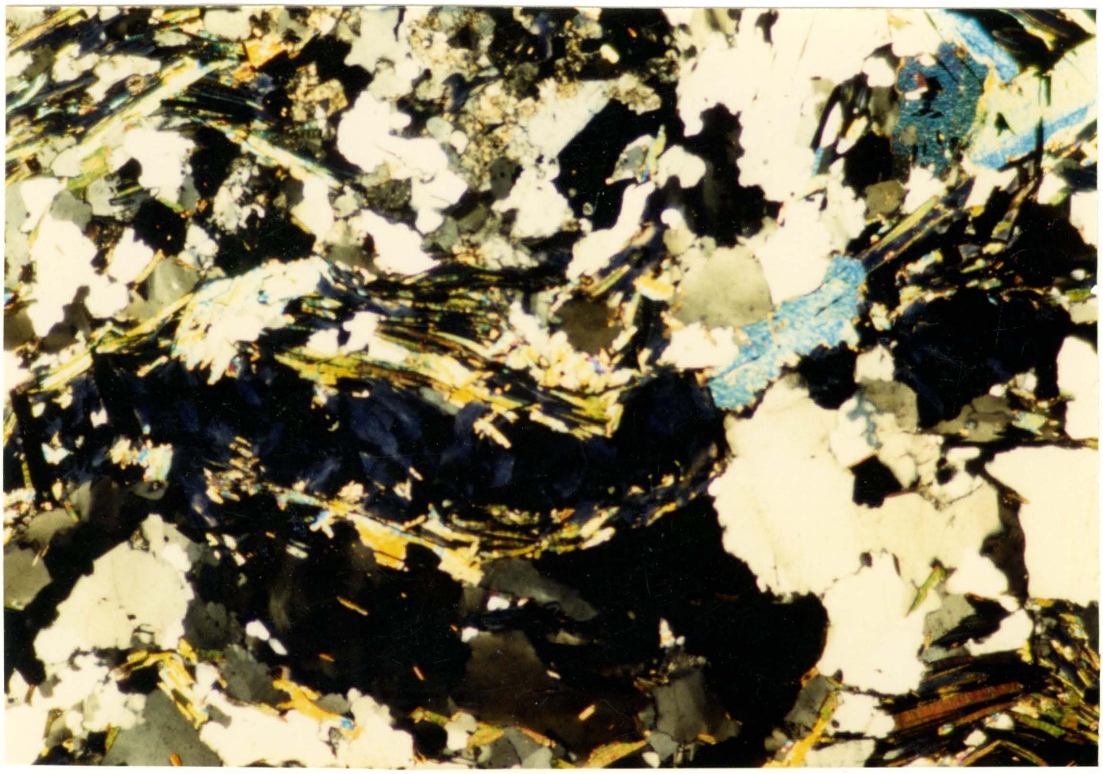
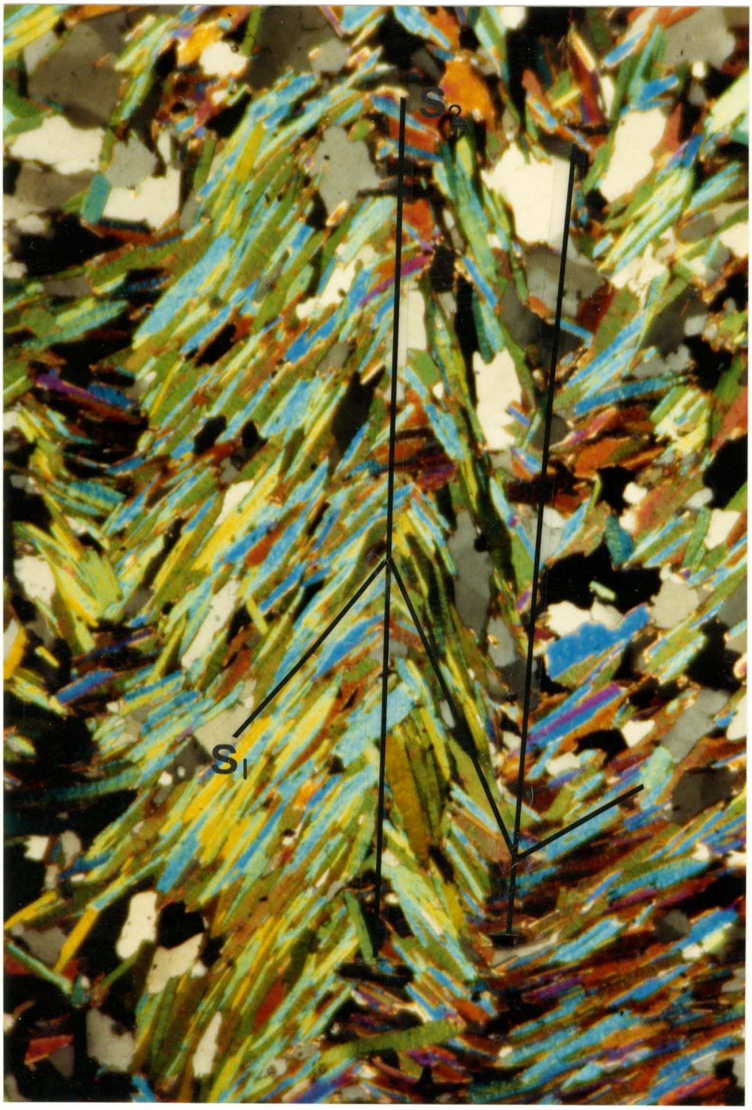


Plate 4-9: Zonal type crenulation cleavage (S2) corresponding to "micaceous" zones coincident with the limbs of asymmetric microfolds in semipelite. The micas within the cleavage define the original schistosity (S1) and are oblique to the boundaries of the zone. Field of view is 4.5 mm x 3.1 mm.


S2
S1

Like most D2 fold structures, the plunge of their folds axis is dominantly to the southwest.

Crenulation cleavages (S2) and microfolds have been developed in numerous localities across the study area. Crenulation cleavages are thought to develop as a result of buckling an anisotropic material (microfolding). Mineral migrations responsible for the cleavages form in response to stress gradients established during the development of these buckling instabilities [Cosgrove, 1976]. In some instances, they are strongly developed while only incipiently developed in others. More easily noticeable is the associated corrugation of S1 and crenulation lineation {Plate 4-8 A,B}. The trends of these microfold axes (F2) are characteristically to the southwest and commonly coincide with the trends of the mesoscopic folds with which they are associated. Microscopically, the crenulations are asymmetric and their limbs are defined by a concentration of micas, fine grained lenticular quartz and feldspar. In thin sections with strongly developed crenulations, the limbs are distinctly quartz-feldspar poor and mica rich when compared to the matrix encompassing the crenulations. Several of the thin sections examined are marked by very strong crenulations and a good axial planar crenulation cleavage. The cleavage is generally prominent on the more steeply dipping limb {Plate 4-9}. These cleavages occur as wide, diffuse cleavages zones with indistinct boundaries. According to Gray [1979] this is known as zonal type crenulation cleavage. It is distinguished from the discrete type of crenulation cleavage which has thin, sharply defined cleavage discontinuities which truncate the initial fabric. Discrete type crenulation cleavage is not present in any thin section from the study area. According to Gray [1979] discrete crenulation cleavage is at a more advanced stage of zonal cleavage and a transition should exist between the two. This model will be utilized in a later section when interpretations of the nature of crenulation cleavage present will be made.

The last major structural features which can be attributed to D2 are the moderately plunging mineral lineations present throughout the allochthon, autochthon and basement. Though differing in mineral composition, the lineations are oriented alike in allochthon, autochthon and basement, all plunging moderately to the southwest. In the basement a pronounced lineation, composed of streaks of recrystallized quartz and feldspar, and, on occasion hornblende, is present and also coaxial with D2 folds [King et al., 1987]. In the allochthon the most common lineations are composed of sillimanite crystals and rods of quartz and/or feldspar. Also present are lineations comprised of hornblende and biotite. With very few exceptions, all mineral lineations, crenulation lineations and intersection lineations, from the basement [King et al., 1987] and cover alike cluster fairly tightly in the south west quadrant of an equal area stereographic projection. This is of special significance in determination of their timing.

4.5 D3 Deformation

No direct evidence of D3 has been observed in the study area, but is inferred from the doubly plunging nature of the Exmouth Antiform and other basement-cored antiforms and synforms (refer to FIG 4.3). A Type 3 interference pattern interference pattern [Ramsay, 1967] is present regionally, at the 10 km. scale. The D3 fold which affects the study area is known as the Tree River Cross Arch [Hoffman et al., 1986]. On the southern limb of this antiform, the D2 Exmouth plunges approximately 35° to the southwest.

4.6 Overprinting Relations, Complex Fold Patterns

Most of the complex structural features present are the result of interference between structures produced by D1 and those produced by D2. Interference patterns produced by D3 have not been observed within study area. Regional knowledge of the Metamorphic Internal Zone indicates that the Exmouth Antiform is on the southern limb of a broad east-north-east D3 antiform. Thus D3 had the effect of rotating all structural attitudes to the southwest producing an oblique orientation of D2 and D1 structures.

The earliest overprint relationship present is that between bedding and S1 schistosity. These features can be clearly distinguished from one another only in the hinge zones of D1 isoclinal folds. Here, it is apparent that the schistosity is axial planar. Its consistent orientation parallel to the axial planar surface of the D1 folds places the timing of this schistosity as being syntectonic to D1. Another important point to emphasize is that some D1 isoclinal folds and other macroscopic D1 folds involve folds of bedding and schistosity. This indicates that deformation continued beyond the time of formation of the isoclinal folds with axial planar schistosity. For reasons to be discussed shortly they can still be distinguished from the D2 deformation.

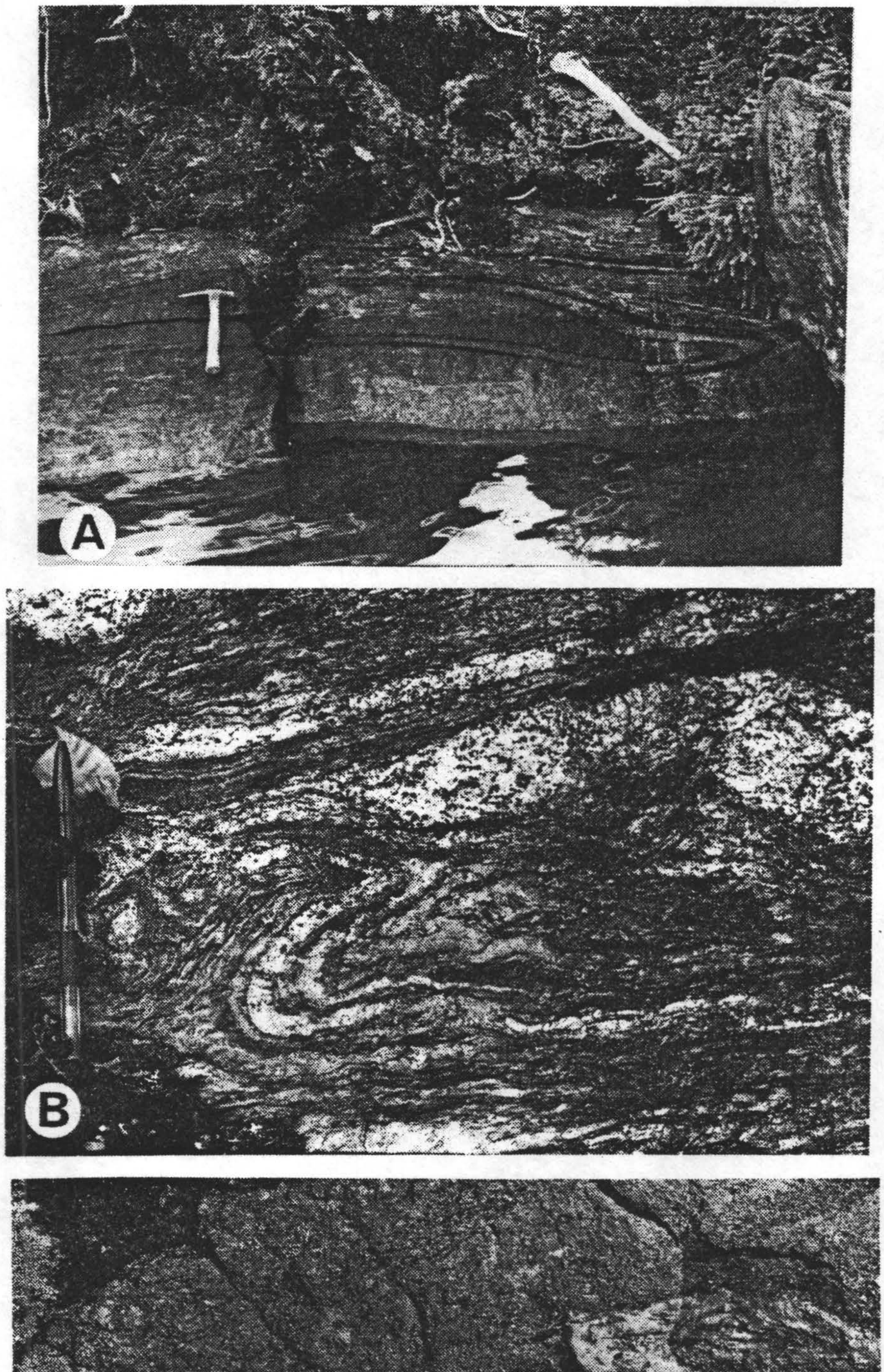
Crenulations of S1 about D2 folds axes, sometimes with the formation of a crenulation cleavage (zonal type) are found at many localities in the study area. Associated with the crenulations are crenulation lineations. The orientation of the lineations align consistently with other linear elements such as minor fold axes {Plate 4.10C} and mineral lineations measured in the allochthon as well as the basement [King et al., 1987].

On a mesoscopic scale many interference patterns between D1 isoclinal folds and later D2 folds occur {Plate 4-10A,B}. Most of these mesoscopic fold patterns are characteristic of a Type III interference pattern [Ramsay, 1967]. The best

Plate 4-10A: Type III interference pattern involving tight to isoclinal D1 fold and open D2 fold in layered metasediments. West islands of study area. Isoclinal fold is doubly plunging. Hammer length is 40 cm.

Plate 4-10B: Complex multifolded layers indicating Type III interference pattern in layered metasediments. Allochthonous units at decollement in study area. Pen is 15 cm in length.

Plate 4-10C: Folding of quartz-muscovite-sillimanite layer and schistosity (S1) by D2 fold. F2 axial plane at high angles to layering (S1 bedding). Scale is in centimeters.


A
B
C
Cm

examples involve the folding of isoclinal folds of bedding with an associated axial planar schistosity folded about fold axes plunging moderately to the south-west. The axial orientations of the latter fold axes are typical of D2. The isoclinal folds are typical of D1. Therefore these interference patterns are interpreted as D1 folds folded over a D2 fold.

Mineral lineations are important in elucidating the deformational history of the study area. Their consistent orientation towards the southwest in the study area, regardless of mineralogy suggests a D2 origin. This direction is parallel with the macroscopic basement fold axes. Nowhere in the study area or basement are they observed curved about a minor D2 oriented fold axes. That they occur with little variation, except perhaps in mineralogy, both in the allochthon and rocks of the basement and autochthon (the latter two show no indication of D1 strain) and are parallel to the macroscopic basement fold axes indicates an origin coeval with deformational strain in the basement, that is, D2.

Lastly, on a macroscopic scale the over-rotated thrust folds interpreted to be syn D1 are clearly affected by the basement-cored antiform. Examination of FIG 4.2 shows that these calculated axial traces clearly wrap around the nose of the D2 basement fold. This suggests that these axial traces may have been distorted from some prior orientation during production of that antiform.

4.7 Stereographic Projections

Several stereographic projections have been produced from structural data collected in the field. These projections are confined to mesoscopic and macroscopic linear elements (fold axis, crenulation lineations and mineral lineations). For comparison and interpretation purposes, a stereographic projection of linear elements from the basement [King et al., 1987] is included. Examination of these reveals a confirmation of interpretation of deformation phases just noted. These projections are presented in FIG 4.4.

The plot of all minor fold axes shows a wide variance in orientations with a conspicuous clustering of points in the south west quadrant {FIG 4.4}. However, after separating D1 and D2 fold axes based mainly on the distinct deformation styles previously noted, a clear pattern becomes evident. FIG 4.4 includes a plot of linear elements from the alochthon which include fold axes interpreted to be D2 folds. This plot shows a fairly good clustering of points in the southwest quadrant with not a great deal of variance. A similar such configuration occurs with the fold axes measured in the basement {FIG 4.4}. However, the clustering of points is tighter for the basement fold axes than those of the allochthon. If it is accepted that both plots truly represent D2 fold axes, then the wider variance in allochthonous orientations can be explained as resulting from a modifying effect of pre existing D1 surfaces, whereas no such D1 structures have been observed in the basement. According to Ramsay [1967], the directional stability of any new fold axes is in part, affected by the attitudes of pre existing surfaces. Among the various resultant D2 fold axes orientations possible, he suggested that when the angle between the axial surfaces of the new folds (D2) and surfaces undergoing folding (D1 fold limbs) is small, then the spread of F2 fold axes will be small {FIG 4.5}. Implications derived from these stereographic projections are consistent with those derived from the various complex fold patterns present in outcrop. The implications are that D1 and D2 were approximately coaxial, D1 producing recumbent folds, D2 producing upright folds which once superimposed, locally produced a Type III interference pattern [Ramsay, 1967].

The allochthon and autochthon are of similar lithologies and should respond in approximately the same way for a given applied stress. The autochthon shows none of the effects resulting from D1 that are so in the allochthon [King et al., 1987]. It follows therefore that the D1 episode affected <u>only</u> the allochthon. It further follows that the minor fold axes, mineral lineations and crenulation cleavages, all ap-

FIG 4.4: Orientation of Linear Elements in the Study Area

Orientations of measured mesoscopic linear elements in basement (a) [King, 1987] and allochthon (b). Note that the plots are very similar for D2 fold axes, mineral lineations, and crenulation lineations in basement and allochthon. D1 fold axes in allochthon are generally random. Plot of all D2 fold axes in allochthon (c) showing little variation from west to east as shown by plots (d, e, f respectively) on either limb of the basement D2 antiform. The plots are seperated by generally north-south dashed lines. The tight clustering of D2 fold axes in allochthon indicate a high axial direction of stability.
Symbols; dots- mineral lineations, open triangle- F1, closed triangle- F2, squares- crenulation lineations.

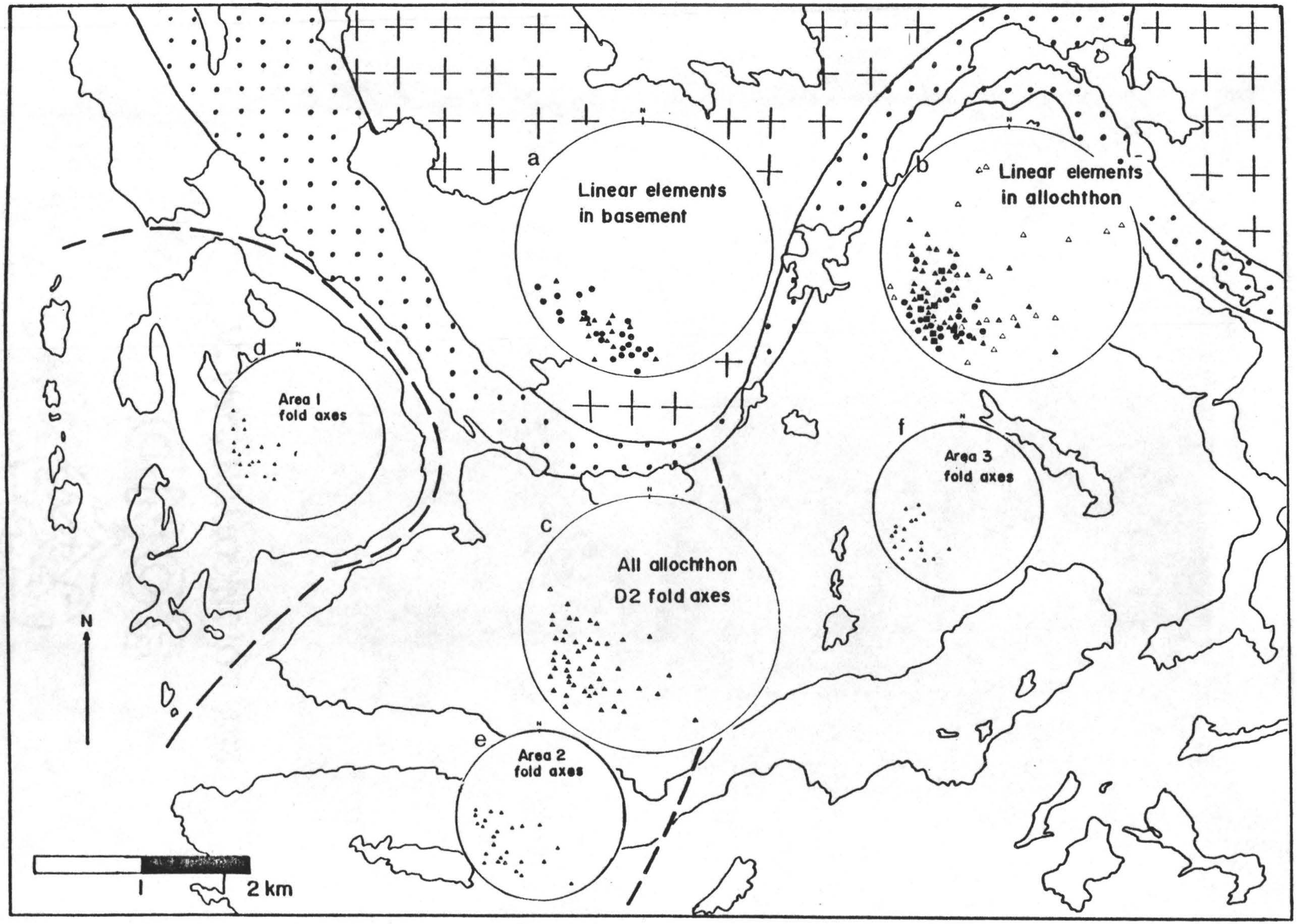
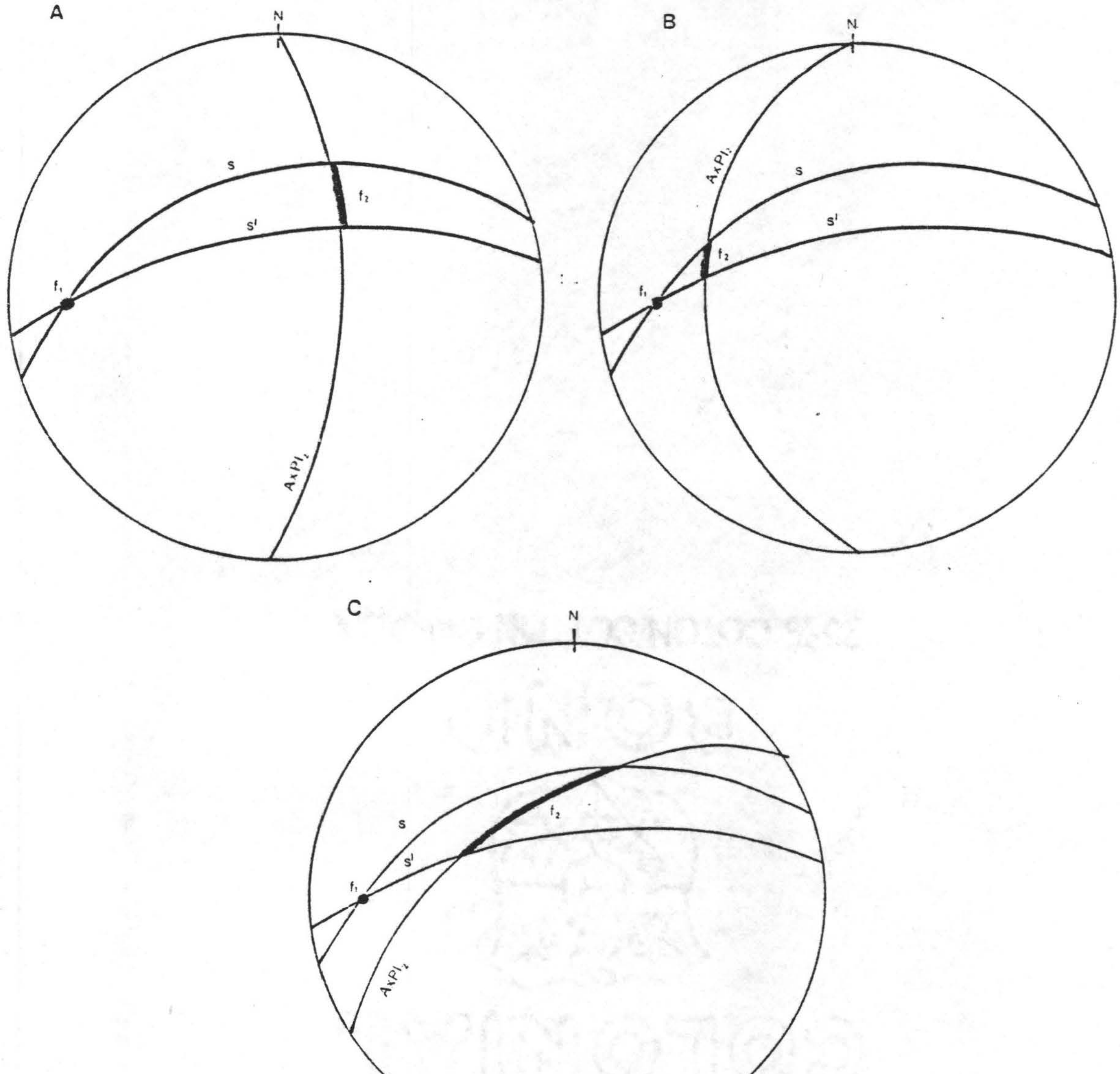

a
Linear elements
in basement
b
Linear elements
in allochthon
c
All allochthon
D2 fold axes
d
Area 1
fold axes
e
Area 2
fold axes
f
Area 3
fold axes
N
1
2 km

FIG 4-5: Theoretical Axial Direction Stabilities of Second Folds (D2) Developed on Curving Surfaces

Variations in attitudes of the surfaces on which new folds are developed are developed into the new folds (D2) as a variation of axial directions. Where the axial directions of the new folds make high angles with the surfaces undergoing folding (which would be the case with coaxial folding, Type III resultant interference pattern, FIG 5.5d) any initial variation in the orientation in surfaces being folded results in a change in pitch of the fold equal to (FIG 5.5a) or less (FIG 5.5b) than that of variation of the initial surface. Thus the D2 fold axes, F2, have a high axial direction stability. This is the case in the study area. If the initial angle between the axial surfaces of new folds and surfaces being folded is small, (FIG 5.5c) there will be a great variation in axial direction. This style is not present in the study area [modified from Ramsay,1967].


A
N
s
s'
f₁
f₂
AxPl₂
B
C

proximately co-axial with the basement-cored antiform, are D2 features. Thus the segregation of linear features from the allochthon into D1 and D2 is plausible. Hence, by comparing the plots of F2 for allochthon and for basement, the conclusion that the increased variance in allochthonous F2 orientations is due to pre existing recumbant F1 folds, coaxial with F2, is a very likely one. It explains most of the local structural data obtained; it agrees with the current tectonic model for the area.

4.8 Synthesis

Three episodes of deformation have been documented in the study area. The first episode occurred as a result of east-west shortening and thrusting of allochthonous rift deposits over the Slave Craton and Proterozoic cover. It produced recumbent isoclinal folds with axial planar cleavage, isoclinal folds of bedding and cleavage and overrotated thrust folds.

It was noted earlier that some asymmetric D2 type parasitic folds are found on the limbs of the D1 overrotated thrust folds. Rather than disprove previous interpretations, it makes necessary the adoption of a continuous east-west compressive deformation model (the Calderian Orogeny), as opposed to discrete D1 and D2 events. This is also made apparent by the occurrence of D1 isoclinal folds, some with axial planar cleavage, but others with folded bedding and cleavage. D1 then, was a continuous deformation within the Calderian Orogeny, with a transition of structural styles with continued deformation. It graded continuously into D2, with which it was coaxial. It should be more appropriate therefore to talk in terms of early Calderian deformation and late Calderian deformation.

By the late stages of the Calderian Orogeny, (D2), thrusting had ceased and the basement and autochthonous cover were folded into a large scale broad antiform. Evidence for the completion of thrusting by this time is the way in which the axial traces of the over-rotated thrust folds are folded around the D2 antiform. In light of the previous discussion the time interval between the termination of thrusting and commencement of "D2" folding was very small and was probably not a discrete transition but a continuous one.

At this point it is useful to tie in the maximum prograde metamorphic grade attained, with the structural history. As previously noted, prograde metamorphic conditions are thought to be pre D2 and post D1. Evidence suggests that the metamorphic culmination occurred immediately prior to the folding of the basement (late Calderian deformation). This evidence includes;

(1) the (S1) schistosity was onced defined by elements of the prograde metamorphic assemblages, as indicated by the occurrence of faserkiesel.

(2) the pretectonic {Plate 4.4A} and syntectonic growth of garnets {Plate 4.5}, as indicated by inclusions in garnets and the behaviour of S1 adjacent to these porphyroblasts, suggests that metamorphic mineral growth (garnets etc., at low to medium greenshist grade followed by higher grade muscovite, sillimanite etc. [Turner, 1980]) before and during D1. Geobarometry and geothermonetry on garnets is very useful in determining metamorphic history of a rock (v. St. Onge, 1984c) but is beyond the scope of this thesis.

(3) the folding of faserkiesel pods in places by "D2" folds. Since faserkiesel is thought to be a result of retrogression from prograde sillimanite-K feldspar metamamorphic assemblages in the study area, then maximum prograde metamorphism must have occurred prior to this D2 folding.

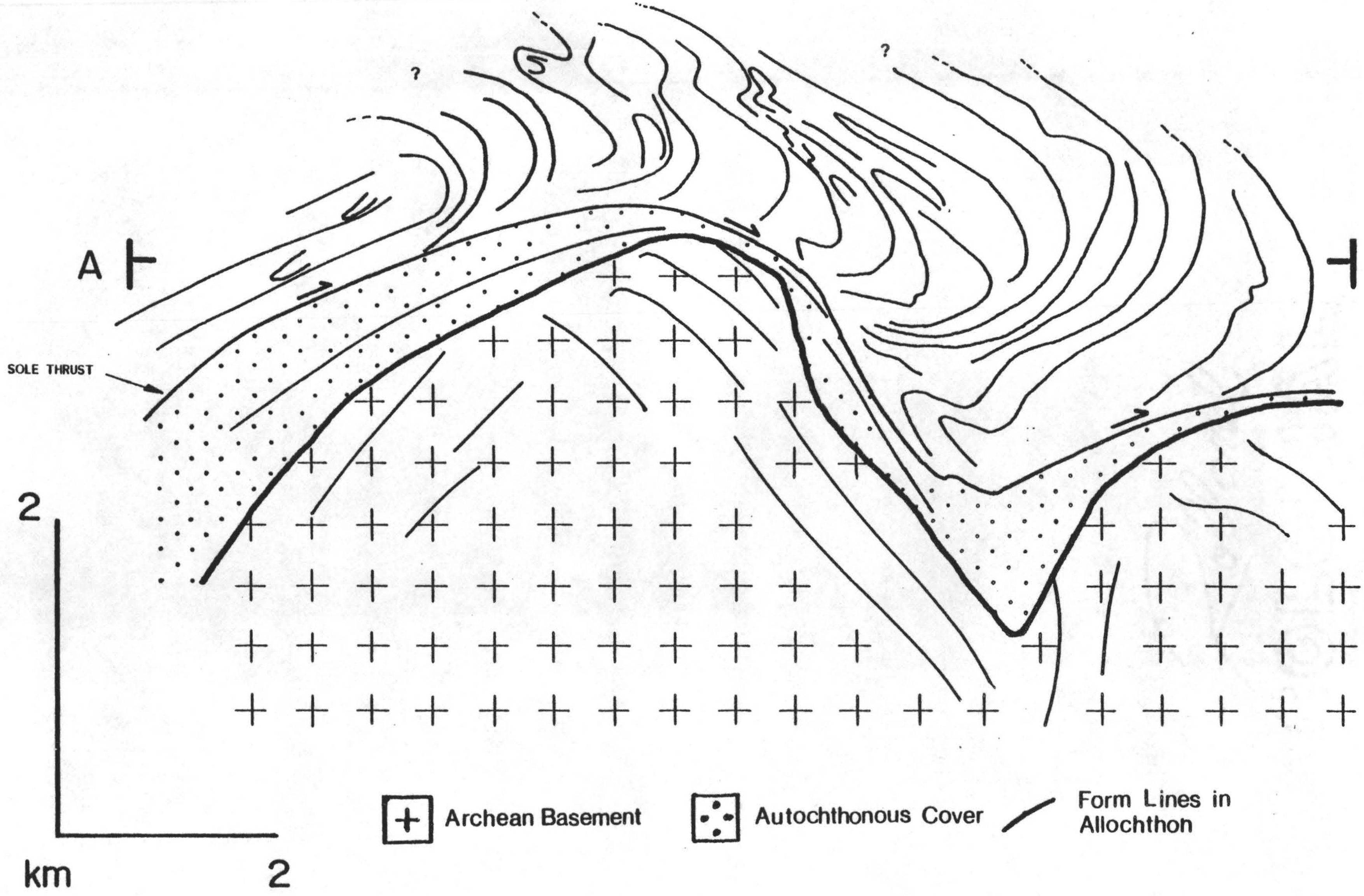
(4) the parallelism displayed by mineral lineations with the F2 fold axes, implying that temperatures necessary for the partial recrystallization and remobilization of these minerals had just occurred (late D1)

(5) the parallelism of parasitic F1 folds on the over-rotated D1 thrust folds with typical F2 orientations suggesting a close temporal relationship between D1 and D2

(6) the fact that zonal type crenulation cleavage is observed and coeval with D2; the more advanced discrete type is absent. This suggests that induced strain was not high in D2, perhaps a result of low metamorphic grades (lower than in D1).

FIG 4.6: Down Plunge Cross Section of Study Area

Post folding cross section showing modification of D1 over-rotated thrust folds by D2 basement folding. Line of section A-B is shown in FIG 4.2.



?
?
A
B
SOLE THRUST
2
km
2
Archean Basement
Autochthonous Cover
Form Lines in Allochthon

Thus the metamorphic characteristics can be correlated to the structural history as presented. It complies to both metamorphic and structural features in the study area and is consistent with the regional model of emplacement of the hot Hepburn Batholith suite in the allochthon coevel with thrusting (D1).

The last major deformational event, D3, was temporally discrete from the first two. A completely different direction of compression is implied by the east-north-east trending fold axes producing the regional "saddles" and "culminations" in the basement. All earlier structural elements located on the southern limb of this late cross folds were rotated into a south westerly plunge accounting for the map expression of the originally upright Exmouth Antiform. The plunge of the southern limb of the cross fold is interpreted to be 35°. Looking down the plunge of the Exmouth Antiform, the 35° plunge has been removed in order to visualize the structures in right (transverse) cross section {FIG 4.6} [methodology from Mackin, 1950]. This method assumes cylindrical folding, which, clearly is true only to a first approximation. Despite this, the schematic cross section displays reasonably the interpreted macrostructures of the study area.

CHAPTER 5 : CONCLUSIONS

Metamorphism in the allochthonous rocks above the Exmouth Antiform in the Wopmay Orogen is Buchan type (low P) in nature with muscovite-sillimanite and sillimanite-K feldspar zones present in sequence. Metamorphic conditions in the study area may be classified as varying from medium to high grade [Turner, 1980] or upper greenschist to lower amphibolite. Metamorphic grade increases systematically from east to west across the study area. Minimum pressure and temperature estimated from the lowest grade assemblages in the study area are on the order of 550°C and 2.1 Kbar. Estimates from the higher grade assemblage suggest minimum temperatures of 630°C and pressures of 1.7 Kbar. The current tectonic model suggests a "hot side up" metamorphic gradient resulting in increased metamorphic grades with decreasing depth. It is not possible to resolve whether metamorphic grades encountered in the study area are the result of a "hot side up" metamorphic gradient or a normal geothermal gradient. A larger study area would be necessary to resolve this issue.

The first identifiable deformation event produced significant crustal shortening by thrusting and associated recumbent, isoclinal folding and over-rotated thrust folds. An axial planar schistosity was produced during the early stages of this deformation event. A second, roughly coaxial event folded both cover and basement at all scales of observation. A significant layer parallel extension occurred at this time. A late cross folding event provides an oblique view through the crust on a major D3 fold limb in the study area. The structural relief of over 6 km so provided allows the direct observation of the results of the two preceding deformation events.

The deformation/metamorphic pathway determined resembles other documented orogenic belts, such as the Eastern Alps [Hoffman et al., 1986] and the Eastern Cape Smith Belt, Quebec [Scott, 1986]; where crustal thickening by thrusting (D1) leads to a thermal culmination. This is followed by a coaxial folding event (D2) during cooling by uplift and erosion. An overall westward transport is suggested by allochthonous thrust geometry and lithology.

The observations and results of this study are consistent with the current tectonic models proposed for the area [King et al., 1987]. The metamorphic characteristics and polyphase deformational characteristics describe those of a hot allochthon being thrust above a cold basement with continued compression.

REFERENCES

Bowring, S.A., Van Schmus, W.R. and Hoffman, P.F. 1984. U-Pb Zircon Ages from Athapuscow Aulacogen, East Arm of Great Slave Lake, N.W.T., Canada; Canadian Journal of Earth Sciences, v.21, pp. 1315-1324.

Breaks, F.W. and Shaw, D.M. 1973. The Silent Lake Pluton, Ontario: A Nodular Sedimentary, Intrusive Complex; Lithos, v.6, pp. 103-122.

Carmichael, D.M. 1969. On the Mechanism of Prograde Metamorphic Reaction in Quartz-Bearing Pelitic Rocks; Contributions to Mineralogy and Petrology, v.20, pp. 244-267.

Cosgrove, J.W. 1976. The Formation of Crenulation Cleavage; Journal of the Geological Society of London, v.132, pp. 155-178.

Easton, R.M. 1981a. Geology of Grant Lake and Four Corners Lake Map Area, Wopmay Orogen, District of MacKenzie; in Current Research, Part B, Geological Survey of Canada, Paper 81-1B, pp. 83-94.

Easton, R.M. 1981b. Stratigraphy of the Akaitcho Group and the Development of an Early Proterozoic Continental Margin, Wopmay Orogen, Northwest Territories; in Proterozoic Basins of Canada, ed. Campbell, F.H.A., Geological Survey of Canada, Paper 81-10, pp. 79-95.

Easton, R.M. 1982. Tectonic Significance of the Akaitcho Group, Wopmay Orogen, Northwest Territories; unpublished Ph.D. Thesis, Memorial University of Newfoundland, St. John's, Newfoundland, p. 432.

Easton, R.M. 1983. Crustal Structure of Rifted Continental Margins: Geological Constraints from the Proterozoic Rocks of the Canadian Shield; Tectonophysics, v.94, pp. 371-390.

Eugster, H.P. 1970. Thermal and Ionic Equilibria Among Muscovite, K Feldspar, and Aluminosilicate Assemblages; Fortschr. Mineralogie, v. 47, pp. 106-123.

Evans, B.W. and Guidotti, C.V. 1966.
The Sillimanite-Potash Feldspar Isograd in Western Maine, U.S.A.; Contributions to Mineralogy and Petrology, v.12, pp. 25-62.

Fraser, J.A., Hoffman, P.F. and Irvine, T.N. 1982.
The Bear Province; Special Paper, Geological Association of Canada, v.11, pp. 445-503.

Gray, D.R. 1979a.
Geometry of Crenulation-Folds and Their Relationship to Crenulation Cleavage; Journal of Structural Geology, v.1, pp. 187-205.

Gray, D.R. 1979b.
Microstructure of Crenulation Cleavage: An Indicator of Cleavage Origin; American Journal of Science, v.279, pp. 97-128.

Guidotti, C.V. 1963.
Metamorphism of the Pelitic Schists in the Bryant Pond Quadrange in Maine; American Mineralogist, v.48, pp. 772-791.

Hildebrand, R.S. 1981.
Early Proterozoic Labine Croup of Wopmay Orogen:Remnant of a Continental Volcanic Arc Developed During Oblique Convergence; Geological Survey of Canada Paper 81-10, pp. 133-156.

Hildebrand, R.S. 1983.
Geology of the Camsell River-Conjouror Bay Area, Early Proterozoic Cauldrons, Stratovolcanoes and Subvolcanic Plutons; Geological Survey of Canada Paper 83-20, pp. 1-42.

Hildebrand, R.S. and Bowring, S.A. 1984,
Continental Intra-Arc Depressions:A Non Extensional Model for their Origin, with a Proterozoic Example from Wopmay Orogen; Geology, v.12, pp. 73-77.

Hildebrand, R.S., Bowring, S.A., Steer, M.E. and Van Schmus, W.R. 1983.
Geology of the U-Pb Geochronolgy of Parts of the Leith Peninsula and Riviere Grandin Map Areas, District of MacKenzie; in Current Research, Part A, Geological Survey of Canada, Paper 83-1A, pp. 329-342.

Hildebrand, R.S., Hoffman, P.F. and Bowring, S.A. in press.

Tectono-Magmatic Evolution of the 1.9 Ga Great Bear Magmatic Zone, Wopmay Orogen, Northwestern Canada; Journal Volcanol. Geotherm. Res..

Hildebrand, R.S., Hoffman, P.F. and Bowring, S.A. 1986a.
Wopmay Orogen as an Evolving Continental Arc:Magmatic Response to Contrasting Stress Regimes; Geological Association of Canada Program Abstracts, v.11, p. 81.

Hildebrand, R.S. and Roots, C.F. 1985.
Geology of the Riviere Grandin Map Area (Hottah Terrane and Western Great Bear Magmatic Zone) District of MacKenzie; Geological Survey of Canada, Paper 85-1A, pp. 373-383.

Hobbs, B.E., Means, W.D. and Williams, P.F. 1976.
An Outline of Structural Geology. New York: Wiley and Sons. P. 555.

Hoffman, P.F. 1980.
Wopmay Orogen:A Wilson Cycle of Early Proterozoic Age in the Northwest of the Canadian Shield; in Strangway, D.W., ed., The Continental Crust and its Mineral Deposits:Geological Association of Canada, Special Paper, v.20, pp. 523-549.

Hoffman, P.F. 1981.
Revision of Stratigraphic Nomenclature, Foreland Thrust-Fold Belt of Wopmay Orogen, District of MacKenzie; in Current Research, Part A, Geological Survey of Canada, Paper 81-1A, pp. 247-250.

Hoffman, P.F. 1984.
Wopmay Fault Zone and its Role in Evolution of Wopmay Orogen; Geological Association of Canada Program Abstract, v.9, p. 74.

Hoffman, P.F. 1985.
Is the Cape Smith Belt (Northern Quebec) a Klippe?; Canadian Journal of Earth Sciences, v.22, p. 1361-1369.

Hoffman, P.F., King, J.E., Lalonde, S.B., St. Onge, M.R., Tirrul, R. 1986.
Axial Projections and Modes of Crustal Thickening, Eastern Wopmay Orogen, Northwest Canadian Shield; Geological Survey of Canada, in press.

Hoffman, P.F. and Bowring, S.A. 1982.

Short Lived 1.9 Ga Continental Margin and Its Destruction, Wopmay Orogen, Northwest Canada; Geology, v.12, pp. 68-72.

Hoffman, P.F. and McGlynn, J.C. 1977.
Great Bear Batholith:A Volcano-Plutonic Depression; Special Paper, Geological Association Of Canada, v.16, pp. 170-192.

Hoffman, P.F. and St. Onge, M.R. 1981.
Contemporaneous Thrusting and Conjugate Transcurrent Faulting During the Second Collision in the Wopmay Orogen: Implications for the Subsurface Structure of Post-Orogenic Outliners; in Current Research, Part A. Geological Survey of Canada, Paper 81-1A, pp. 251-257.

Hoffman, P.F., Tirrul, R. and Grotzinger, J.P. 1983.
The Externides of Wopmay Orogen, District of MacKenzie; in Current Research, Part A. Geological Survey of Canada, Paper 83-1A, pp. 383-395.

Jeletzky, O.L. 1974.
Unroofing of the Hepburn Batholith:A Petrographic Study of the Aphebian Recluse Formation, Epworth Group, Northwest Territories; unpublished B.Sc. Thesis Carlton University, Ottawa, p. 73.

King, J.E. 1985.
Structure of the Metamorphic Internal Zone, Northern Wopmay Orogen, Northwest Territories, Canada; unpublished Ph.D. Thesis, Queen's University, Kingston, Ontario, p. 208.

King, J.E. 1986
The Metamorphic Internal Zone of Wopmay Orogen (Early Proterozoic), Canada: 30 km of Structural Relief in a Composite Section Based on Plunge Projection; Tectonics, v.5, pp. 973-994.

King, J.E., Barrette, P.D. and Relf, C.D. 1987.
Contrasting Styles of Basement Deformation and Longitudinal Extension in the Metamorphic-Internal Zone of Wopmay Orogen, N.W.T.; in Current Research, Part A. Geological Survey of Canada, Paper 87-1A, pp. 515-531.

Lalonde, A.E. 1984.
The Characterization and Tectonic Significance of Plutonic Suites Within Central Wopmay Orogen, N.W.T.; Geological association of Canada Program Abstract, v.9, p. 80.

Losert, J. 1968.

On the Genesis of Nodular Sillimanite Rocks; International Geology Congress, 23 Session, v.4, pp. 102-122.

Macklin, J.H. 1950.
The Down-Structure Method of Viewing Geological Maps; Journal of Geology, v.58, pp. 55-72.

McGlynn, J.C. 1970.
Bear Province; in Geology and Economic Minerals of Canada, ed. R.J.W. Douglas; Geological Survey of Canada, Economic Geology Report 1, pp. 77-84.

McGlynn, J.C. 1976.
Geology of the Calder River (86 F) and Leith Peninsula (86 E) Map Areas District of MacKenzie; Geological Survey of Canada, Paper 76-1A, pp. 359-361.

Price, R.A., Monger, J.W.H and Muller, J.E. 1981
Cordilleran Cross Section-Calgary to Alberta; in Calgary '81 Annual Meeting, Geological Association of Canada, pp. 261-334.

Ramsay, J.G. 1967.
Folding and Fracturing of Rocks. New York: McGraw Hill, 568p..

Ramsay, J.G., Huber, M.I. 1983.
The Techniques of Modern Structural Geology Volume 1: Strain Analysis. New York: Academic Press, 301p..

Scott D.J. 1986.
Metamorphism, Structure, and Tectonics of the Eastern Cape Smith Belt, Northern Quebec; unpublished B. Sc. Thesis, McMaster University, Hamilton Ontario.

St. Onge, M.R. 1981.
"Normal" and "Inverted" Metamorphic Isograds and Their Relation to Syntectonic Proterozoic Batholiths in the Wopmay Orogen, Northwest Territories, Canada; Tectonophysics, v. 76, pp.295-316.

St. Onge, M.R. 1984a.
Geothermometry and Geobarometry in Pelitic Rocks of North-Central Wopmay Orogen (Early Proterozoic), Northwest Territories, Canada; Geological Society of America Bulletin, v.95, pp. 196-208.

St. Onge, M.R. 1984b.
Evolution of the Early Proterozoic Calderian Thermal Culmination:Mobility in Pressure, Temperature and Physical Space, Wopmay Orogen, N.W.T.; in Program with Abstracts, v.9 Geological Association of Canada, p. 108.

St. Onge, M.R. 1984c.
The Muscovite-Melt Bathograd and Low P Isograd Suites in North-Central Wopmay Orogen, Northwest Territories, Canada; Journal of Metamorphic Geology, v.2, pp.215-326.

St. Onge, M.R. and King, J.E. in press.
Evolution of Regional Metamorphism During Bac-Arc Stretching and Crustal Shortening in the 1.9 Ga Wopmay Orogen, Canada.

St. Onge, M.R., King, J.E. and Lalonde, A.E. 1984.
Deformation and Metamorphism of the Coronation Supergroup and its Basement in the Hepburn Metamorphic-Plutonic Zone of Wopmay Orogen: Redrock Lake and the Eastern Portion of Calder River Map Areas, District of MacKenzie; in Current Research, Part A. Geological Survey of Canada, Paper 81-1A, pp. 171-180.

St. Onge, M.R., Lalonde, A.E. and King, J.E. 1982.
Geology of the Central Wopmay and the Eastern Portion of the Calder River Map Area; in Current Research, Part A. Geological Survey of Canada Paper 83-1A, pp. 147-152.

St. Onge, M.R., Lalonde, A.E. and King, J.E. 1983.
Geology, Redrock Lake and Eastern Calder River Map Areas, District of MacKenzie:The Central Wopmay Orogen (Early Proterozoic), Bear Province, and the Western Archean Slave Province; in Current Research, Part A. Geological Survey of Canada, Paper 83-1A, pp.147-152.

St. Onge, M.R., Lucas, S.B., Scott, D.J. and Begin, N.J. 1986.
Eastern Cape Smith Belt:An Early Proterozoic Thrust-Fold Belt and Basal Shear Zone Exposed in Oblique Section, Wakeham Bay and Cratere du Nouveau Quebec map areas, Northern Quebec; in Current Research, Part A. Geological Survey of Canada, Paper 86-1A, pp. 1-14

Tippett, C.R. 1980.
A Geological Cross-Section Through the Southern Margin of the Foxe Fold Belt, Baffin Island Arctic Canada, and its Relevance to the Tectonic Evolution of the Northeastern Churchill Province; unpublished Ph.D. Thesis, Queen's University.

Tirrul, R. 1982.
Frontal Thrust Zone of Wopmay Orogen Takijug Lake map area, District of MacKenzie; in Current Research, Part A.Geological Survey of Canada, Paper 82-1A, pp. 119-122.

Turner, F.J. 1968,
Metamorphic Petrology. New York: McGraw-Hill, 524p..

Winkler, H.G. 1979.
Petrogenesis of Metamorphic Rocks, Fifth ed.. New York: Springer Verlag, 348p..

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