STRUCTURAL STUDIES IN THE MORETOWN AND CRAM HILL UNITS NEAR LUDLOW, VERMONT

A thesis presented to the Faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of Master of Science

William J. Gregg
1975

ACKNOWLEDGMENTS

The writer is indebted to Dr. Paul F. Williams for introducing him to the field area and providing assistance in the early part of the study. Throughout the study Dr. Winthrop D. Means provided advice and encouragement and made many helpful suggestions pertaining to methods of field investigation. Bruce W. Nisbet provided the author with excellent companionship and expertise in the field area on many occasions. Dr. George W. Putman, Dr. William S.F. Kidd, and Dr. Winthrop D. Means are gratefully acknowledged for patiently reviewing this manuscript and offering suggestions which helped to improve its quality in many ways.

Finally, the author wishes to express his sincere thanks for the encouragement and knowledge passed on to him by Roger N. Miller and especially Winston A. Dezaine, whose familiarity with the geologic structure and mining history of the ultramafic rocks of Vermont is surpassed by none.

CONTENTS

Tables

Illustrations

Figure Number | Page

Illustrations

Illustrations

Figure Number Page

INTRODUCTION

The geology of the eastern limb of the Green Mountain Anticlinorium consists of a series of Paleozoic metasedimentary rocks with lithologic boundaries arranged in a remarkably straight trend approximately parallel to the axis of the Green Mountains. Published reports of the area, consisting largely of reconnaissance mapping, have treated this complex series of polyphase deformed rocks as an essentially upright autocthonous sedimentary sequence. Boundaries between rock units have, for the most part, been assumed to be primary in origin, as have various structural elements within the rock units. More recent work in selected areas within the Ludlow Quadrangle (Fig. 1) has revealed problems in the application of stratigraphic techniques toward field mapping in these medium grade metamorphic rocks.

One of the main problems is the consistent misidentification of demonstrably secondary mesostructures as primary sedimentary features by early workers. Many rocks in the area have been shown to contain secondary structures which mimic such primary features as bedding and conglomeratic pebbles (Gregg and Nisbet, in preparation) and which have in fact been mistaken for these structures in the past. In some cases so-called conglomerate horizons, recently shown to be pseudo-conglomerates produced by transposition of early layering (Gregg and Nisbet, in preparation), have been used to mark basal units of various formations. The layering in the surrounding rock has been thus assumed parallel to the "conglomerate" on a large scale. It is more likely that the "conglomerates" occupy zones of

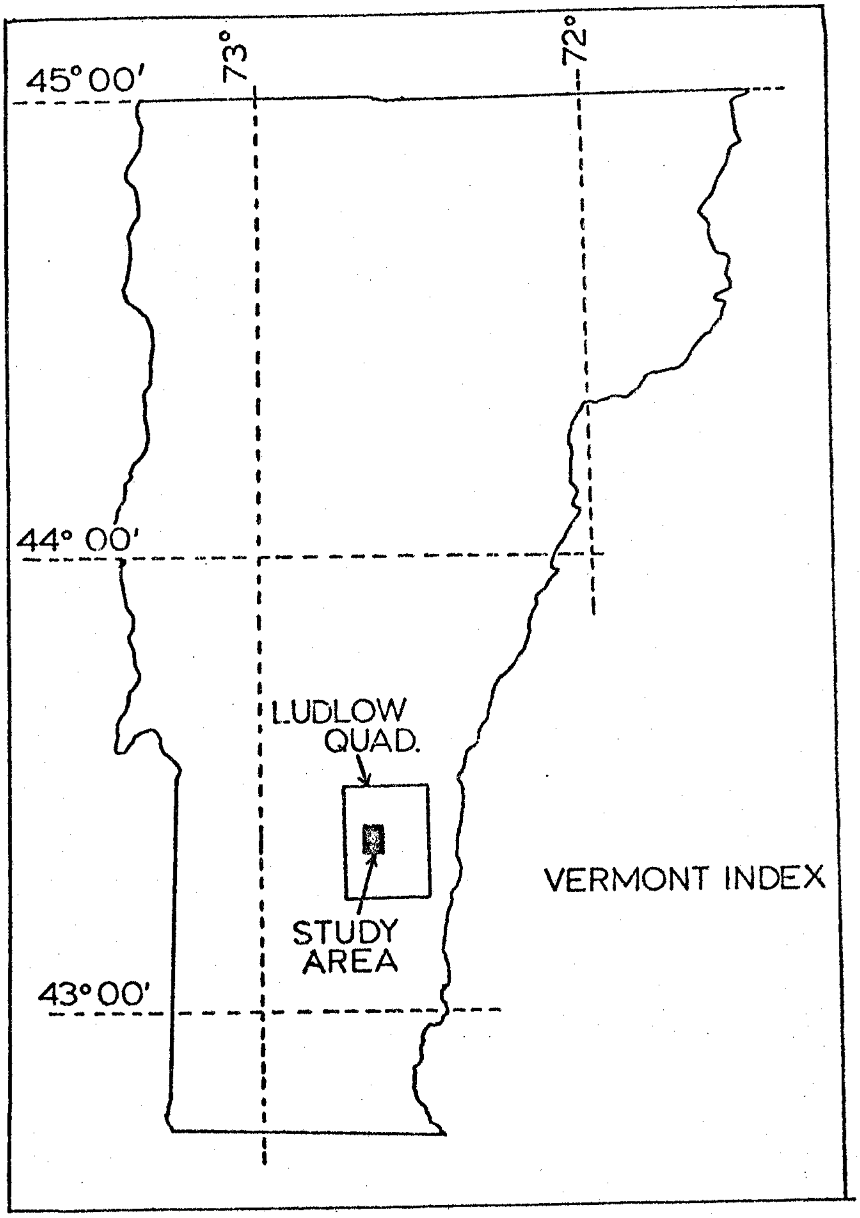


Fig. 1 Index map of Vermont showing study area.

high strain and the transposition mesostructures may in fact be due to the proximity of the rocks to a series of faults between adjacent rock formations. Thus the transposed rocks probably had initial compositions similar to the surrounding rocks, but have been altered to "pseudo-conglomerates" near the fault controlled rock boundary.

In addition to problems arising from incorrect interpretation of mesostructures, other problems have resulted from the failure of structural geologists to distinguish various fold groups on the basis of overprinting relationships. For example, in areas where early and late folds are known to exist, even by earlier workers, one often finds that all folds are grouped together, plotted on the same diagrams, or represented on maps without indicating which folds overprint other groups. Work such as this has led to the formation of rather novel, though unsubstantiated, ideas concerning the interpretations of regional structure; for example, the so-called "spruce tree" folds on the limbs of large mantled gneiss domes (Thompson, 1950; Skehan, 1961) which actually represent misinterpreted early folds, refolded about the domes (B. W. Nisbet, unpublished Ph.D. thesis).

This investigation deals with a subarea within the Ludlow quadrangle where detailed structural mapping was performed by the author from 1971 to 1974 largely during the summer months. The area is about 5 km long and 2 km wide and occupies the central portion of Figure 2. Mapping was performed on scales of 30 feet per inch in most areas and 100 feet per inch in a few outlying areas. In most cases the geologic contacts and outcrops close to the central ultramafic zone were surveyed by steel tape and brunton from permanent monuments established by transit survey.

The central feature of the area is the ultramafic zone consisting dominantly of serpentinized ultramafic rock masses up to 1 km long with minor zones of talc-carbonate rocks around the boundaries. The core rocks within the serpentinite bodies contain up to 50% early minerals, dominantly orthopyroxenes, although clinopyroxenes may be as high as 5%. Olivine is rarely present in the thin sections examined, probably because of the degree of serpentinization. Early mapping by Richardson (1928) delineated the serpentinite" as a single enormous body of peridotite and pyroxenite over 6 km long and 2 to 3 km wide. Subsequent mapping by Thompson (1950) reduced the mass to three large bodies from 1 to 2 km long (Fig. 2). Mapping by the author has further defined the small scale structures in the ultramafic zone and reduced the masses to at least 6 smaller bodies, the largest of which is about 1 km long and 0.5 km wide. A number of smaller bodies of serpentinite around the larger cores have also been mapped. These are usually on the order of 50 to 70 meters long and 30 to 40 meters wide.

The ultramafic zone is situated along the boundary of two rock units, the Moretown member and the Cram Hill member of the Missisquoi Formation. Although these units have been called members of the same formation by early workers, the author has found a number of striking contrasts in the deformational features of the units. For example, the Cram Hill phyllites have been involved in only two phases of deformation, while the Moretown gneisses contain structures from at least one earlier deformational phase. The Cram Hill phyllites contain layering which may be of sedimentary origin and which is moderately deformed in most cases. The Moretown member has been severly deformed, and all trace of initial layering is obliterated. Many other contrasts between these two "members" will be discussed in the text and evidence for a tectonic contact between the two rock types will be presented.

The author considers that there is little basis for classifying the various rock types by the "Formational" and "Member" designations, and that future structural work in the Central Vermont area will result in the abandonment of the stratigraphic nomenclature now being applied. For these reasons the author will not use the terms applied by Thompson (Fig. 2) and will return to rock names similar to those used by the earliest workers (such as Richardson, 1928). The Moretown "member" and the Cram Hill "member" will be referred to as the Moretown gneiss and the Cram Hill phyllite respectively; the rock names reflecting the dominant rock type observed in each unit.

During field work and thin section examination the author observed a number of tabular garnets in the rocks of the Moretown member. In most cases the garnets were formed during the earliest deformational event and were deformed into tabular shape by later deformation. The deformation, however, was not the typical flattening assumed by most workers, but a slicing process in which segments of garnets from an initially equidimensional crystal are sheared parallel to rock layering. A number of examples of partially sliced crystal sections were observed on mesoscopic and microscopic scales.

Field mapping was performed according to the methods outlined by Turner and Weiss (1963), and further developed by later workers (for example, Means, 1963; Williams, 1967; and many others).

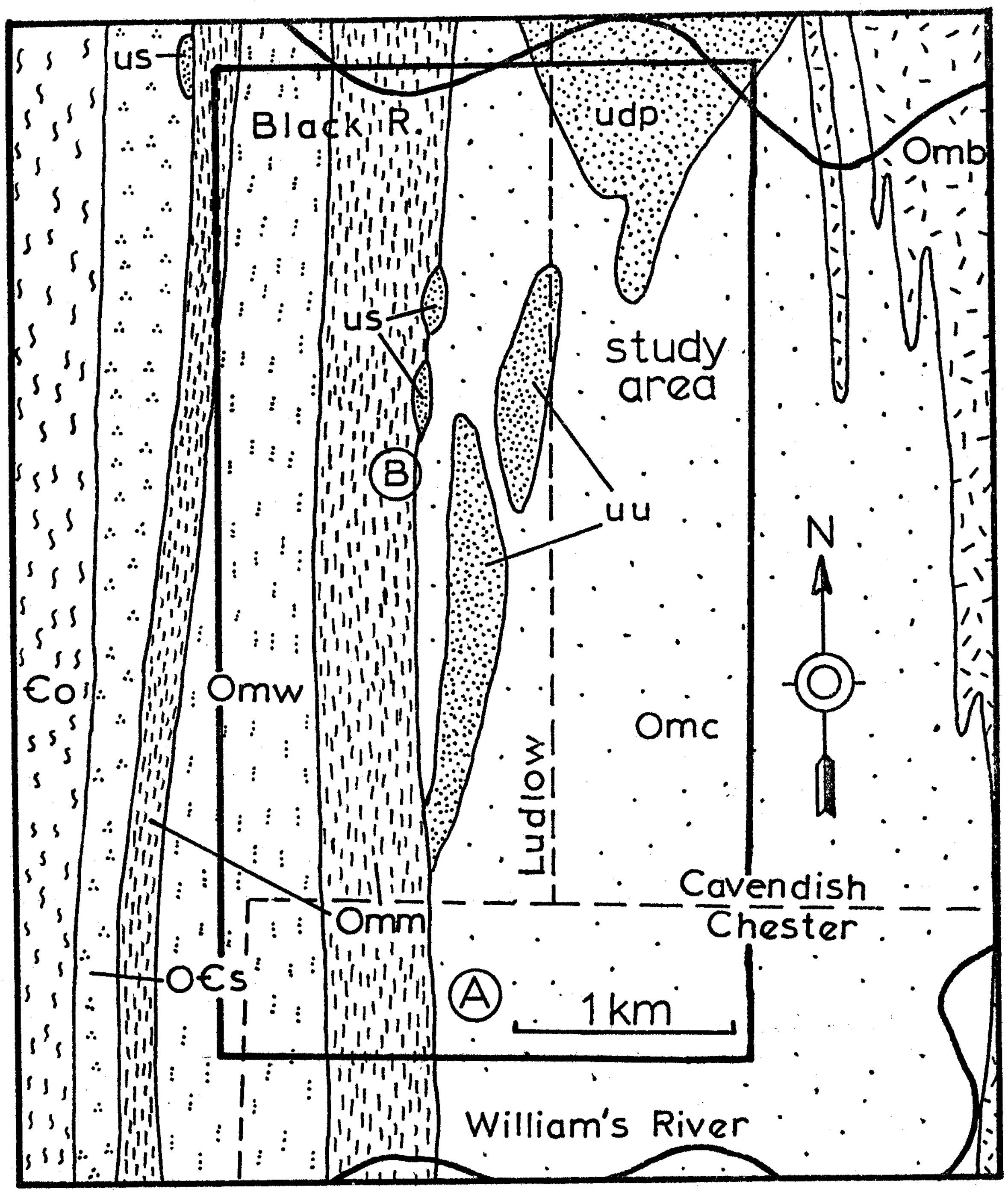


Figure 2 - Geologic setting of study area showing geology as mapped by Thompson (1950). Area "A" is referenced in text as the vicinity of the Cram Hill quartzite horizon; area "B" is subarea B in text.

Missisquoi Formation:

- Omc = Cram Hill member
- Omb = Barnard Volcanic member
- Omm = Moretown member
- Omw = Whetstone Hill member

- OCs = Stowe Formation
- Co = Ottauquechee Formation
- uu = undifferentiated ultramafics
- udp = dunite, peridotite, serpentinite
- us = serpentinite, carbonate rock, talc-carbonate rock & steatite

CHAPTER I

GEOLOGY OF THE MORETOWN GNEISS

Introduction

Rocks of the Moretown unit form the western boundary of the central ultramafic zone (Fig. 2). The Moretown consists for the most part of thinly banded gneisses containing infrequent layers of chlorite-garnet schists usually no greater than 20 cm thick (Fig. 3). The most prominent foliation is a north-south trending regional "pinstriping" consisting of alternating quartzo-feldspathic and mica-rich domains. The rocks are of the epidote-amphibolite facies with abundant blue-green amphibole and plagioclase compositions below AN_{30} in the meta-basites found near the ultramafic zone. Although garnet is ubiquitous in the gneissic rocks, only a single occurrence of garnet amphibolite was noted. Most of the amphibolites mapped are discontinuous lenses usually 1m to 5m thick; these will be discussed in another section.

Mineralogy of the Moretown Rocks

Table I lists the estimated modes for gneissic and schistose varieties of Moretown rocks. Thin sections were made from hand specimens, and drill core taken from a single 135 meter diamond drill hole at 3 meter intervals through 30 meters of the Moretown in the vicinity of subarea "B" (Fig. 2). The averages for the Moretown are grouped according to gneissic and schistose categories. Typical gneissic rocks contain 69% quartz + albite, 14% muscovite, 9% chlorite, 5% biotite and minor proportions of magnetite, carbonate, pyrite, epidote, and garnet. Typical chlorite schists interbedded infrequently with the gneisses contain 29% quartz + albite, 39% muscovite,

Figure 3 – Typical outcrop appearance of minor chlorite-garnet schist layers interbedded with more massive units of gneissic rocks in the Moretown. The fold is a large Group M-3 fold of S_2 layering. S_3 is weakly developed in the schist. Schist band is approximately 5 cm thick.

TABLE I

ESTIMATED MODES FOR MORETOWN ROCKS
(in percent mineral const.)

SAMPLE *	N	QUARTZ +ALBITE	MUSCOVITE	MAGNETITE	CARBONATE	BIOTITE	CHLORITE	PYRITE	EPIDOTE	GARNET
DDH 18-R-73:										
G 327'	200	61	9	1	0	0	25	0	4	0
G 337a'**	200	73	19	tr.	0	4	3	0	tr.	0
G 337b'**	200	67	20	2	0	0	8	0	0	3
G 347'	200	62	19	1	0	0	18	0	0	0
G 357a'#	200	63	18	1	2	0	12	0	0	5
S 367'	200	34	39	2	1	1	15	0	0	9
S 387'	200	23	49	2	0	10	14	1	0	0
G 397'	200	78	5	0	0	5	11	1	0	0
G 407'	200	76	2	2	0	18	2	0	0	0
S 414'	200	42	30	3	1	8	15	1	0	0
S 427'	200	17	50	2	0	1	13	0	0	17
S 436'	200	28	26	0	0	20	17	1	0	8
G 357b'#	1000	56	25	1	1	13	3	0	0	2
G, U-13-74b	200	68	20	2	0	0	8	0	1	2
G, C-5-73	203	73	12	0	0	13	1	0	0	tr.
G, U-14-74	200	79	9	2	0	4	4	0	2	tr.
5 Schists---Averages		29	39	2	tr.	8	15	1	0	7
11 Gneisses Averages		69	14	1	tr.	5	9	tr.	1	1

* Prefix G denotes gneissic sample, S denotes schistose rock.
** G 337a and G 337b are two different thin sections from same footage.
\# G 357a and G 357b are from the same thin section. G 357a was taken in the quartzo-feldspathic layering in the section while G 357b was taken over the entire section, and includes a mica rich domain outside field of G 357a.

15% chlorite, 8% biotite, 7% garnet and minor proportions of magnetite, carbonate, pyrite, and epidote. Albite is listed with quartz because it is not usually possible to distinguish fine-grained, untwinned albite from quartz during thin section modal counts, although selective staining of polished rock slabs usually demonstrated the presence of albite. X-ray diffraction scanning produced albite peaks of low amplitude compared to those of quartz and sericite in the same sample. Powder samples were prepared from rock specimens U-14-73 and 18-R-73 (337') which are listed in Table I.

Selective staining of polished rock slabs indicated the absence of orthoclase in the Moretown gneisses and schists, a fact confirmed by x-ray diffraction scanning. Albite is dispersed irregularly throughout the gneisses on various scales. Certain specimens may contain up to 30% albite in the whole rock, with the mineral concentrated in layers between mica-rich domains. More commonly, albite is distributed uniformly in single hand specimens (for example, Fig. 9) as fine grains comprising 1% or less of the total mineralogy. The extreme variations in distribution make it overly speculative to suggest a bulk content of albite for the Moretown rocks as a whole; however, in most of the samples observed, albite was less than 5%. It may be that higher albite contents occur in the schists listed in Table I but only a few specimens of schist were stained for feldspars. Thus, the percentages given for quartz + albite in the gneissic rocks probably contain up to 5% of the amount as albite, with the remainder as quartz. Likewise, the schistose rocks in Table I may contain up to 30% of the quartz + albite portion as albite, with the remainder as quartz. In comparing the compositions of the schists and gneisses as components of the Moretown, one must bear in mind that the schists appear to be volumetrically minor relative to the gneissic rocks.

Structual Studies in the Moretown Gneiss

(A) Foliations in the Moretown

The most prominent layering in the gneisses is the "pinstripe" layering already mentioned in the introduction to this chapter. The pinstripe layering (S_2) consists of alternating quartzo-feldspathic and layer-silicate domains. The layer silicate domains are usually less than 0.5 mm thick and the quartzo-feldspathic domains vary from 1 to 5 mm in thickness (Fig. 4).

Within quartzo-feldspathic domains of S_2 an earlier layering (S_1) identical in mineralogy and appearance to S_2 is recognizable (Fig. 5). The earlier S_1 layering is usually disposed in tight to isoclinal folds, with the S_2 pinstriping subparallel to the axial planes of these folds. Although usually difficult to recognize in the field, the early layering may form a fine pseudo-crossbedding within the S_2 layering, but when the specimen is cut parallel to the observed face, the secondary[1] layering and crenulated S_1 layering within the microlithons between S_2 mica-rich domains are always distinguishable.

Various authors (Thompson, 1950, p. 42; Ern, 1963, p. 31; Chang *et al.*, 1965, p. 31) have referred to the pinstriping as sedimentary layering. Both Cady (1961) and Skehan and Hepburn (1972, p. 8) consider that locally the

---

[1]"secondary layering" refers to any demonstrably non-primary layering observed in deformed rocks, and may include "differentiated, metamorphic, tectonic or deformational" layering. "Primary layering" is usually considered to be an initial layering of sedimentary or igneous origin. The recognition of an earlier layering, cut by a later layering, is usually considered as proof that the latter is secondary.

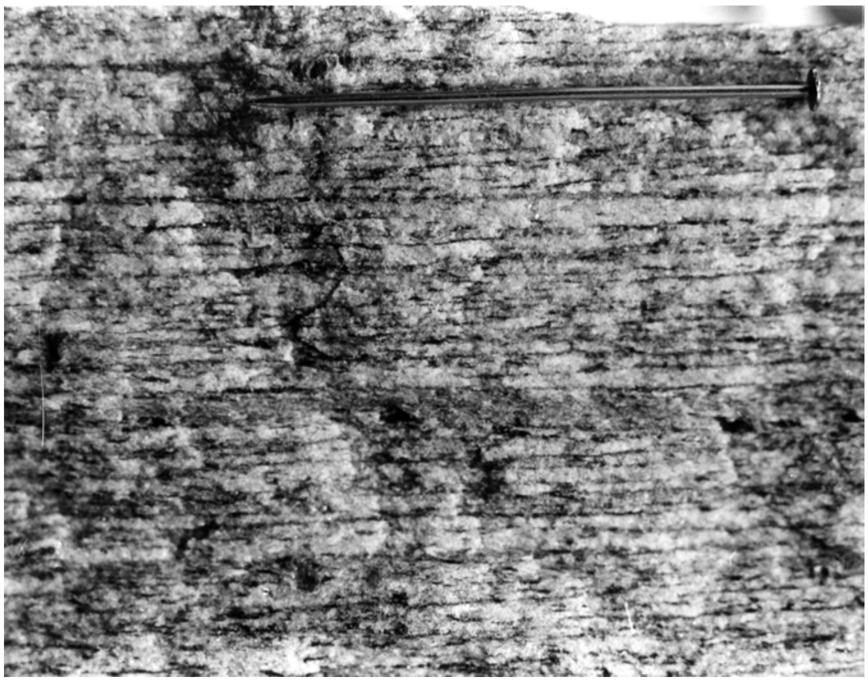


Figure 4 – Typical outcrop appearance of S_2 “pinstripe” layering in the Moretown gneisses. This layering has been referred to as sedimentary by previous workers but is in fact secondary in origin. Pin is 25 mm long.

Figure 5 – Moretown gneiss specimen showing relationship of S_2 pinstripes to earlier S_1 layering. S_2 is vertical in photograph and axial plane within folds in early S_1 layering which crosses photograph diagonally from right to left. Area above penny is enlarged in Figure 6. Diameter of penny is 19 mm.

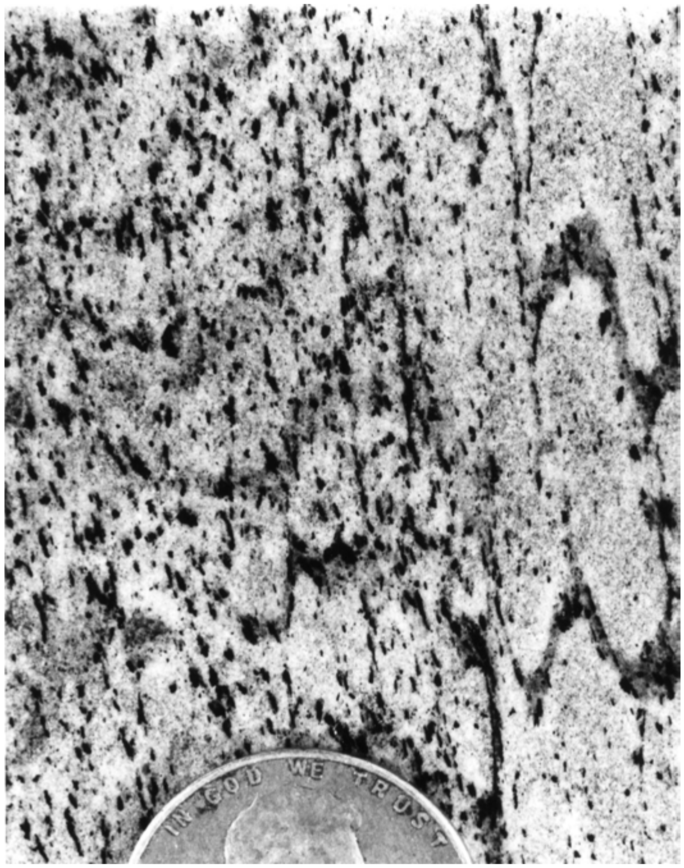


Figure 6 – Enlargement of area in Figure 5 just above penny showing early layering folded by Style Group M-2 folds with axial plane secondary layering S_2 oriented vertically in photograph

pinstriping may be secondary, but Cady goes on to say that on a larger scale it is a bedding foliation because of its parallelism with quartzite beds (Cady, 1961, p. 5).

The secondary nature of the pinstriping can be demonstrated at many localities in the Ludlow area, but the early S_1 layering can only be recognized where secondary micaceous domains in S_2 are relatively widely spaced. Where this is so, the microlithons between the micaceous pinstripe domains become large enough to trace individual S_1 layers across them. As the spacing between S_2 layer-silicate domains decreases, recognition of early layering becomes increasingly difficult, and when the S_2 micaceous domains are less than 3 mm apart (as is usually the case), field recognition of S_1 layering is impossible. Even in this case, however, the intersection lineation between S_1 and S_2 can be observed on S_2 surfaces, attesting to the existence of S_1 in the rock. S_1 layering appears to be accentuated by the formation of secondary quartz layering parallel to S_1 possibly associated with layer-parallel straining in the rock. This evidence will be discussed in a later section on the textural relationships between garnet microstructures and layering.

Although no earlier layering than S_1 has been directly observed, it appears likely that S_1 is also secondary in origin since it is identical in morphology to S_2. On the point of similarity between S_1 and S_2 an interesting observation bears mention. Rare specimens collected show the usual S_2 layering closely spaced and all but obliterating S_1; but an additional layering (S_3) has formed across S_2. This new layering is again nearly identical to both S_1 and S_2 (Fig. 7). Apparently the Moretown gneisses responded in a relatively consistent manner to each successive deformation, forming nearly identical types of layering with respect to each episode. It seems likely that all sedimentary mesoscopic structures have been destroyed during the polyphase deformational history of the Moretown gneisses.

(B) Macroscopic Structures in the Moretown

On a macroscopic scale, the Moretown consists of an essentially featureless unit, due in part to the poor outcrop, but primarily due to the total absence of any lithological "marker" horizons. Throughout the entire area the only other mappable rock types in the Moretown consist of the layered amphibolites infrequently encountered near the borders of the ultramafic zone. Mapping of these units has shown that they have a high degree of parallelism with the ultramafic zone and a corresponding discordancy with mesoscopic early layering in the Moretown gneisses. The fact that all the amphibolite bodies lie sub-parallel to S_2 in the country rock implies emplacement for these units late in the development of S_2. There is no evidence that transposition is responsible for the arrangement of the amphibolite bodies subparallel to S_2.

(C) Style Groups in the Moretown

Within the study area two style groups have been identified in the Moretown. The elements of each style group are the fold style, the type of axial plane secondary layering if developed, and the relationship of new layering and metamorphic minerals. The Style Groups are discussed in order from the oldest to the youngest groups based on age relationships established by overprinting of elements among different groups. Table II is a summary of the structural elements in each Style Group and should be referred to throughout the text.

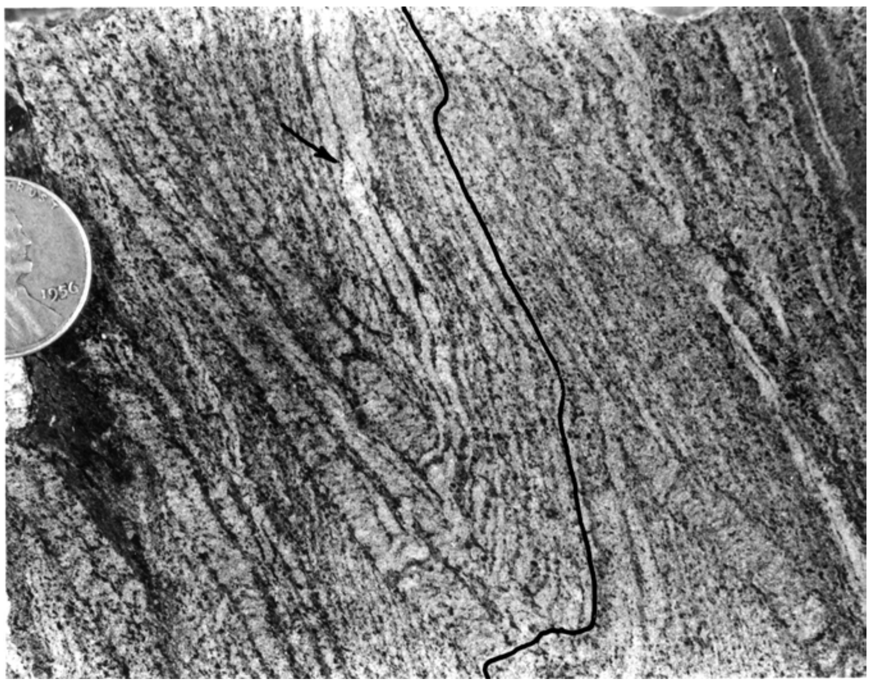


Figure 7 – Specimen U-31-74 showing three generations of layering in the Moretown gneisses. S_3 cuts across photo at about 45° from lower center to upper left. S_2 trends more vertically than S_3 and is marked by the ink line. Note that S_2 is folded by Group 3 folds with S_3 as axial plane foliation. Arrow shows early S_1 layering lying within quartzo-feldspathic domains of S_2 layering.

1. Early layering in the Moretown -- S_1 layering has already been discussed with respect to its close resemblance to S_2; and the probable secondary origin of S_1 has been commented upon. S_1 layering usually consists of alternating quartzo-feldspathic layers and mica-rich layers. The quartzo-feldspathic layers range from 1 to 3 mm thick and the mica-rich layers are usually less than 1 mm thick. Garnets are contained within the S_1 layering, often as equant, undeformed porphyroblasts. In all cases observed thus far, S_1 layering is the only layering in which garnet growth occurred. Layer-parallel shearing in much of the S_1 layering produced tabular garnet forms parallel to S_1. The garnet microstructures of the Moretown are described in more detail in a later section.

The nature of the fine scale S_1 layering is such that, like S_2 it tends to obliterate early mesoscopic structures. Because of this it is difficult to adequately demonstrate an earlier foliation (S_0) and, consequently, the folds associated with S_1 development are difficult to observe. For this reason it is not possible to classify a Style Group for S_1 related structures. There are occasional examples of folds believed to be associated with S_1 in the area, but the relationships between these folds and S_1 is not clear.

2. Elements of Style Group 2 -- The structural elements which deform S_1 layering and its associated microfabrics are collectively referred to as Style Group M-2. They are referred to as Style Group 2 because, although it is not possible to adequately define a Style Group 1, the existence of S_1 indicates that an early Style Group may yet be defined in an area where deformation due to Style Group M-2 elements is not as severe. Group M-2 folds are commonly observed on a mesoscopic scale, usually on the order of

TABLE II

STYLE GROUP ELEMENTS IN THE MORETOWN

STYLE GROUP	FOLD STYLES	SECONDARY LAYERING	LINEATIONS	PORPHYROBLASTIC METAMORPHIC MINERALS
M-3	Usually open, may have tight angles in hinge. Planar axial surfaces.	S_3- new layers are mica-rich.	L_3- weak crenulations.	none
M-2	Isoclinal to tight folds usually less than 10-20 mm in area.	S_2- new layers are mica-rich.	L_2- intersection between S_1 and S_2 on S_2 surfaces.	none
?	(not yet demonstrated)	S_1- new layers are mica-rich.	none observed	Garnet growth before development of S_1 is cut by S_1 surfaces.
-----	------	S_0-early layering?	----	-------

a few millimeters or centimeters in wave length. They usually consist of very tight to isoclinal folds of S_1 layering (Fig. 5). Large examples of Group M-2 folds are relatively rare, but where seen consist of tight folds of S_1 layering, in some cases with strongly attenuated fold hinges (Fig. 8). The rarity of large scale Group M-2 folds is due to the tendency toward obliteration of early structures by the secondary axial plane "pinstripe" layering (S_2) associated with these folds.

The intersection of S_1 and S_2 forms a strong lineation (L_2) throughout the Moretown gneisses. L_2, usually observed parallel to Group M-2 fold axes, is the only lineation seen in the Moretown rocks and is observed only on S_2 surfaces. L_2 at first appears to be a mineral lineation defined by oriented amphibole minerals, but on closer inspection it can be seen to consist of the edge-on view of biotite and muscovite crystals which lie in the plane of S_1, intersected by S_2.

3. <u>Elements of Style Group M-3</u> -- Group M-3 folds are infrequently observed and are usually open folds of S_2 layering. Small scale examples of Group M-3 folds are usually restricted to irregular crenulations which deform S_1 and S_2 and overprint folds of style group M-2. Group M-3 folds are most common within subarea "B" (Fig. 2) where the regional foliation (S_2) is disposed in open folds on the order of ten meters in wavelength. Subarea "B" is of importance in that no other area within the 5.6 square miles mapped showed such a scale of deformation of S_2 in the Moretown. S_2 layering outside of subarea "B" is deformed by Group M-3 crenulations which rarely exceed a few millimeters in wavelength, (refer to Fig. 7). Group M-3 folds within subarea "B" overprint Group M-2 folds of S_1 layering in outcrops where S_2 is poorly developed, as for example in Figure 8.

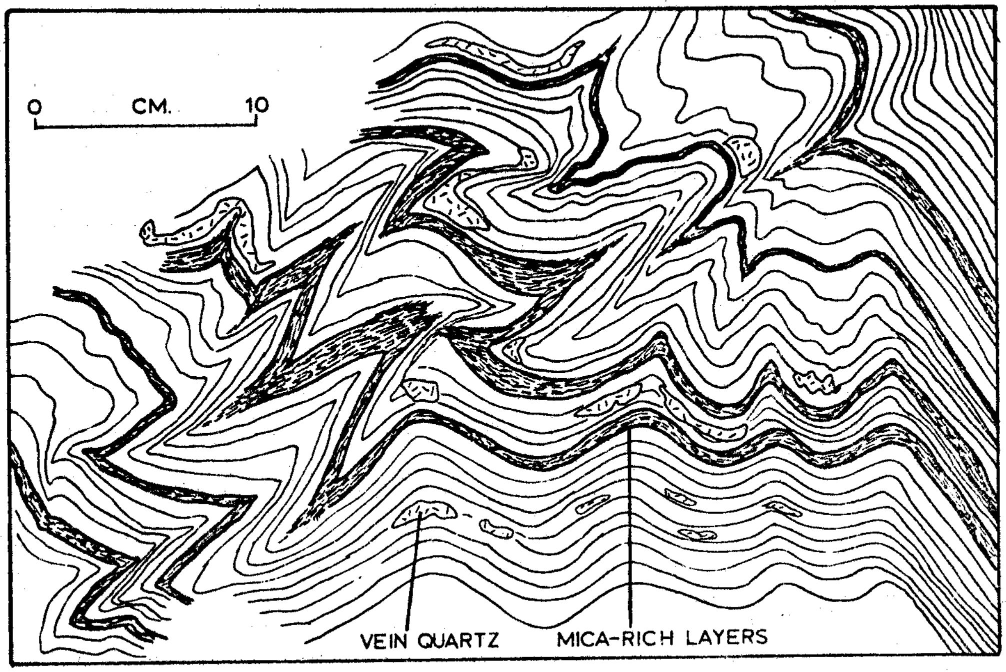


Figure 8 - Group M-2 folds of S_1 layering in Moretown gneiss refolded by Group M-3 folds with vertical axial surfaces. Photo from subarea "B" (Fig. 2).

Group M-3 crenulations throughout the area often form complex mesostructures when overprinting typical Moretown gneisses, especially when both S_1 and S_2 exist in the same rock on a coarse scale. For example, in Figure 9, group M-3 crenulations are oriented with axial plane traces running from lower left to upper right and are difficult to distinguish in much of the layering. The heavy mica-rich domains which are oriented approximately vertically in the photograph are secondary S_2 layer-silicate layers. Finally, the thinner, mica-rich layers oriented approximately 15 to 30^o counterclockwise from vertical in the photograph are early S_1 mica domains. These S_1 domains are cut by the vertically oriented S_2 domains at angles from 15^o to 30^o to S_2.

Figure 10 illustrates an example similar to Figure 9 in which S_2 and S_1 are more easily distinguished. Examples such as this are never seen in the field, due to weathering as well as rarity, and are usually only discovered in polished rock specimens.

The schematic illustrated (Fig. 11) represents a possible interpretation of the mesostructures in Figure 10. Foliations of this type are often mistakenly referred to as transposed layering because the intersections between S_1 and S_2 may mimic isoclinal folds (especially so in the case of Figure 9).

(D) Tabular Garnets in the Moretown

The various style groups observed in the Moretown are presented in Table II. With reference to the early layering S_1, garnets have been categorized as associated metamorphic minerals, although the justification for this has not yet been presented. The problems and implications associated with work on garnet microstructures in the Moretown are sufficiently

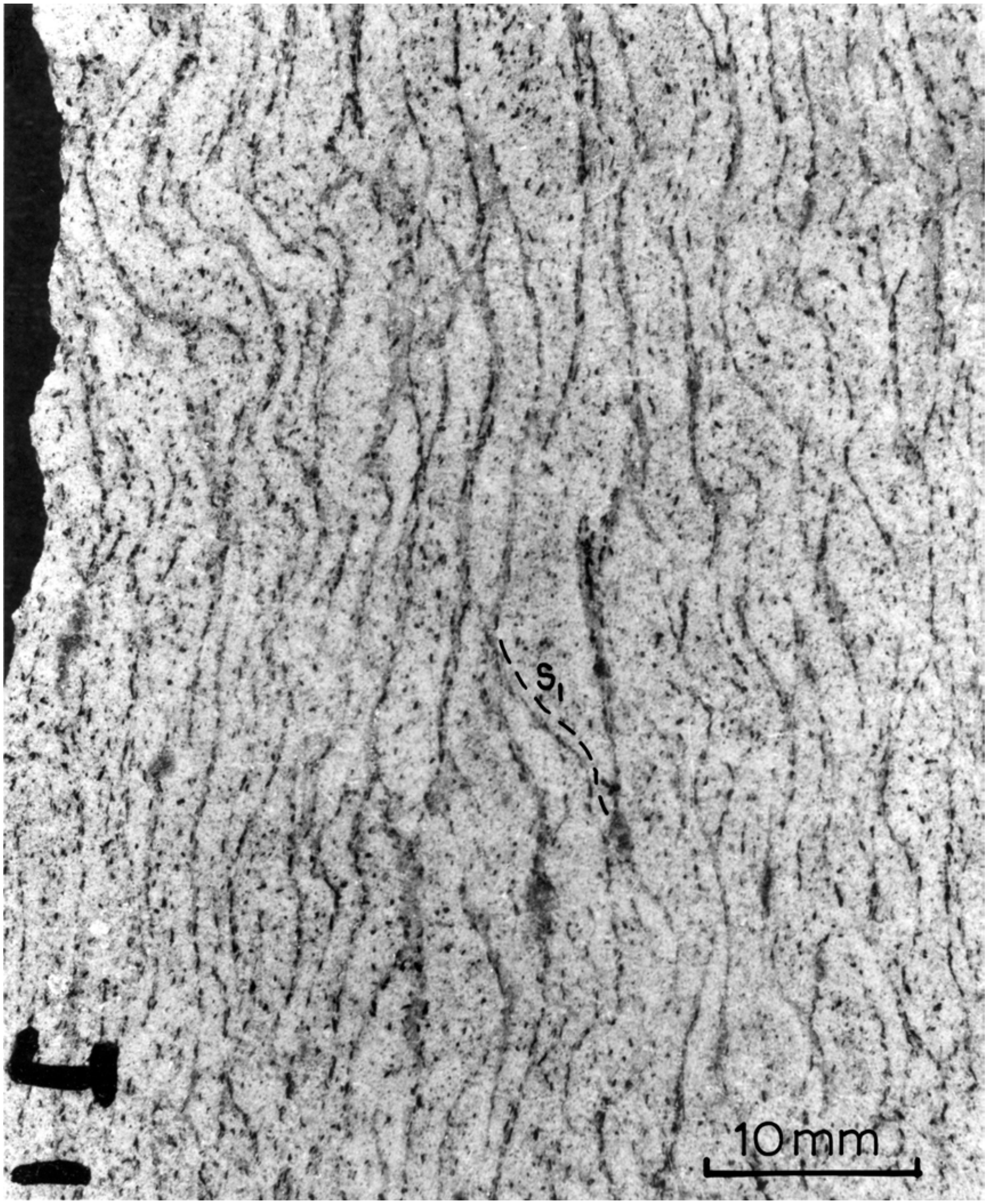


Figure 9 – Complex mesostructures produced by Group M-3 and Group M-2 folds of S_2 and S_1 layering. Group M-3 folds are crenulations with axial planes crossing photo 30º from horizontal from lower left to upper right. S_2 trends dominantly vertical and is folded by Group M-3 folds. S_1 trends 15º to 40º to left of vertical in photo; a single S_1 layer is outlined in center of photo.

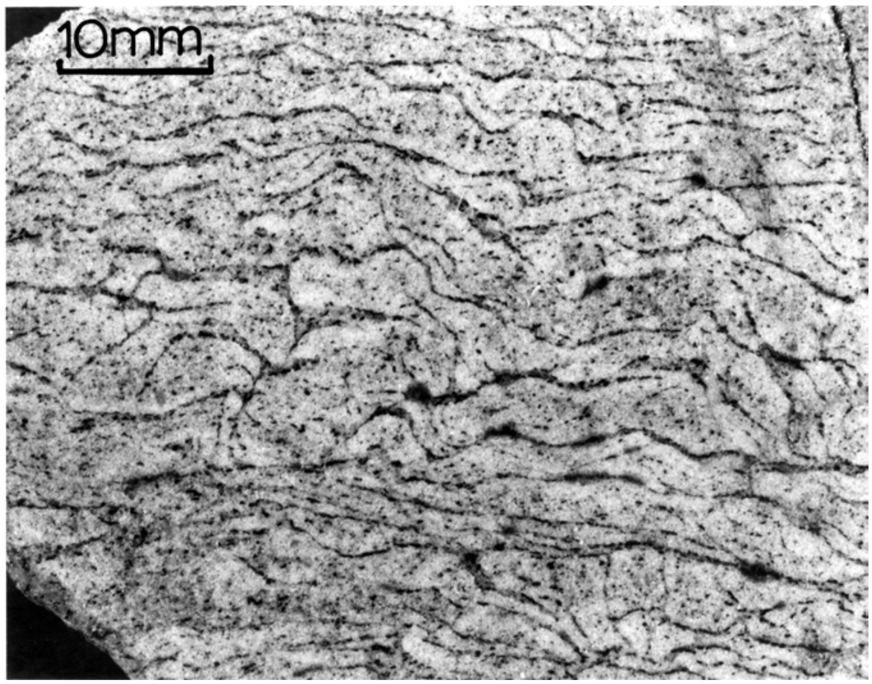


Figure 10 – Complex mesostructures similar to those in Figure 9 but with S_2 mica-rich domains more widely spaced and S_1 at higher angles to S_2 in some areas of the photograph. Group M-3 crenulations trend toward vertical axial planes; S_2 is dominantly horizontal in the photo, but folded by Group M-3 folds; S_1 usually is at various angles to vertical and is truncated by S_2 in various sections of the photo. Refer to Figure 11 for a schematic drawing.

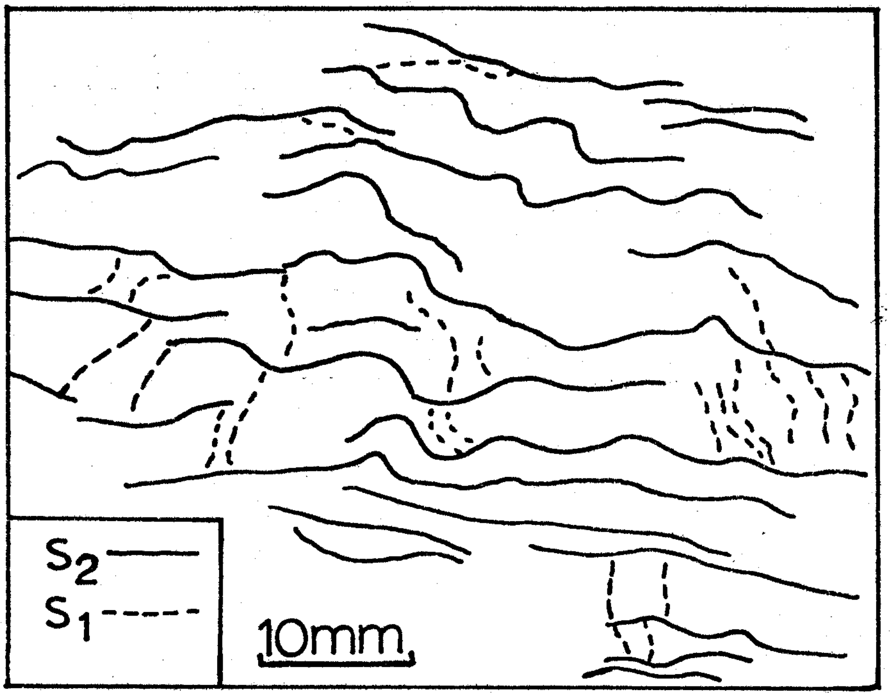


Figure 11 - Schematic drawing of Figure 10 to same scale showing possible interpretation of mesostructure in certain areas. Interpretation of all layering is not possible, especially in areas such as the upper quarter of Figure 10 where the S_2 domains are closely spaced and nearly obliterate S_1 traces.

complex as to require additional discussion.

Specimens of garnetiferous quartzo-feldspathic gneisses have been collected throughout the Moretown and have been found to contain garnets of unusual tabular shape, apparently due to the slicing of originally equant garnet crystals along a direction parallel to the rock foliation.

Most of the garnet crystals are sliced parallel to S_1 in the Moretown although one occasionally finds garnet slices parallel to S_2. Some examples of individual garnet crystals show slicing parallel to both of these secondary layerings. All garnet crystals observed in equidimensional form (i.e., before slicing) indicate early growth in S_1 layering. No evidence has been seen indicative of growth in S_2. Thus, S_2 layering always cuts through any garnets intersected, while this need not be the case for S_1.

Garnets are dispersed throughout the Moretown in grain sizes ranging from 5 mm to 20 mm with shapes varying from undeformed, relatively equidimensional prophyroblasts to tabular grains with aspect ratios ranging as high as 30 to 1. When observed normal to S_1 or S_2 foliations, surfaces, all garnets collected showed no elongation in any particular direction (Fig. 12).

The tabular garnets have been observed in progressive stages of slicing both within individual crystals and among different crystals within the same rock specimen. Figure 13 depicts a relatively equidimensional garnet in which prominent fractures can be observed in a subparallel orientation to the foliation. Small segments along the fractures are filled with quartz, usually in sections parallel to the rock layering. Along the top of the grain a quartz-rich layer is developed, possibly along a shear surface, as may be evidenced by small trains of garnet fragments extending

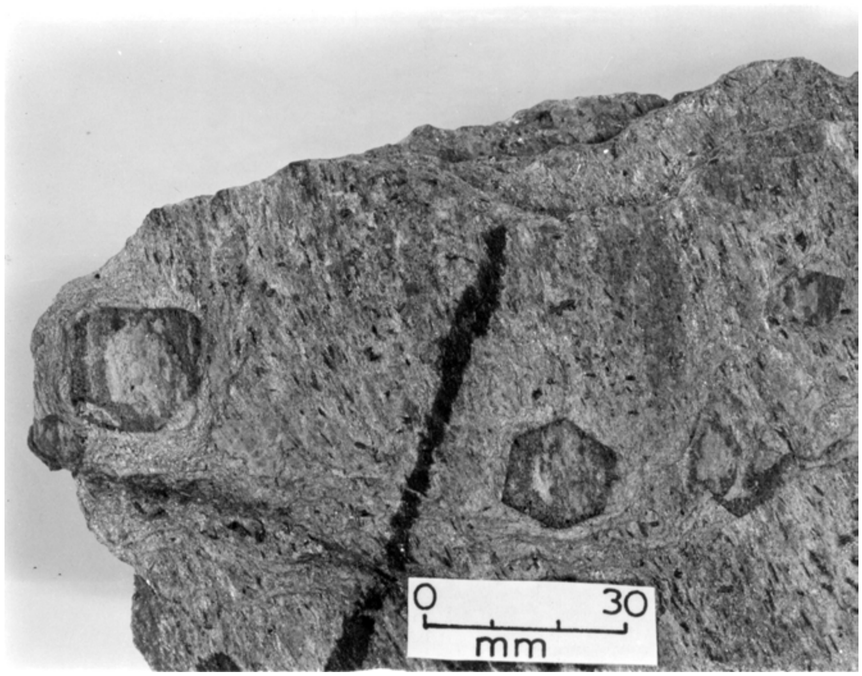


Figure 12 – View of tabular garnets normal to S_2 layering. The garnets, though sliced into tabular sections within the foliation, are equidimensional in this plane. The nearly vertical L_2 lineation indicates the trace of S_1 layering in which the garnet growth occurred. The absence of quartz striping parallel to the lineation on the surface of the garnets indicates no slicing occurred in S_1 in this example.

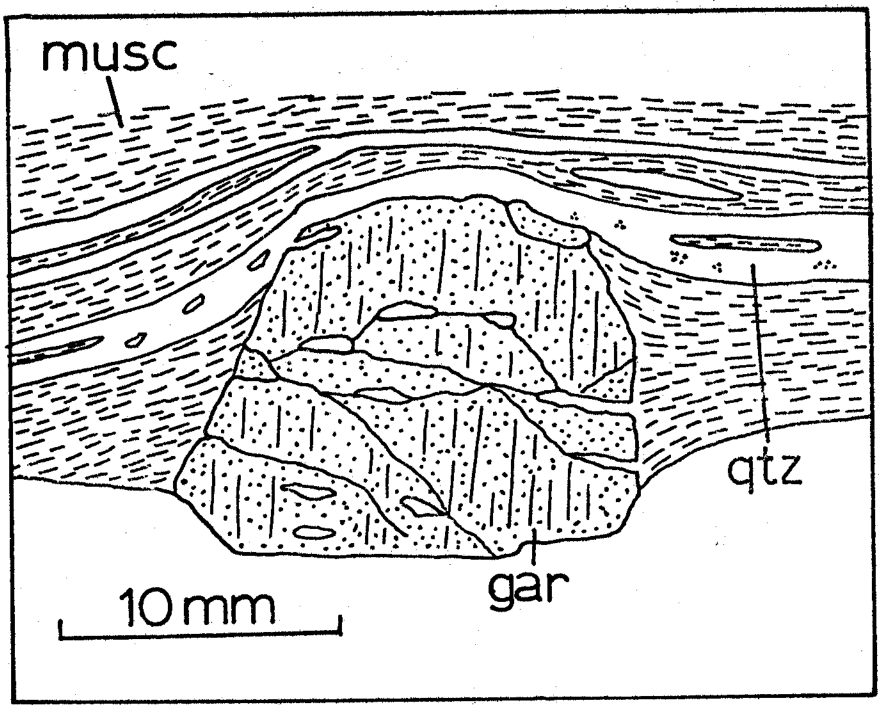


Figure 13 - Garnet crystal in S_2 layering. Early stage fractures subparallel to the foliation are filled with vein-quartz inclusions. The quartz vein above the garnet marks a surface along which other crystals nearby are sliced into sections. Angular fragments of garnet are dispersed in the quartz veins. Vertical lines are cracks of unknown origin.

from the surface of the garent crystal. Note the closely spaced vertical cracks of uncertain origin within the garent crystal.

Figure 14 indicates a further stage in which a tabular section within the crystal has been delineated by quartz veining within fracture surfaces parallel to the rock foliation. Displacement between sections is not apparent. On closer inspection than provided in Figure 14, small garnet fragments can be observed apparently displaced from the bottom surface of the crystal within the quartz vein. Also within the vein, fragments of the surrounding rock constituents appear to be broken away from the walls.

Displacement between adjacent sections of a single garent crystal can be noted in Figure 15. The actual direction of relative displacement lies close to the plane of the photograph, so that the measured faces and angles of the largest sections are still complimentary. Displacements commonly cannot be observed throughout the area of study since it becomes overly speculative to attempt reconstruction of a single crystal after its sections have been separated over distances of a few inches. Rather, what one commonly observes are crystal fragments such as that depicted in Figure 16. Both the top and bottom surfaces are bounded by quartz rich layers, marking the planes along which other sections of the original crystal have been removed. Although sections such as this can be observed in grains as small as 5 mm, the bounded surfaces defined by quartz rich layers may not always be retained.

The garnets discussed so far are mesoscopic examples of sliced crystals. Microscopic sliced garnets are abundant in the Moretown gneisses. The textures of the microscopic garnets are in general similar to the larger sliced garnets; with opaque inclusions and quartz-vein boundaries

Figure 14 – Garnet crystal in which tabular segments have been delineated by quartz veins parallel to S_2. Displacement between sections is not yet apparent. Scale lines are millimeters.

Figure 15 – Sliced garnet segments from the same original crystal. The actual displacement direction lies close to the plane of the photograph, thus the measured dimensions of the respective sliced surfaces remain equivalent. Magnetite inclusions are parallel to layering. Scale lines are millimeters.

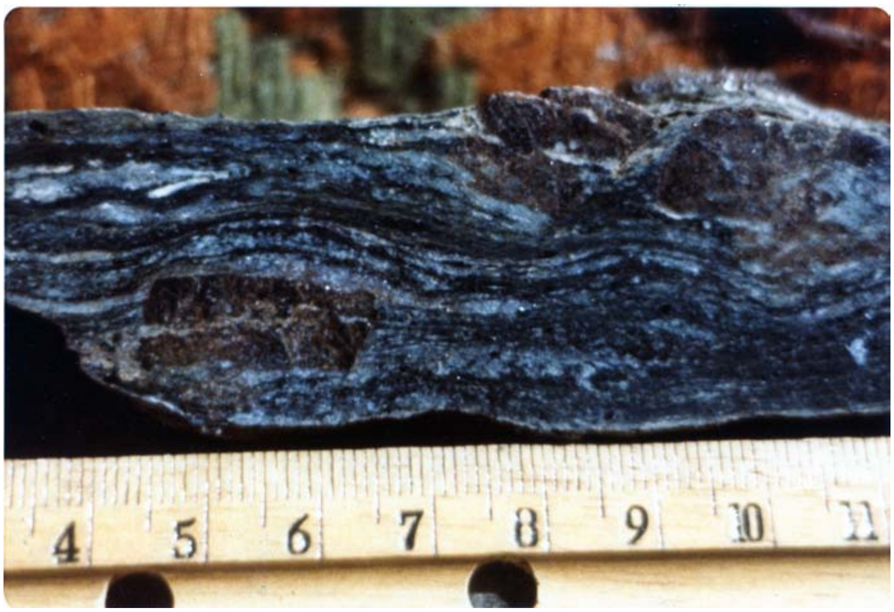


Figure 16 – Tabular garnet section typical of mesoscopic sliced garnets. Segment contains magnetite parallel to S_2 layering and is bounded on top and bottom by vein quartz layering. Weak fracture traces can be seen in garnet at nearly right angles to foliation.

often apparent. The major difference in texture is the irregular boundaries seen on the microscopic crystals in some specimens. In addition to this, one can also observe garnets folded by late Group M-3 folds on the microscopic scale. This situation is less commonly encountered on larger scales because Group 3 folds are not usually well-developed on scales larger than that of crenulations in the rock.

In Figure 17 a typical folded tabular garnet is presented from rock recovered during diamond drilling in the Moretown gneiss. The crystal is bounded by vein quartz on both sides and is folded along with the foliation by what is believed to be a Group M-3 fold. Inclusion trails of opaque minerals are not apparent, but the crystal contains many small quartz inclusions parallel to the foliation. This rock is taken from subarea "B," where S_2 is not well-developed, thus the layering is probably S_1, although it is difficult to be certain about the type of layering observed in drill core samples.

The garnet presented in Figure 18 is also from subarea "B" and the layering is also considered to be S_1. In this crystal magnetite grains are abundant as are the quartz inclusions which also parallel the S_1 layering. This example shows well the type of irregular borders usually seen on microscopic sliced garnets.

It is more common to observe garnet crystal segments only slightly offset when working with thin sections. For example, in a selection of 8 samples prepared from rocks known to contain tabular garnets, 3 out of the 8 studied showed different sections of the same crystal only slightly displaced from the neighboring sections. The remaining 5 sections showed

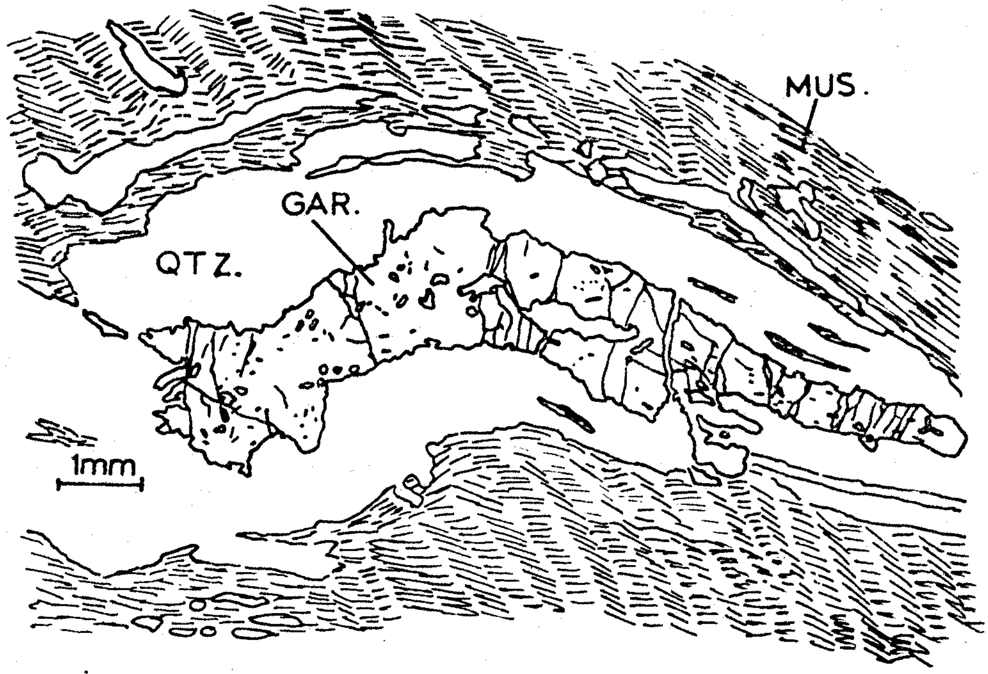


Figure 17 - Tabular garnet from DDH 18-R-73 at 307', in Moretown gneisses. Garnet section is folded along with layering by open crenulations of Style Group M-3.

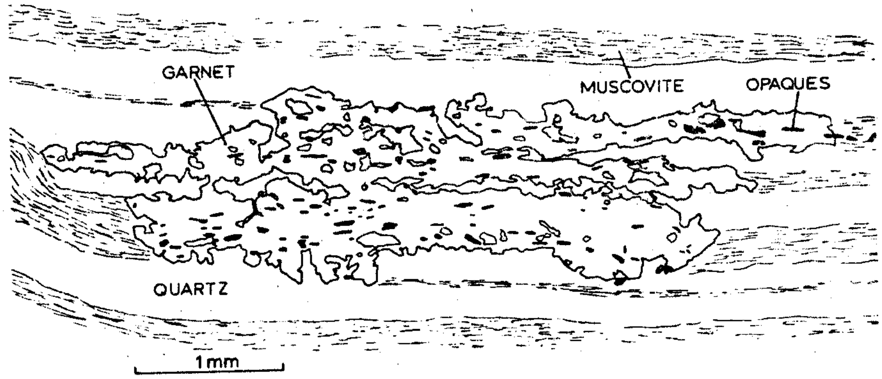


Figure 18 - Tabular garnet segment in Moretown gneiss from DDH 18-R-73, 357'. Garnet is bounded on upper and lower surfaces by quartz-rich layers parallel to rock foliation. Included opaque mineral grains in garnet are parallel to foliation.

nation of oriented specimens in the laboratory may prove useful in determining whether or not slicing maintains a consistent sense of shearing on a certain scale.

(E) Other Garnet Structures in the Moretown

Within the gneissic rocks of the Moretown all of the garnets observed are pre-kinematic with respect to S_1. There are volumetrically insignificant zones of garnet-chlorite schist in the Moretown (refer to Fig. 3) that contain different garents than those in the gneisses, which will be mentioned in passing. Within these rare, centimeter-thick schist layers are undeformed garnets commonly containing trails of opaque minerals always parallel to the surrounding foliation. Quartz inclusions are not usually found within these garnets and the garnets appear to be post-kinematic with respect to the foliation in which they are found.

These schist zones are of uncertain origin. There is at least one type which is apparently parallel to S_1 layering (Fig. 3), and which usually consists of a greenish chlorite-garnet schist as much as 6 cm thick. Similar schist zones are often seen sub-parallel to S_2 layering and it seems possible that both types could be large scale examples of secondary layering associated with S_1 and S_2. Because of the problems cited, the relationships between the post-kinematic garnet forms in the schist and the respective layerings remains unclear.

(F) Alternative Interpretations of Tabular Garnets

Tabular garnets from both metamorphic and pegmatitic surroundings have been described by a number of workers. In some cases the "flattened" appearance has been ascribed to constraints imposed by a pre-existing or simultaneously developing foliation within which the nucleation and growth of the garnet crystal is occurring (Jahns, 1946; Gresens, 1966; Blackburn and Dennen, 1968). Alternatively, tabular shape has been attributed to tectonic flattening of garnet from an initially equidimensional porphyroblast (for example, Dalziel and Bailey, 1968).

Both Jahns (1946) and Gresens (1966) have described pegmatitic garnets in which the crystals are contained within large single crystals of muscovite. These garnets have a "flattened" appearance in the plane of mica layering, with the largest crystal faces bounded by mica (ool) layer surfaces.

Gresens (1966) reasons that garnets growing within the confines of the mica crystal assumed a tabular shape due to differential diffusion of garnet constituents through the anisotropic mica structure. The constituents to form garnet may have diffused from outside the mica crystal or may have resulted from exsolution; in either case, the control of garnet shape by the surrounding muscovite structure seems clear.

The constraint imposed upon growth by a layered mineralogic environment has also been used to explain the existence of tabular garnets in metamorphic rocks (Blackburn and Dennen, 1968). In this instance, garnets from the Grenville gneisses, Gananoque, Ontario, were described as being both tabular and elongated in the direction of a pronounced sillimanite lineation. The elongation was attributed solely to increased rates of ionic diffusion along foliation surfaces, especially in the direction of the lineation, which parallels major and minor fold axes in the rock.

Dalziel and Bailey (1968) describe tabular garnet porphyroblasts from the Grenville front (near Sudbury, Ontario) which they attribute to tectonic flattening. Evidence cited for a flattening process includes the marked asterism and splitting of spots in Laue photographs of deformed garnets, which they attribute to distortion of the crystal. An orthorhombic microfabric of flattened quartz and feldspar in the associated rock fabric is also cited as supportive of the flattening mechanism. They suggest that, garnets, being initially equidimensional, represent deformation ellipsoids, providing no grain boundary migration has occurred.

In many areas of metamorphic rocks field workers commonly attempt to analyze strain in rocks by selecting certain deformed objects which they interpret as approximating strain ellipsoids. Often the most precarious assumptions are made concerning the initial size and shape of the object in question. In the case of deformed garnets, it may be impossible to distinguish a sliced garnet, subsequently modified by deformation under metamorphic conditions, from a garnet deformed entirely by flattening in the same metamorphic conditions. In Table V the three mechanisms which might produce tabular garnets are listed along with different types of internal inclusion structures one might expect to observe. If there are no inclusions of any kind in the garnet, there may be no way to distinguish between the different mechanisms on the basis of internal structure. In this case, however, sliced garnets may be recognized by other observations such as offsets among sections of the same crystal. In the case of zoned inclusions in garnet, the three mechanisms may produce results which are barely distinguishable. Inequant growth may produce some type of zoning with preferred growth in particular direction; therefore, resulting in zoning which is non-concentric. Tectonic flattening of a crystal which contained concentric zoning in its original state would produce similar results (as shown in Table IV). In the case of the slicing mechanism, all early zoning

TABLE III

INTERNAL STRUCTURES DEVELOPED BY
SLICING VARIOUS TYPES OF INCLUSION PATTERNS IN GARNETS

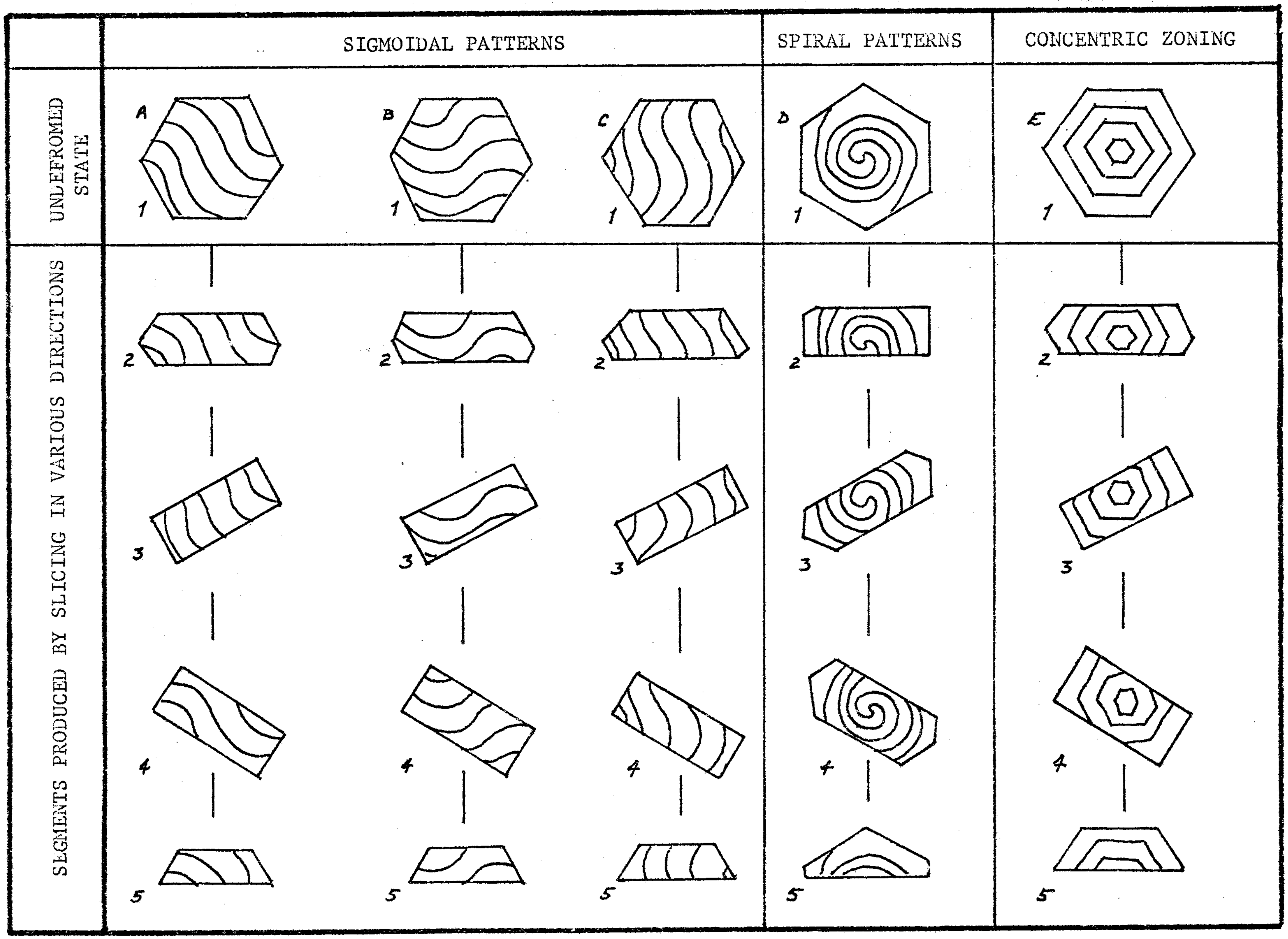


TABLE IV

INTERNAL STRUCTURES DEVELOPED BY FLATTENING
VARIOUS TYPES OF INCLUSION PATTERNS IN GARNET

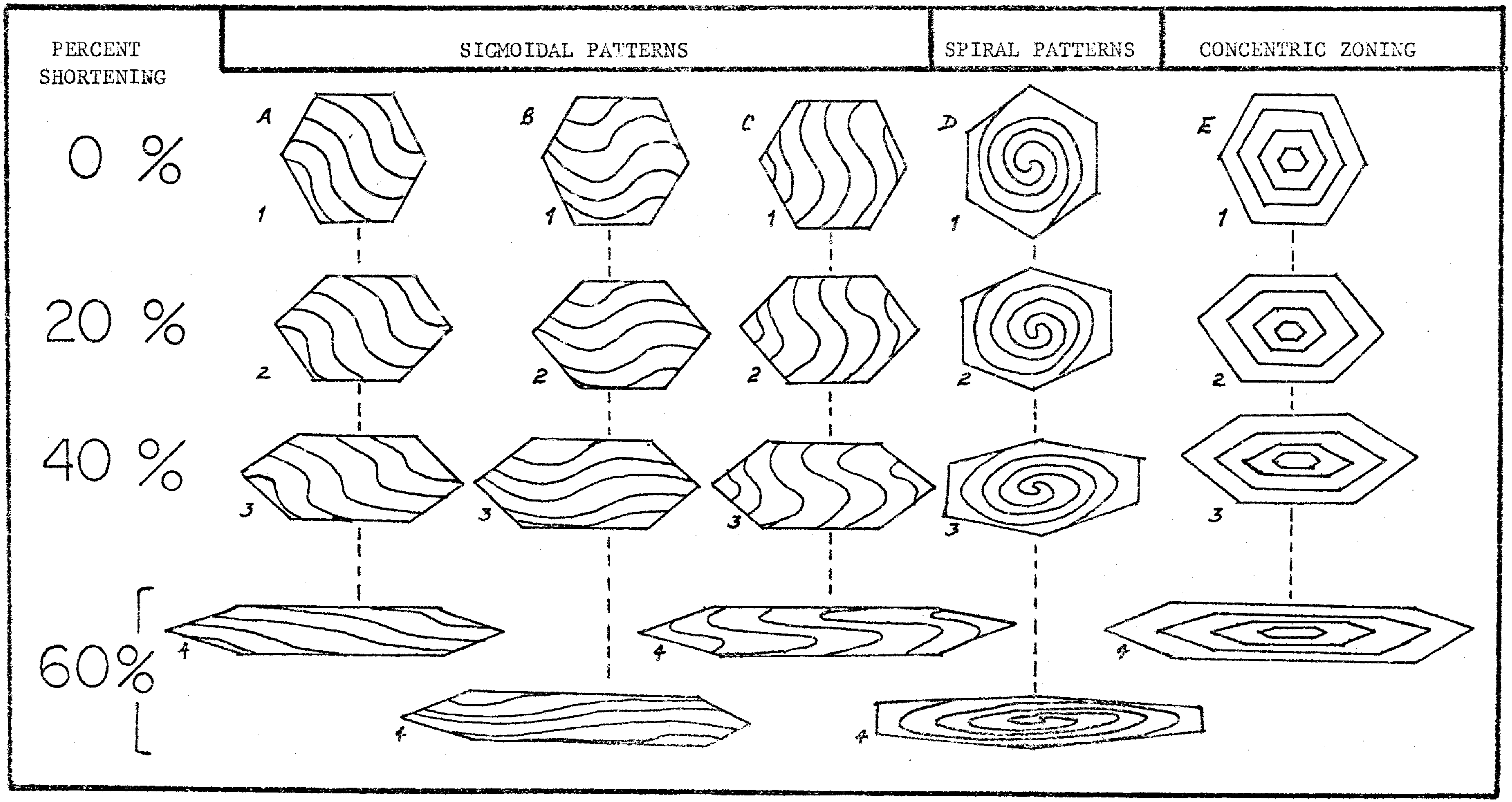
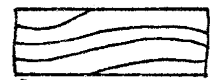
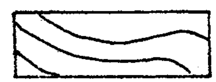


TABLE V

SUMMARY OF POSSIBLE STRUCTURES DEVELOPED IN
TABULAR GARNET CRYSTALS BY VARIOUS MECHANISMS OR PROCESSES

Mechanism \ Inclusions	No inclusions in original or newly growing garnet	New zoning or concentric zoning in original crystal	Spiral patterns in original crystal	Sigmoidal patterns in original crystal
Inequant growth	NO VISIBLE DIFFERENCES		(not applicable)	(not applicable)
Tectonic Flattening		 E-3	 D-4	 B-3
Tectonic Slicing		 E-3	 D-2	 B-3

will be truncated by the sheared surfaces (see Table III), thus it may be possible to distinguish this mechanism from the others.

In the case of inequant growth, no spiral or sigmoidal inclusion structures are expected, since these structures involve other mechanisms relating to pre-kinematic and syn-kinematic growth. Tectonic flattening of garnet with pre-existing structures of these types may result in flattened spiral inclusions or, less obviously, flattened sigmoidal inclusions (Table IV). Again in the case of sliced garnets, any early sigmoidal or spiral structures would be truncated by the sheared surfaces, making recognition of the mechanism possible (Table III).

Thus, of the structures illustrated in Table V, it is thought that only slicing produces results which allow one to postulate the original mechanism. Even then one can never be sure just how much effect either or both of the other mechanisms had on the final shape of the garnet grain. It seems likely that even the best examples of garnets deformed by obvious slicing have had at least some amounts of strain imposed on the internal crystal structure other than by shearing.

Since deformed garnets may have an uncertain deformational history, the basic assumption that all observed grains were equidimensional prior to "flattening" is precluded. Similarly, the assumption that the aspect ratios observed in the tabular garnet are related in some simple way to the total strain imposed on the rock fabric is also problematic.

In a discussion of increased ionic diffusion rates along foliation surfaces in metamorphic rocks (Blackburn and Dennen, 1968) the authors dismiss the possibility of tectonic mechanisms as possible factors influencing the shape of tabular garnets in the rocks they have observed. As evidence excluding the effects of deformation, the authors point to sillimanite lineations which pass undisturbed from the foliation through garnet crystals. An alternative interpretation would be that the sillimanite needles are late overgrowths in the layering and may not have been present during the deformation of the garnet. This is supported by the observation that the sillimanite is a mineral lineation associated with late folds. Thus there is no evidence to exclude deformation as a factor influencing garnet shape.

In summary, although constraints imposed by an anisotropic layering on ionic diffusion rates can be demonstrated in the case of grains enclosed in large, single crystals of layer silicates associated with pegmatites, such a specific mode of origin has not been proven in the case of grains growing in a pre-existing layering under metamorphic conditions. On the other hand, there are abundant examples of equidimensional garnet growth within pre-existing layering of both primary and secondary origin.

DeWit (1972) has discussed the formation of syntectonic porphyroblasts of garnet and criticized all existing explanations by previous workers (Ramsay, 1962; Cox, 1969; Rosenfeld, 1970, for example). The main objections voiced by de Wit are the assumptions made by earlier workers that:

1) the crystal acts as a rigid sphere during the synkinematic process;

2) crystal growth occurs by constant volume increments over the entire surface of the original nucleus during rotation.

Most authors would agree that the first assumption is an approximation to the assumed actual behavior. The second assumption is certainly open to question and many authors have recognized the problems caused by preferred dimensional growth with respect to this assumption (for examples refer to the discussion of previous work on tabular garnet mechanisms). Although de Wit objects to the second assumption, he does not substantially demonstrate a contradiction to it in his observed garnet microstructures. His major thesis is that the garnets may not be growing from single nuclei in what he describes as a "homogeneous" growth phenomenon, but that they may initially grow in a series of nuclei along a veinlet caused, for example, by a hydraulic fracturing mechanism. While his drawings clearly demonstrate what such veinlets might look like, often resembling sigmoidal "tension gashes" with many individual garnets nucleating within, he does not demonstrate any features such as this in actual rock photographs. The closest approximation to the "tension gash" form which de Wit presents (Fig. 19) shows a striking resemblance to the sliced tabular garnets discussed earlier (for example refer to Figs. 17 and 18). The garnets presented by de Wit (1972) usually show trails of quartz inclusions and many are bordered by vein quartz zones similar to those illustrated in Figures 17 and 18. Of interest are the fine dark lines at right angles to the crystal boundaries in the garnet which de Wit assumes to be indicative of layeritic growth from each seperate nuclei of the "heterogeneously" growing garnet. Figures 13 and 14 suggest that these are fractures probably developing as deformational rather than growth phenomena. In these garnets, minute cracks in sub-parallel arrangement normal to rock layering can be observed in both well-deformed and weakly deformed grains. Sections of separated grains show more pronounced and frequent microscopic cracks than do crystals where slicing is less apparent.

De Wit presents a new theory of synkinematic growth from the veinlets that he illustrates, based upon his assumptions that:

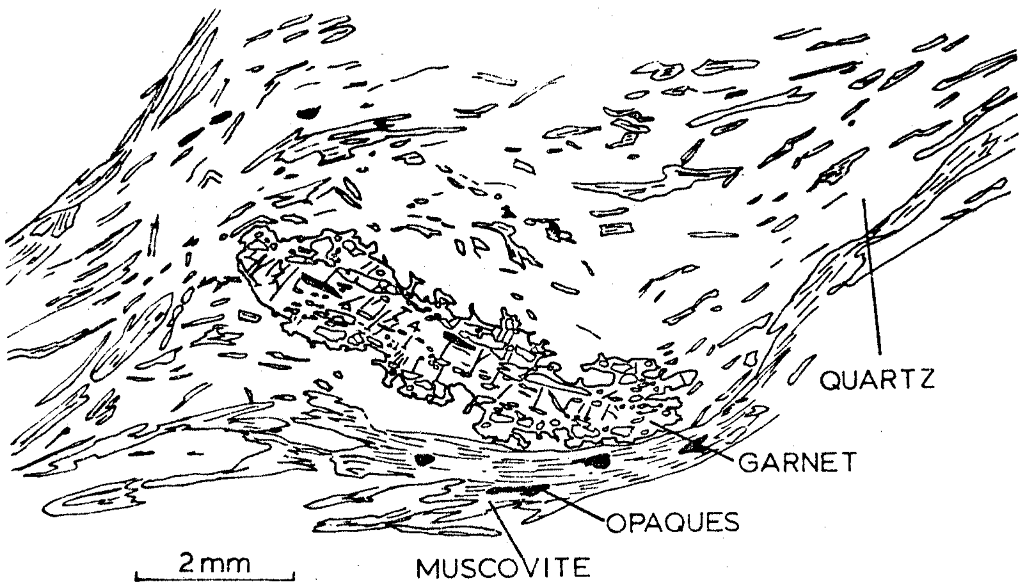


Figure 19 - Tabular garnet segment from the Fleur-de-Lys metasediments in Newfoundland (from de Wit, 1972). Garnet shows striking resemblance to sliced garnets from the Ludlow area; for example, Figures 17 and 18. Sliced garnets on the microscopic scale tend to have more irregular boundaries than mesoscopic garnets. The crystal above is the type example of a grain nucleating in a tension-gash-like fracture according to de Wit (1972).

1) the garnet segments he observes owe their shapes to a growth process;

2) any associated sigmoidal folds in the surrounding layering in which the garnet crystal lies formed before the garent;

3) the development of the sigmoidal, pre-garnet folds in some way influenced the growth of garnet in that region, perhaps by opening up tension fractures within which garnet might begin nucleating.

This writer objects to these points on the basis that the field evidence presented does not unequivocally support the conclusions.

When one considers point three further, it seems more realistic to consider the association of sigmoidal folds around the garnets as evidence for the presence of garnet before the deformation which produced the folds. The quartzo-feldspathic and layer-silicate layering in areas with no garnets may be considered capable of being deformed in a relatively homogeneous manner compared to the areas where garnets are present. The rigidity contrast between the garnet crystal and the neighboring rock foliation probably causes localized structural heterogeneities in the layering as a whole. It is around these points that heterogeneous strains (folds, microfaults, etc.) are likely to occur. This seems to be an equally plausible case in the examples presented by de Wit. This writer believes that the photographic evidence in de Wit (1972) suggests many similarities in the respective histories of the garnets in the Fleur-de-Lys area and those from the Ludlow area.

CHAPTER II

GEOLOGY OF THE CRAM HILL PHYLLITES

Introduction

Rocks of the Cram Hill lie along the eastern contact of the ultramafic zone (Fig. 2). The Cram Hill consists for the most part of dark grey to black "carbonaceous" phyllites with occasional quartzite interbeds from 4 cm to 10 cm thick. A distinctive mappable unit of blue-grey, massive quartzite outcrops in the southern portion of the area.

The rocks in general appear to be less deformed and of lower grade than the Moretown gneisses, although the presence of rare garnets suggests a common point in their respective metamorphic histories. An epidote-amphibolite grade is suggested from amphibolites contained in the Cram Hill phyllites near the ultramafic zone.

Mineralogy

Estimated modes for the Cram Hill are given in Table VI and the details concerning sample collection are the same as those discussed in the previous chapter. The Cram Hill phyllites typically contain 47% quartz + albite, 24% muscovite, 11% magnetite and other opaques, 6% carbonates, 5% biotite, and 6% chlorite. Pyrite is present in quantities of about 1% and garnet is present in trace quantities.

Structural Studies in the Cram Hill Phyllites

(A) General Description of the Cram Hill Mesostructure

The overall structural development of the Cram Hill is relatively simple in comparison to the Moretown gneisses. In the Moretown, work on structural elements relies upon the classification of foliations

TABLE VI

ESTIMATED MODES FOR THE CRAM HILL PHYLLITES IN VOLUME PERCENT CONSTITUENTS BY TEN FOOT INTERVALS IN DIAMOND DRILL HOLE 18-R-73

FOOTAGE	N	QUARTZ + ALBITE	MUSCOVITE	MAGNETITE	CARBONATE	BIOTITE	CHLORITE	PYRITE	GARNET
47'	200	52	24	5	5	9	3	3	tr.
57'	200	42	30	18	tr.	7	1	2	0
67'	200	54	25	3	4	5	8	2	0
77'	100	32	42	13	0	1	12	0	0
87'	200	42	22	12	19	2	tr.	4	0
97'	300	60	15	11	7	4	0	3	0
106 a *	200	53	19	12	8	5	5	tr.	0
106 b *	200	51	16	12	10	5	5	2	0
117'	200	49	27	10	10	1	3	tr.	0
126'	200	39	26	23	3	3	6	0	0
138'	200	45	27	5	8	4	12	0	0
167'	200	47	12	12	0	9	19	1	0
Average		47	24	11	6	5	6	1	tr.

*Different areas of same thin section counted.

and their relationships to different styles of folds with relatively infrequent examples of overprinting. In the Cram Hill, however, style groups of structural elements are more easily defined. In addition, there are many examples of overprinting which could allow a straightforward classification of fold types into generations. For the sake of continuity, however, the folds will be discussed in this section in terms of the style group classification.

(B) Style Groups in the Cram Hill

The style groups described in this section are characteristic only of the Cram Hill and are not to be confused with style groups of the same Arabic numerals in the Moretown. The relationships between the style groups in the Moretown and those in the Cram Hill will be discussed in a later section.

1. Early layering in the Cram Hill -- The earliest layering in the Cram Hill phyllites is a compositional banding consisting of layers of quartzite of varying thickness alternating with thicker sections of phyllitic rock. In some cases, specifically in very thick units of quartzite, a gross internal layering is visible. As one traverses such sequences, individual layer thicknesses change abruptly across the section and can range from 5 mm to 40 mm. These sequences do not show the typical A-B-A-B repetition of textural or mineralogic character commonly seen in many types of rocks containing secondary layering. For these reasons the author classifies these layerings as possible sedimentary structures. These types of S_0 layering are illustrated in Figures 20 and 21.

2. Elements of Style Group C-1 -- Group C-1 folds are characteristically isoclinal to tight folds of compositional layering (S_0) (Figure 20). Within the "carbonaceous" phyllite sequences these folds are best

seen in the interbedded quartzite layers as illustrated in Figure 20. Well-developed axial plane foliations (S_1) are found in some Group C-1 folds, but such foliations are not ubiquitous. S_1 usually consists of finely spaced layer-silicate domains ranging from .5 mm to 2 mm in thickness, which cross-cut S_0 layering in isoclinal fold hinges and parallel S_0 layering in Group C-1 isoclinal fold limbs. In many cases, segments of type C-1 folds are offset across S_1 in a manner similar to a crenulation cleavage. S_1 occasionally forms anastamosing textures around S_0 domains, and S_0 layers in isoclinal group C-1 fold limbs may become isolated in this manner (Fig. 23). Although examples such as this are referred to as "transposed layering" by some, the structures have developed by the intersection of an early layering by a later layering rather than by attenuation and severing of early layering in the manner discussed by Sander (1911).

A weak lineation (L_1) is occasionally seen on S_1 surfaces and is defined by the intersection of S_0 and S_1. A weak color banding may be observed on S_1 surfaces as part of this lineation and is due to low angle intersections of S_1 with layers of different compositions in S_0 (Fig. 26).

Within the thick quartzite unit found in the southern section of the study area Group C-1 folds are usually tight folds of S_0 layering, and are often overprinted by group C-2 folds (Fig. 24). Group C-1 folds in the quartzite contain a distinct secondary layering (S_1) as an axial plane foliation. S_1 here consists of light-colored quartz-rich bands ranging from a few mm to 2 cm thick, which alternate with darker colored domains containing S_0 layering. S_1 is often more prominent in an outcrop than S_0 and could be mistakenly referred to as sedimentary layering.

TABLE VII - SELECTED STYLE GROUP ELEMENTS
IN THE CRAM HILL PHYLLITES

	PHYLLITIC ROCKS			QUARTZITES	
STYLE GROUP	FOLDS	FOLIATIONS	LINEATIONS	FOLDS	FOLIATIONS
C-2	Usually open with planar axial surf. (figs. 20, 22)	S_2 crenulation cleavage. (fig. 22)	L_2 crenulations (fig. 26)	Open with planar axial surf. (figs. 21, 27)	S_2 quartz-rich layering (figs. 27, 28)
C-1	Usually isoclinal with folded axial surfaces. (fig.20)	S_1 mica-rich layering. (fig. 23)	L_1 intersection lineation of S_1 and S_0. (fig. 24)	Open to tight folds. (fig.24)	S_1 quartz-rich layering (figs. 27, 28)
---	----	S_0 sedimentary compositional layering. (fig. 20)	----	----	S_0 compositional layering, probably sedimentary. (fig. 21)

Figure 20 – Early layering, probably sedimentary in origin, in the Cram Hill phyllites. Group C-2 folds with nearly vertical axial planes can be seen folding S_0 and overprinting isoclinal Group C-1 folds. View faces north in study area.

Figure 21 – Early coarse compositional layering (S_0) in Cram Hill quartzite, probably sedimentary in origin. Note that quartzite layering is parallel to lower contact with the phyllites. Weak S_2 layering may be seen as an axial plane foliation to Group C-2 folds, oriented roughly vertically in photo.

Figure 22 – Open to tight folds of Group C-2 style overprinting early isoclinal Group C-1 folds of S_0 with S_1 . Strong S_2 crenulation cleavage (vertical) is visible in the surrounding phyllite. Top of photo is towards north in area.

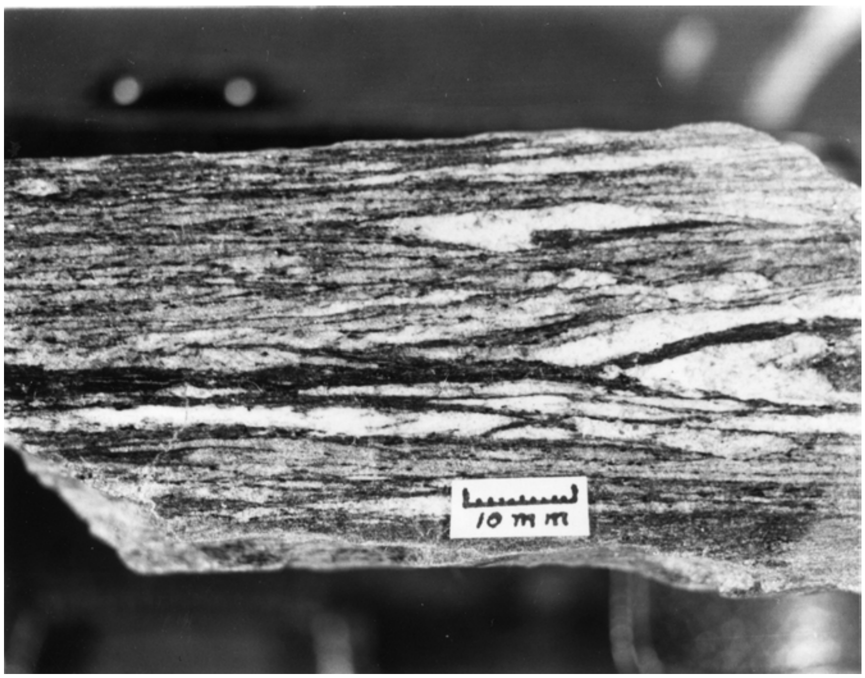


Figure 23 – Strongly developed secondary layering parallel to axial planes of Group C-1 folds. S_1 is not usually developed to this extent in the Cram Hill phyllites. Note the isolated fold closures producing an appearance easily mistaken as transposed layering. In this case the old layering can be seen to be cut by S_1 , not true of transposed layering.

Figure 24 – Massive quartzites from the southern portion of the study area in the Cram Hill. Group C-2 folds with nearly vertical axial planes overprint Group C-1 folds of the early layering.

Figure 25 – Quartzite layering in massive Cram Hill phyllite. Group C-2 fold axial surfaces run horizontally in the photo, subparallel to S_2 in the surrounding phyllite. Early folds, refolded by Group C-2 folds, can be seen. Axial planes to early Group C-1 folds are folded but generally cross the photo in a vertical direction. North in photo is the direction towards the upper right hand corner from the center.

With respect o metamorphic minerals associated with Style Group C-1, it is sufficient to note that garnet crystals are undisturbed within S_1 layering, but are consistently cut and deformed by elements of Style Group C-2. Undeformed garnets in S_1 are therefore considered an element of Style Group C-1. A later section on garnet microstructures in the Cram Hill will discuss Style Group C-1 garnet in detail.

3. Elements of Style Group C-2 -- Mesoscopic folds of this style group are open folds of S_0 and S_1 layering (refer to Figs. 20, 22 and 24). Within the phyllites, Group C-2 folds usually range in wavelength from 1 to 10 cm and commonly contain an axial plane crenulation cleavage (S_2) usually consisting of thin micaceous films or layers up to 1 mm which cut S_0 and S_1 layering.

Minute crenulations of S_1 surfaces usually define L_2, the only lineation associated with Group C-2 folds (Fig. 26). S_1 surfaces, such as that illustrated in Figure 26, commonly contain intersecting L_1 and L_2 fabric elements.

Within the quartzites, Group C-2 folds are more common on scales ranging from 5 m to 10 m in wavelength, but are also seen on scales ranging from 1 to 10 cm. The fold shapes are usually open with fairly rounded hinges and are similar to the fold styles observed in the phyllites. Group C-2 folds overprint Group C-1 folds in the quartzites (Fig. 24), and contain a well-developed axial plane secondary layering, here referred to as S_2. S_2 in the quartzites is defined by light-colored, quartz-rich bands ranging from a few mm to 2 cm thick, which alternate with microlithons containing both S_0 and S_1 layering. In Figure 27, S_2 is oriented vertically and can be seen as an axial plane foliation to Group C-2 folds in

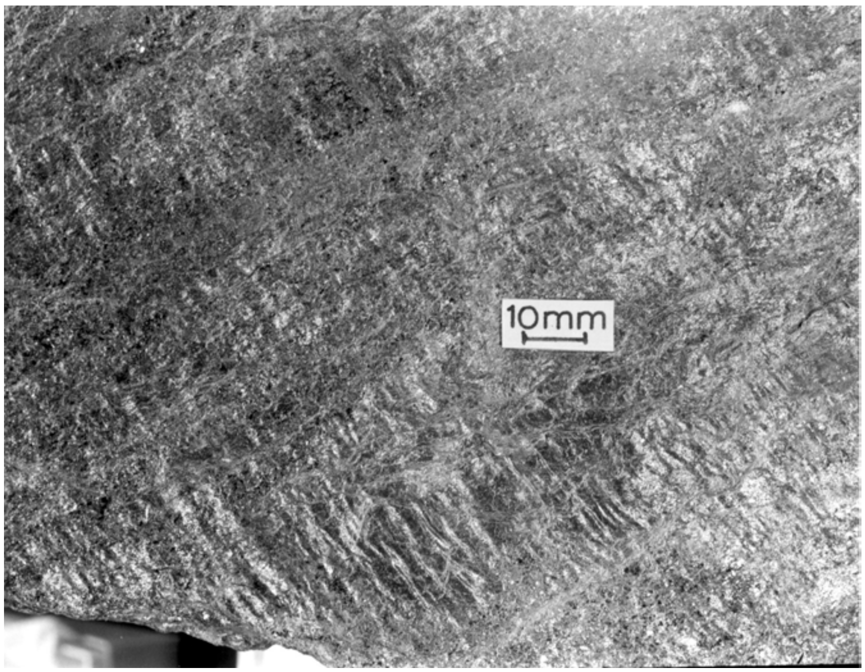


Figure 26 – Lineations observed in the Cram Hill phyllites. L_2 , which crosses the photograph from lower right to upper left, consists of the axes of small crenulations of Style Group C-2. L_1 , which crosses from lower left to upper right, is a faint color banding formed by the intersection of S_1 with S_0 on S_1 surfaces. L_2 crenulations overprint L_1 .

S_1 layering. Note that S_2 and S_1 are identical in appearance and may be impossible to distinguish without the benefit of overprinting relationships (Fig. 28). As a further complication to field mapping, both S_1 and S_2 are always more prominent in an outcrop than the primary S_0 layering. All of these factors make the interpretation of layering in this area very difficult.

(C) Garnet Microstructures

While garnets are not often seen in the Cram Hill phyllites, the known examples contain microstructures usually similar to those in Figures 29 and 30. The sketch below Figure 29 represents the setting of both Figure 29 and 30 on a larger scale. The quartz-rich areas of both figures correspond to the white areas in the sketch, and the mica-rich areas in both figures represent the black S_2 cleavage domains in the sketch. The classic explanation for the garnet microstructures in these figures would be an overgrowth of late garnets on the S_1 and S_2 layering; whereby the inclusions in each garnet crystal are merely relicts of the early layering that existed in the region now overgrown by garnet. One implication of this suggestion is that the growing crystal does not mechanicly disturb the pre-existing layering. This is the basic assumption underlying the interpretation of so-called "helicitic" post-tectonic crystal textures observed in a variety of minerals. If one accepts the interpretation without further examination, a late growth of garnet is seen on S_1 and S_2, and thus the conclusion may be made that garnet grade metamorphism occurred after, or perhaps late in, the second deformation.

Before such important conclusions are made, however, it is necessary to inquire whether we need to ask ourselves if there are any other ways to produce the "helicitic" textures, specifically ways which may not involve an overgrowth or post-tectonic mechanism. Of paramount importance to this question is the critical examination of the single piece of evidence available -- the inclusion pattern itself. How do we know whether the inclusions grew before or after the crystallization of the garnet? If we claim that the inclusions grew after the crystallization of the garnet, then the mechanism of formation is not a post-kinematic overgrowth, but rather some type of deformation process affecting a pre-tectonic garnet. One such process might be the effect of shear displacements in a surrounding layering upon pre-existing garnets. If the garnets are affected by such a strain field, it does not seem unlikely that weak shear displacements may occur in the garnet parallel to the layering. These zones of weakness in the garnet crystal may be ideal locations for the nucleation and growth of the "inclusion" minerals. Further deformation of these crystals could account for many of the observed fold shapes seen in the inclusion patterns.

It is probable that both types of mechanisms are operating in metamorphic rocks. The problem is then to distinguish which is the most likely explanation for a given example, in order that something may be said about the metamorphic conditions during various deformation events. In the example shown in Figures 29 and 30, the author favors the prekinematic view of garnet growth with respect to the formation of S_2. The garnets in Figure 29 occupy a quartz-rich "S_1" area for the most part, with some garnets partially protruding into the mica-rich "S_2" area. For those garnets in the quartz-rich "S_1" areas, the author favors the classic "overgrowth" explanation, with respect to S_1 layering, but where these garnets lie within a mica-rich "S_2" area, the author favors a prekinematic origin with

Figure 27 – Style Group C-2 fold in massive Cram Hill quartzite folding earlier S_1 secondary layering offset along new S_2 domains. S_2 is nearly identical to S_1 and is an axial plane foliation to Style Group C-2 folds.

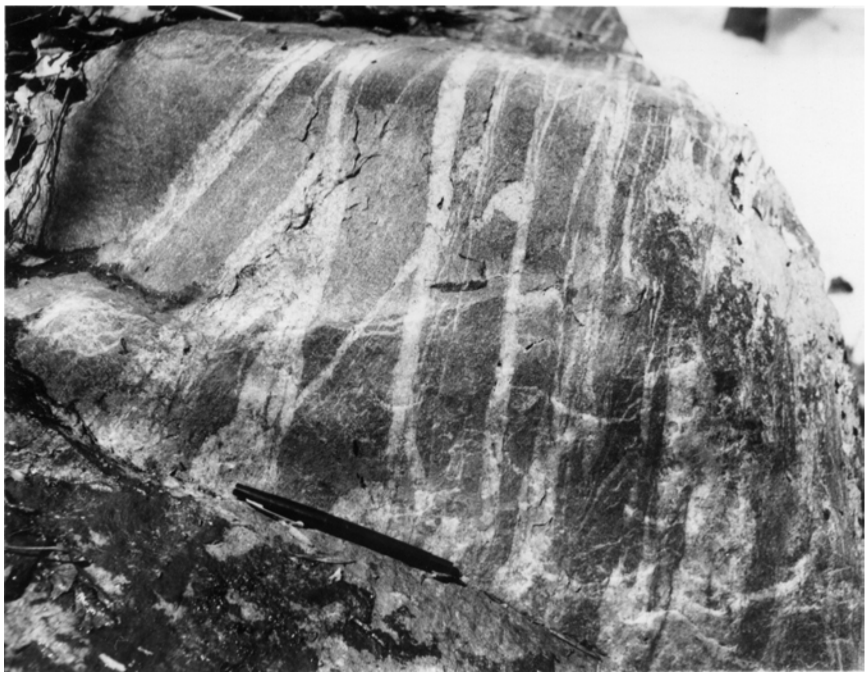
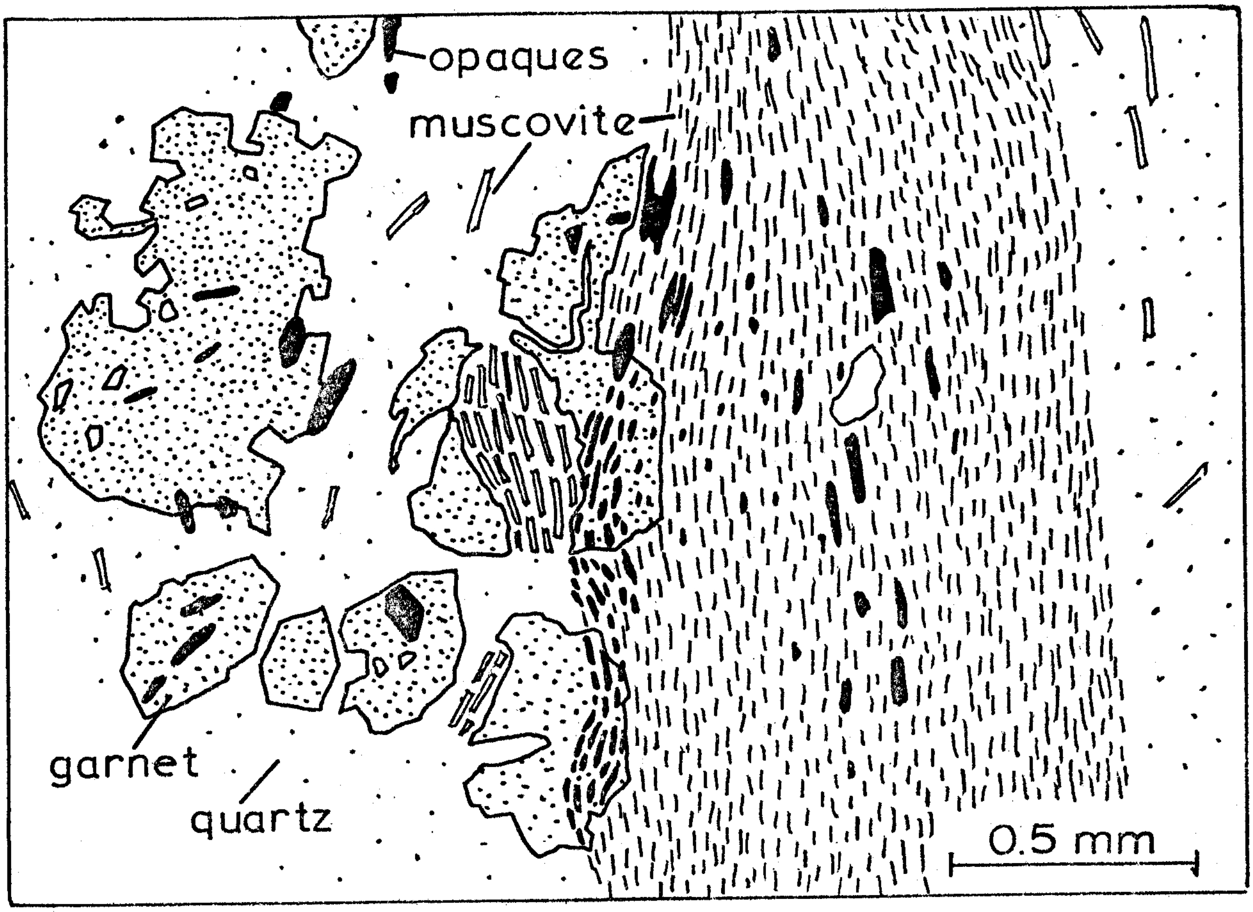


Figure 28 – S_2 and S_1 layering in Cram Hill quartzite. Note the strong similarity between S_1 and S_2 , making differentiation of layering types by the observer very difficult. S_0 layering cannot be seen.



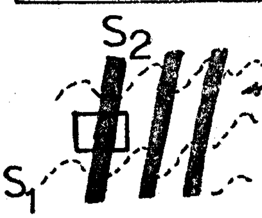


Figure 29 - Deformed garnets in Cram Hill phyllite. The quartz-rich area corresponds to S_1 layering and the vertical mica-rich area represents an S_2 foliation. Note that the inclusion patterns in the garnets near S_2 are much higher in opaques than the S_2 layering in general. The author suggests that garnets grown in S_1 are deformed by S_2 and that the inclusion pattern is a result of deformation.

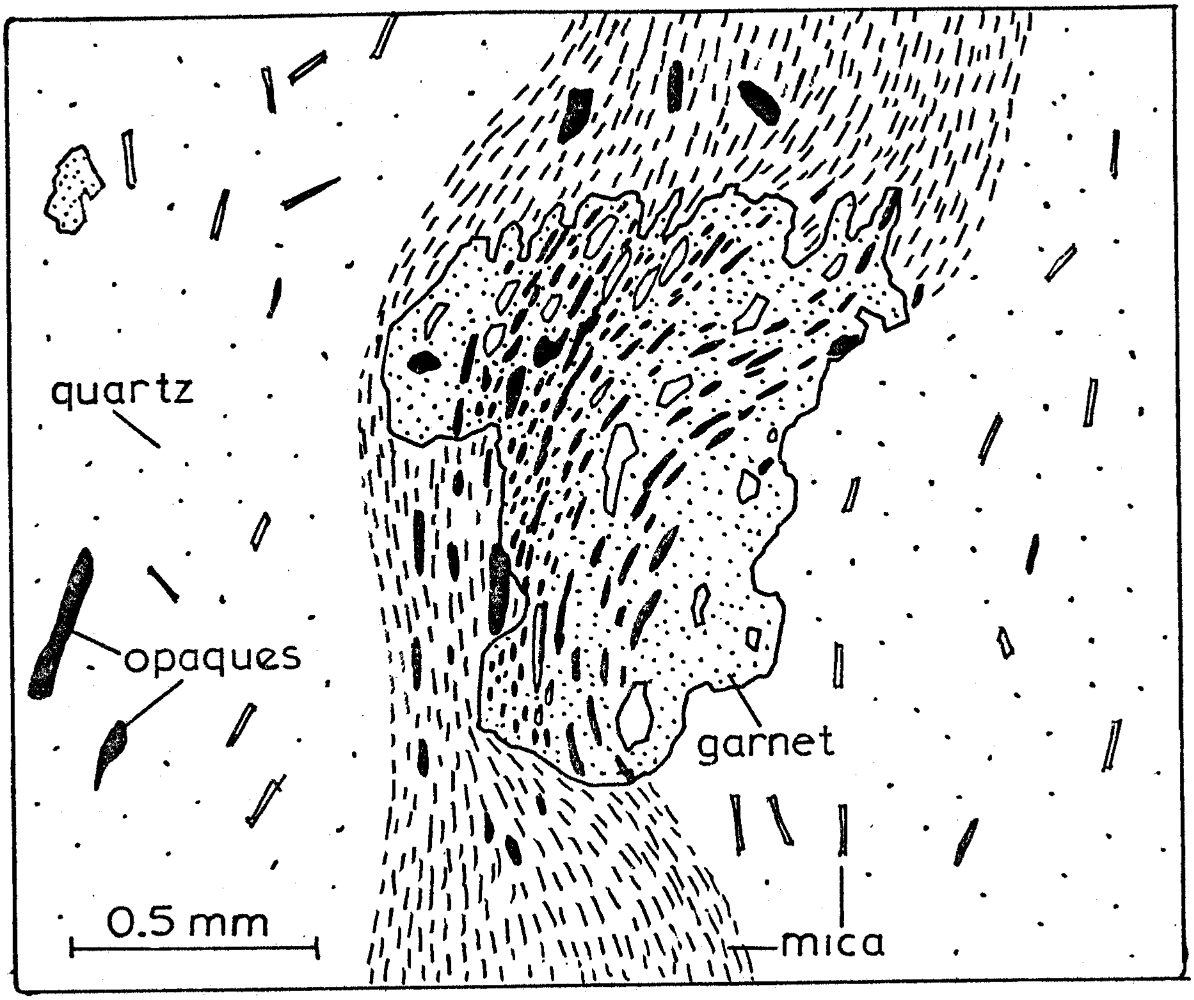


Figure 30 - Deformed garnet in Cram Hill phyllite. The mesoscopic setting is the same as the preceding figure. S_2 is again defined by the mica-rich layering. In this case, however, the garnet is almost wholly contained within the S_2 area. Again the opaque content is higher in the garnet than in S_2 as a whole, and the portion of the garnet falling outside of the S_2 area is nearly free of opaques. The author suggests that the sigmoidal pattern in the garnet is a result of deformation rather than overgrowth.

TABLE VIII

STYLE ELEMENTS ASSOCIATED WITH GARNET MICROSTRUCTURES IN THE CRAM HILL PHYLLITES

STYLE GROUP	LAYERING	GARNET STRUCTURES	MAGNETITE STRUCTURES	CHLORITE STRUCTURES
C-1	S_1	Undisturbed garnets in S_1, minor opaque inclusions.	Early magnetite in S_1 in rod-like forms parallel to layering. Crystal aggregates in S_1 (not parallel to layering).	none apparent.
C-2	S_2	Secondary planar discontinuities in early garnets, including quartz inclusions.	Secondary magnetite parallel to S_2 in S_2 domains and garnet crystals.	Chlorite intergrowths between garnet fragments being disrupted along or near S_1-S_2 boundaries.

respect to S_2. The main reason for this choice is the high opaque contents associated with the garnets in this area, as well as the garnet in Figure 30. The mica-rich S_2 domains show much lower magnetite content than these garnets and one would be hard-pressed to explain this difference in terms of pure overgrowth. In summary, garnet growth is believed to have occurred before the development of S_2 and garnets in general are deformed where they are close to or within S_2 mica-rich domains.

(D) Macroscopic Structural Studies in the Cram Hill

The only adequate structural "marker" in the area is the massive blue-grey quartzite found in the southern portion of the map (Fig. 2, area B). Within this horizon, Style Group 2 elements were studied in detail on a mesoscopic scale.

The Cram Hill quartzite has been shown to contain a number of types of layering on a mesoscopic scale. On a macroscopic scale, however, these layerings are not discernable, and the most important surface is the interface between the massive quartzite and the surrounding phyllites.

The quartzite is well-exposed and outcrops as a series of discontinuous bodies often elongate parallel to the mesoscopic S_2 foliation in the surrounding phyllites. The bodies vary morphologically from remnant isoclinal fold closures up to 40m in width to attenuated and detached fold limbs from 5 to 10m wide (Fig. 31). These dismembered fold segments indicate that macroscopic transposition of the quartzite from an originally continuous horizon has occurred.

The quartzite bodies, when viewed as a whole, define an enveloping surface which approximates the trend of early layering in the Cram Hill sequence on a large scale in this area. Extended mapping of the quartzite

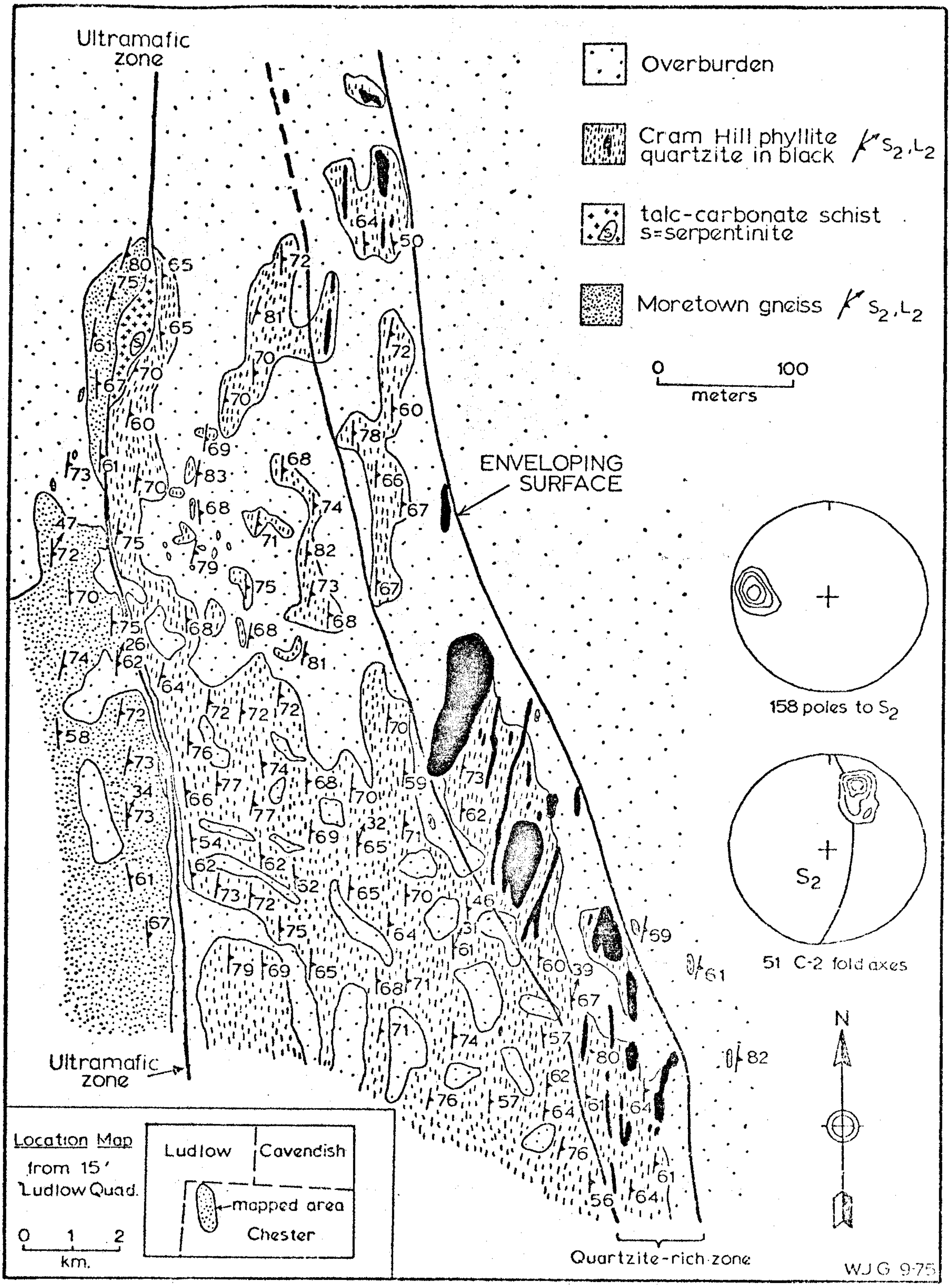


Figure 31 - Geologic map of the southern section of Figure 2 (area "A") along the Moretown - Cram Hill contact. Early layering defined by the enveloping surface of the macroscopically transposed quartzite in the Cram Hill is discordant to the contact and the ultramafic zone.

to the northwest along the trend of the enveloping surface has shown that early layering in the rocks is discordant to the northeast trending contact between the Moretown and Cram Hill members (SIC) of the Missisquoi Formation (Fig. 31). The quartzite can be traced to within one-hundred feet of the contact where it becomes strongly transposed into thin lenses elongate parallel to the regional mesoscopic foliation (S_2). In some cases the early layering is cut at high angles by the blackwall contact of the ultramafic zone. Although the quartzite has not been recognized anywhere along strike on the opposite side of the fault zone, it represents an important example of the truncation of early layering by the ultramafic zone in Vermont.

CHAPTER III

GEOLOGY OF THE ULTRAMAFIC AND RELATED ROCKS

Introduction

The ultramafic zone consists predominantly of serpentinite masses surrounded by talc-carbonate zones of varying thickness. The size of the serpentinite bodies varies from masses 1 km long and 500 meters wide to small discontinuous podiform bodies 30 to 50 meters long and a few tens of meters wide. Most of the serpentinite bodies are elongate along the general strike of the ultramafic zone (refer to Fig. 32).

Situated near the ultramafic zone are various elongate bodies of amphibolite in the country rocks. The amphibolites generally are more abundant near the boundaries of the ultramafic zone, and are rarely encountered over 200 meters away from the zone. The amphibolites are usually less than one meter thick and typically are continuous along strike for 5 to 10 meters. A series of amphibolite lenses form discontinuous zones up to 500 meters long in the Moretown, occasionally parallel to S_2 layering in the gneisses. Cram Hill rocks rarely contain amphibolites but those encountered show deformational histories similar to the surrounding phyllites.

The ultramafic rocks were mapped initially by Richardson (1928) as a single mass of peridotites and pyroxenites about 4 km long and 1 km wide. Subsequent work by Thompson (1950) indicated the body consisted of three separate units rather than a single large mass (refer to Fig. 2 for Thompson's interpretation). Thompson did not observe any primary minerals in the ultramafics and considered the rocks to be completely serpentinized. Work by the author, however, has shown clear zoning in the ultramafic

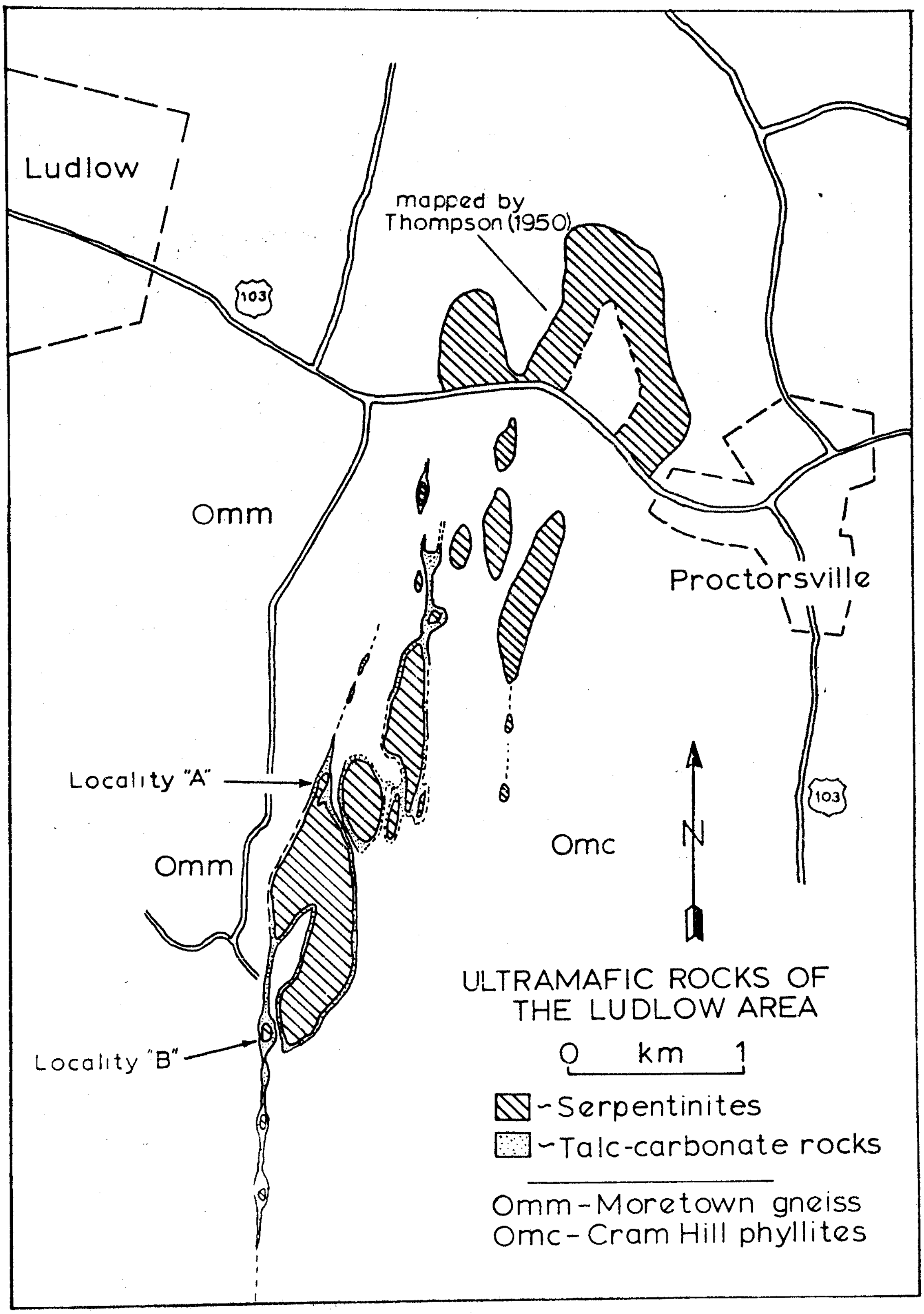

Ludlow
mapped by
Thompson (1950)
103
Omm
Proctorsville
Locality "A"
Omm
Omc
N
103
ULTRAMAFIC ROCKS OF
THE LUDLOW AREA
0 km 1
Locality "B"
-Serpentinites
-Talc-carbonate rocks
Omm-Moretown gneiss
Omc-Cram Hill phyllites

Figure 32.

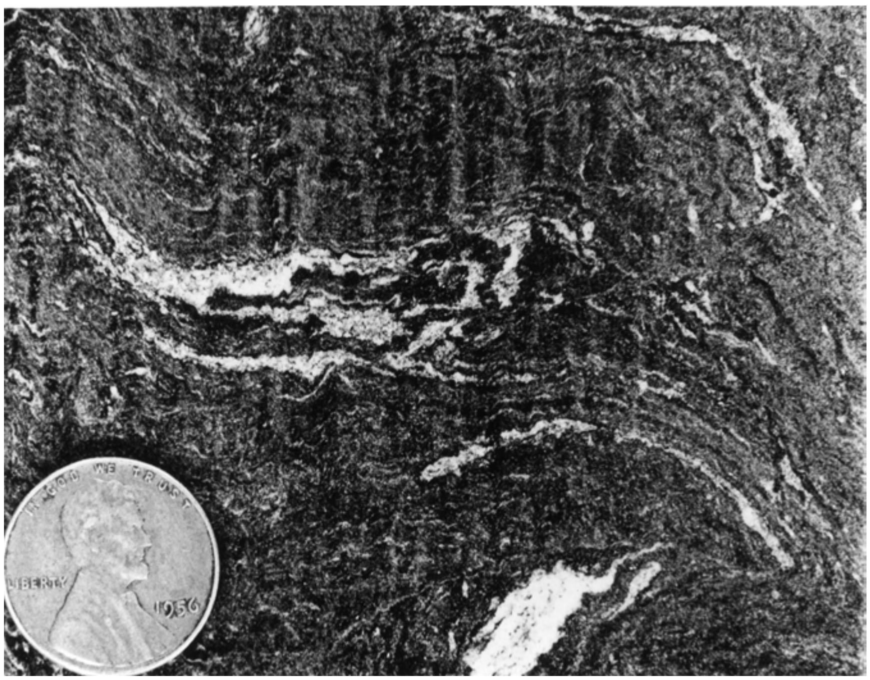


Figure 33 – Layering in the amphibolites of the Cram Hill. S_0 layering (white) is folded by isoclinal folds which are in turn folded by late open folds with vertical axial planes in the photograph. S_2 layering can be seen as an axial plane foliation to the late folds.

masses with up to 50% primary minerals (dominantly pyroxenes) found in specimens from the cores of the ultramafic bodies. Detailed mapping by the author has shown that the units delineated by Thompson can be further defined as a series of at least six smaller, discontinuous masses. The largest body, north of Route 103, has not yet been mapped in detail but the author considers that it is likely to consist of a series of smaller bodies as well.

Structure and Mineralogy of the Amphibolites

The amphibolites within the Cram Hill usually occur within 10 to 20 meters of the ultramafic zone. They are usually fine-grained rocks composed of blue-green pleochroic amphiboles, albite, carbonates, and chlorite in varying proportions. The feldspathic and carbonate minerals are often segregated in layers with little or no amphiboles present. These layers are the dominant element of the earliest compositional layering recognized in the amphibolite (S_0). This layering is disposed in tight to isoclinal early folds and is later refolded by open folds (Fig. 33). A new layering (S_2) is found as an axial plane structure to the late folds and consists of kink-like domains poorer in feldspathic and carbonate minerals than the S_0 layering. Both of these fold types are believed to correlate with the style groups in the Cram Hill phyllites. The amphibolites are not demonstrably oblique to S_0 in the Cram Hill phyllites, but their parallelism to S_1 in the phyllites suggests such a relationship. This structural data seems to indicate an early tectonic emplacement of the amphibolites before the major deformational episodes affecting the Cram Hill phyllites. The close spatial relationship between the serpentinites and the amphibolites may indicate similar, but not necessarily synchronous, emplacement histories.

The rocks in the ultramafic zone do not seem to have been involved in the early deformation that produced features such as the early isoclinal folds in the amphibolite and the Group 1 folds in the Cram Hill phyllite. This may suggest a slightly earlier emplacement for the amphibolite along the same zone.

The amphibolites in the Moretown usually occur not more than 100 to 200 meters away from the ultramafics and are generally more abundant than the amphibolites in the Cram Hill. In rarer cases very thin amphibolite layers less than 1 ft thick may run parallel to the ultramafic zone at less than 2 ft from the contact. These rocks are medium to coarse-grained and may occur as strongly layered lenses or larger, unlayered bodies up to 1 meter thick. The rocks contain abundant blue-green amphibole and albite, and minor amounts of carbonate, chlorite, and opaque minerals. Amphibole minerals are often weakly oriented, forming a lineation which is subparallel to L_2 in the Moretown gneisses. The rocks as a whole are usually lighter in color than those in the Cram Hill and are more strikingly layered. The amphibolite bodies are generally parallel to S_2 in the gneisses and may be continuous for many tens of meters. It is possible that more than one generation of amphibolites exist and that some of the bodies may be earlier structures than those of Style Group 2. In addition to the amphibolites observed in the Moretown, extremely rare amphibolite inclusions have been found in the talc-carbonate rocks of the ultramafic zone. It is not possible to say whether or not they represent tectonic inclusions or late intrusions into the ultramafic zone.

Granitic Rocks Associated With the Ultramafic Zone

Leucocratic granitic bodies have been found in two associations with respect to the ultramafic zone. The first type is associated with the amphibolites in the Moretown, while the second type is contained within the ultramafic zone and extends into the Cram Hill phyllites. Both of these types are relatively small bodies. The first is known only from a single occurrence which runs parallel to an amphibolite body but is separated from it by a 10 cm quartz vein. Potassium feldspar in the granite is distributed in relatively equant grains in undeformed specimens and fragmented grains in examples that are strongly foliated. The potassium feldspar contains characteristic microcline twinning and is often accompanied in the strained zones by recrystallized quartz and layer silicates, which define the foliation.

The second occurrence of granitic rocks consists of large masses of fine-grained tonalite intruded into the talc schists of the ultramafic zone. The intrusion occurred after emplacement of the serpentinite and after steatization as well, since the tonalites pass undisturbed from the ultramafic rocks into the Cram Hill phyllites, cutting across Group 1 folds in the phyllite. The tonalite bodies are usually elongate in the direction of the ultramafic zone. Group 2 folds are very weak in the country rocks around the tonalite, usually consisting of minute crenulations, thus it is not possible to observe whether or not the tonalite is deformed by Group 2 folds. Rocks of this type of occurrence are not usually foliated except along the contacts with the talc schists where a reaction zone of chlorite schist from 4 cm to 20 cm thick occurs. A weak orientation of biotite micas may occur in the tonalite near the alteration zones. The writer considers it possible that the intrusion is associated with Group 2 elements in the Cram Hill, though not enough evidence has been accumulated to prove this relationship. If the relationship is true, however, radiometric dating of the granitic rocks from unaltered zones may reflect the age of the later deformation as well as furnish a minimum age for steatization in the ultramafic zone. The post-steatization tonalite emplacement is an additional event which may indicate recurrent activity along the fractures within which the ultramafic rocks were emplaced.

Contact Rocks and Inclusions in the Ultramafic Zone

The contacts between the wallrocks and the ultramafic zone have been discussed in detail by Jahns (1967) with respect to the Roxbury district in Vermont. In the Ludlow area the so-called "blackwall" zone marking the contact consists of biotite schists on the outward margin grading inward to a chlorite-talc schist. Actinolite lenses are infrequent and usually occur in deformed bodies up to 12 cm thick. The actinolite crystals are usually undisturbed in the centers of the augen but are bent and broken towards the deformed margins of the lens (Fig. 34). The blackwall zone rocks show evidence of considerable post-reaction deformation, often resulting in disturbed blackwall sequences with juxtaposition of the various zones and occasional faulting of the blackwall rocks into the nearby talc schist zones. Rocks of the blackwall are usually less than 20 cm thick with considerable variations in thickness. It is common to observe contacts with a thin rind of chlorite-biotite schist only 2 to 4 cm thick. The uneven distribution of blackwall rocks probably reflects late deformation of the zone rather than initial conditions of formation. The complex faulting and folding in the zone complicates geochemical studies on elemental diffusion during the blackwall crystallization, since the present distribution of these rocks is not the original distribution. Furthermore,

Figure 34 – Typical podiform body of actinolite from the “blackwall” zone. The upper surface of the specimen is the contact with the biotite-chlorite schists, while the lower surface faces inward toward the talc zone. Actinolite needles are deformed into the foliated talc in the center of the specimen.

it is difficult to rule out additional reactions within the zone which may have occurred during later deformation.

Inclusions in the ultramafic zone are very common and fall into four main types: wall-rock fragments, amphibolite lenses, granitic masses, and chlorite-magnetite lenses.

Wall-rock inclusions are relatively rare and consist simply of blocks of schist or gneiss from the wall rocks which have been tectonically emplaced into the ultramafic zone. Most of these bodies are less than 10 meters long and are usually located a few meters from the contact zone itself, often forming a false hanging wall or footwall. These bodies are confined to the talc-carbonate schists in the ultramafic zone, and are usually enclosed in a thin sheet of typical blackwall rocks.

Amphibolite lenses have been described in the wallrocks where they are very common. In the ultramafic zone they are extremely rare, possibly due to complete alteration to chlorite schists. Nonetheless, the presence in the ultramafic zone suggests a possible basic parent rock for some of the chlorite-magnetite lenses, although a clear case where amphibolite masses appear to be partly altered to chlorite lenses has not yet been seen.

Granitic rocks have also been previously described and are mentioned here as inclusions because they are normally surrounded by thick chlorite-magnetite schist contact zones. It is possible that some of the chlorite-magnetite lenses are completely altered and dismembered granitic intrusions, representing reactions between the high Si,Al granitics and the high Mg, Fe ultramafics.

The chlorite-magnetite schists are by far the most abundant inclusions in the ultramafic zone and are found in all rock types within the zone, including the very core serpentinites (Fig. 35). It is likely that some of these bodies may have been formed from granitic rocks or more basic igneous intrusions. They typically range from 1 to 2 meters thick and up to 20 or 30 meters long (Fig. 35) and are generally oriented subparallel to the strike of the zone in the talc-carbonate schists and carbonate-talc rocks. Although not confined to the talc schists, the chlorite lenses are more frequent in these rocks than in the inner carbonate-talc rocks and serpentinites. The chlorite bodies are usually strongly foliated and are often affected by late, open folds with a strong crenulation cleavage as an axial plane structure. Magnetite octahedra are finely dispersed in grain sizes from a few mm to 1 cm within most chlorite schist masses. At locality "B" (Figs. 32 and 35) a number of chlorite schist bodies can be seen in the northeast sector of the serpentinite core. The easternmost body appears to be associated with a structural break in the core which trends north-northeast. This body contains a strong chlorite-rich layering defined by parallel mica foliations about .25 to .5 mm thick. This layering is cut by a vertical crenulation cleavage which strikes subparallel to the "structural break." A few meters northwest of this body is another chlorite-magnetite schist body which lies partly in the serpentinite core and partly in the surrounding foliated and strongly deformed talc schists. Figure 36 presents a southwest facing view of the portion of this body which lies in the serpentinite core. The remaining half of the body which lies in the talc schist zone is strongly deformed with the chlorite layering disposed in tight to isoclinal folds. The portion contained in the serpentinite core, however, is only weakly deformed and no folds are

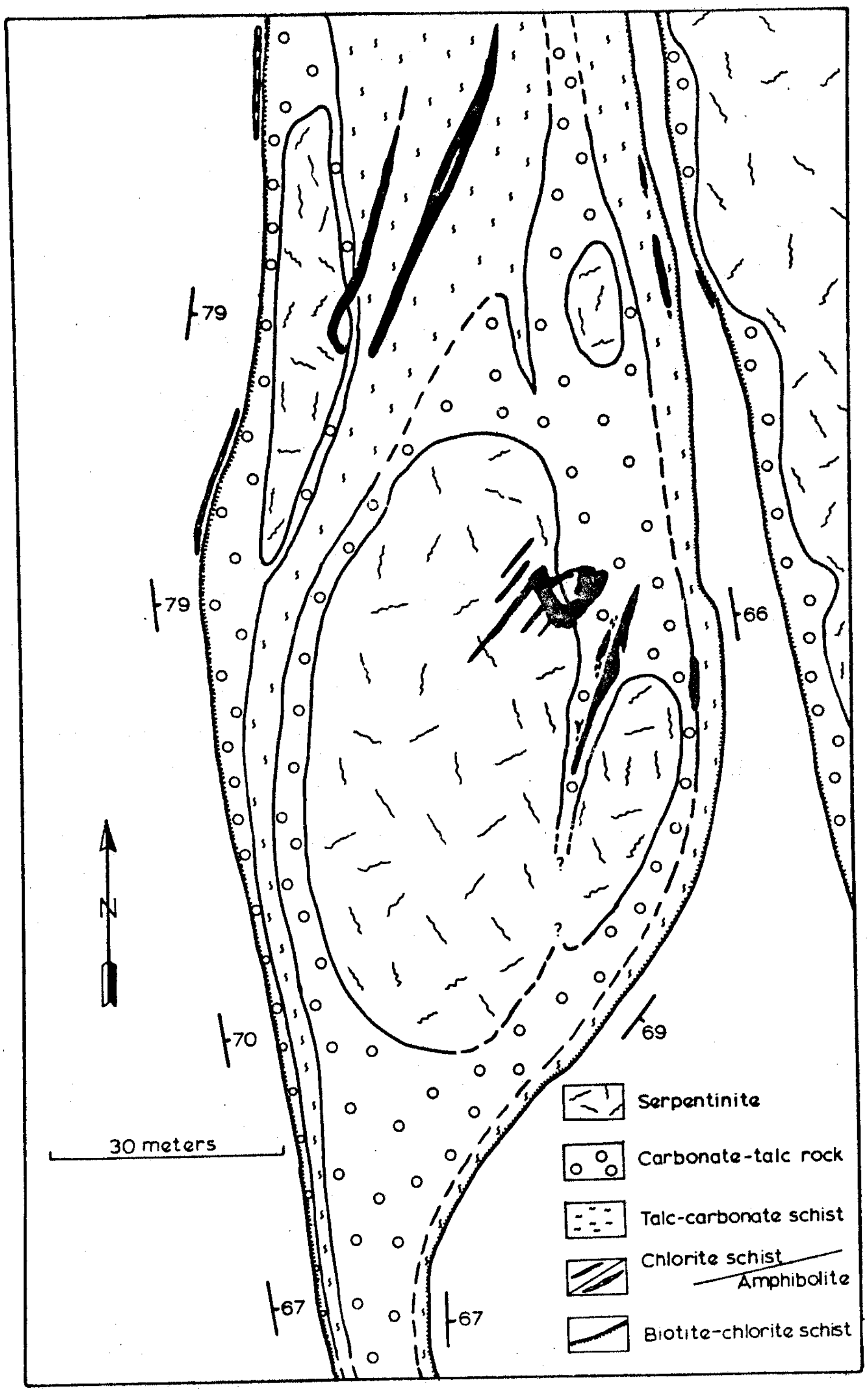


Figure 35 - Geologic map of locality "B" (Fig. 32), showing typical distribution of various rock types in the ultramafic zone.

Figure 36 – View of locality “B” (Fig. 32) facing SW towards the chlorite schist inclusions in the serpentinite core. The chlorite mass in view is the largest body in the center of Figure 34. The top of the hill is the center of the serpentinite core. The body is 10 meters across at the bottom of the photo.

Figure 37 – Details of the center "dike" in Figure 36. The white halo is a zone of talc alteration in the serpentinite body. In the center of the "dike" a chlorite-rich foliation runs vertically in the photo. In the edges of the "dike" a chlorite-talc layering runs horizontally in the photo. The talc zone is believed to represent an alteration zone in the serpentinite, possibly thermal. The black pen in the upper right is 14 cm long.

observed. As can be seen in Figures 35, 36, and 37, this body exhibits an apparently intrusive relationship to the serpentinite core and consists of a series of "dike-like" masses of parallel orientation. The large "halos" of talc around the dikes (Figs. 36 and 37) represent some types of reaction zone and could represent thermal alteration of the serpentinite by a high temperature intrusion, although strong evidence for this is lacking.

The serpentinites in the core will be discussed in the next section. The dikes discussed here do not contain the rather strong patterns of deformation seen in the serpentinites, and probably represent post-tectonic intrusions with respect to the deformations to be discussed in the following section.

Geology of the Ultramafic Rocks

The ultramafic zone usually consists of a core of serpentinite, with up to 50% primary pyroxenes present, surrounded by a rim of foliated antigorite rock with little or no primary minerals present. This foliated serpentinite sheath is commonly enclosed in carbonate-talc rock, which is usually followed by an outer zone of talc-carbonate schists. The relations between each zone have been described in detail by Jahns (1967) and the illustrations presented by Jahns are essentially similar to the geology of the Ludlow area ultramafics.

Table IX is a compilation of modes for the serpentine-bearing rocks in the ultramafic zone and is arranged according to the relative distance of each sample from the center of the serpentinite core. The first three samples are from the actual center of the cores and the remaining samples

TABLE IX - MODAL ANALYSES FOR ULTRAMAFIC ROCKS

GENERAL LOCATION *	SAMPLE NUMBER	pseudomorphs Opx.	Cpx.	Serpentine	Carbonate	Opaques	Talc
Core zone	U-3-74	48	0	47	0	5	0
	U-2-74	32	5	55	0	8	0
	U-1-74	29	0	57	8	8	0
	32-S, 129'	35	tr.	58	1	6	0
	32-S, 130'	19	0	72	0	9	0
	18-R, 230'	0	0	64	35	1	0
	18-R, 237'	0	0	70	30	0	0
	18-R, 247'	0	0	79	17	0	4
Talc Zone	18-R, 257'	0	0	43	40	0	17
	18-R, 260'	0	0	49	27	0	24

* arranged in increasing distance away from the core zone.

no evidence for the effects of late folding of the zone on this particular scale, although the distribution of the ultramafic zone on a regional scale is often controlled by late folding; for example, the distribution of the zone around the Chester Dome to the east of the study area. It is considered that the aspect ratio of each serpentinite body is approximately the same in the vertical and horizontal planes. Thus, the bodies are elongate, podiform structures in three-dimensional view. The partial sections given by Jahns (1967) suggest this type of structure for small scale examples. On a slightly larger scale, the relationship also holds as can be seen in the illustrations of the ultramafic zone rocks at locality "A" (Figs. 38 and 39).

In a number of areas large elongate bodies of Cram Hill phyllite are situated in talc-carbonate zones between adjacent serpentinite masses. The blocks contain Style Group 1 and 2 elements on a mesoscopic scale, but the nature of the terminations of individual blocks of the phyllite is unclear (refer to Fig. 32). In some cases the blocks end abruptly in the talc zone, which in turn terminates in a serpentinite enclosure. In other examples the talc zone bordering the phyllite block contains abundant chlorite-magnetite schist stringers in zones hundreds of meters long, often occupying obvious dislocation zones in the talc body. These relationships may suggest that the blocks are faulted slices included in the ultramafic zone during emplacement, but the outcrop is insufficient to observe whether or not the terminations of the blocks are fold hinges or fault-bounded surfaces.

(B) Mesoscopic and Microscopic Structures

Examination of a number of core serpentinites has revealed a

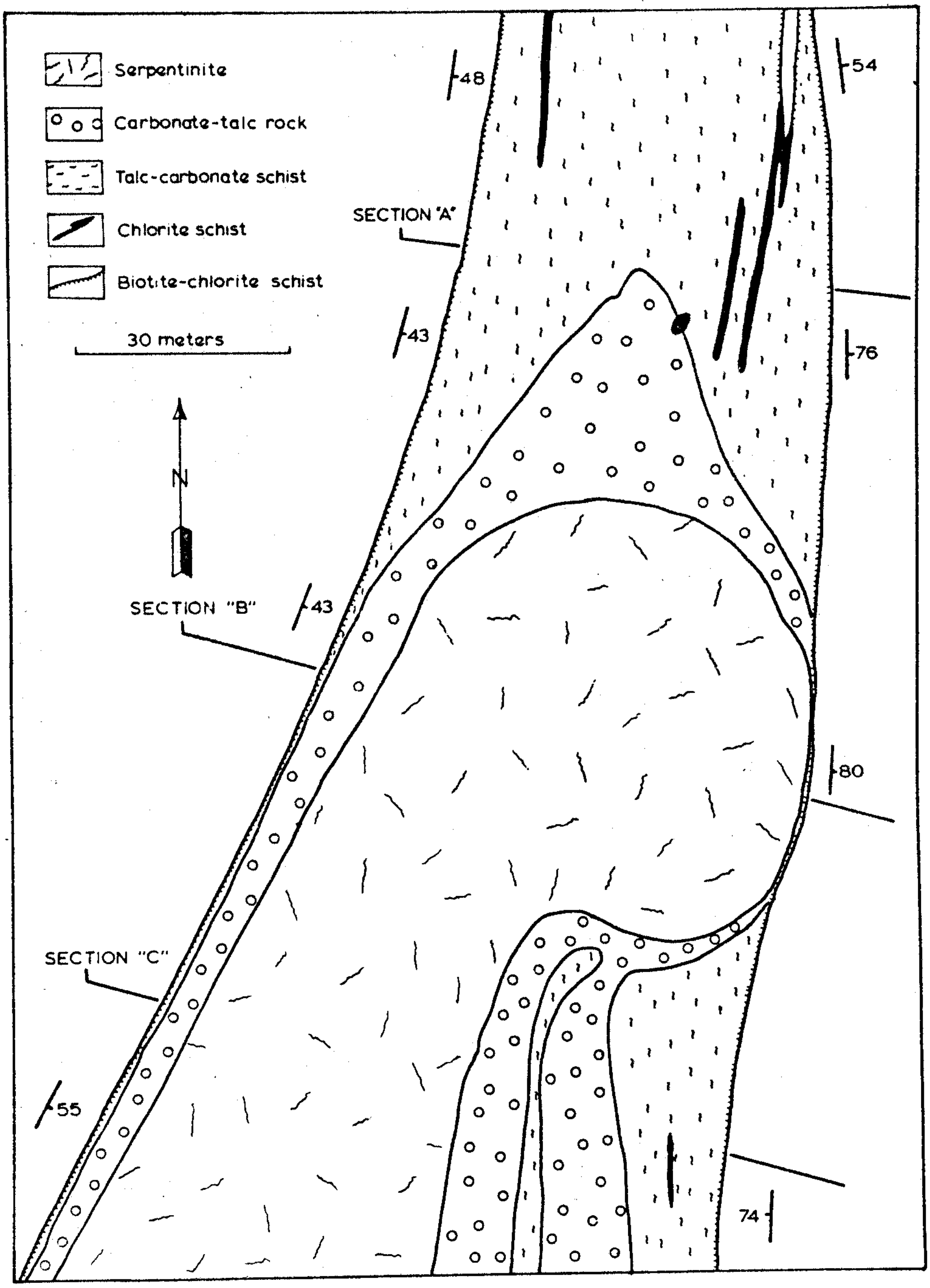


Figure 38 - Geologic map of locality "A" (Figure 32) showing general relationships of various rock types in the ultramafic zone. Sections marked are presented in Figure 39.

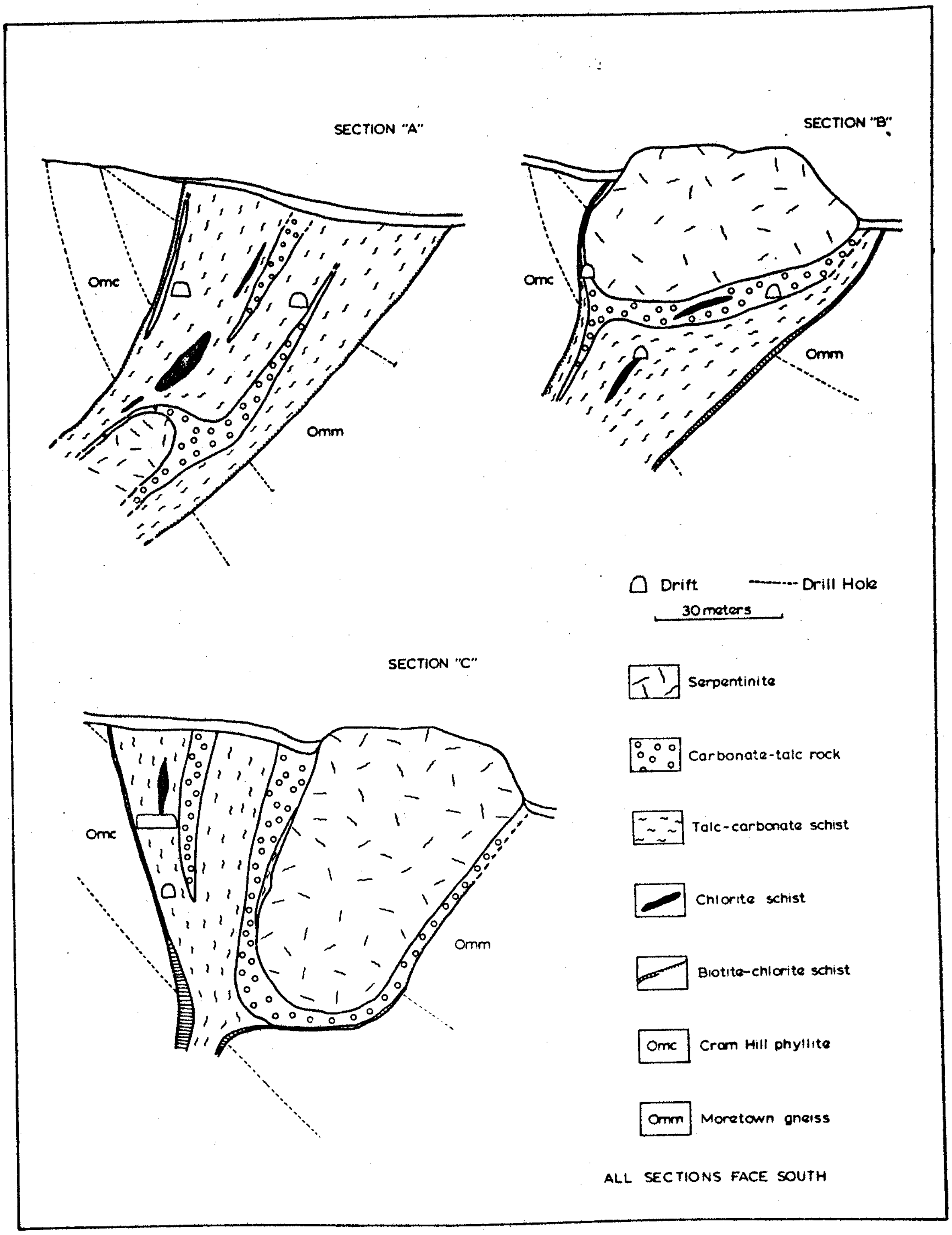


Figure 39 - Structure sections for the geologic map (Fig. 38) of locality "A."

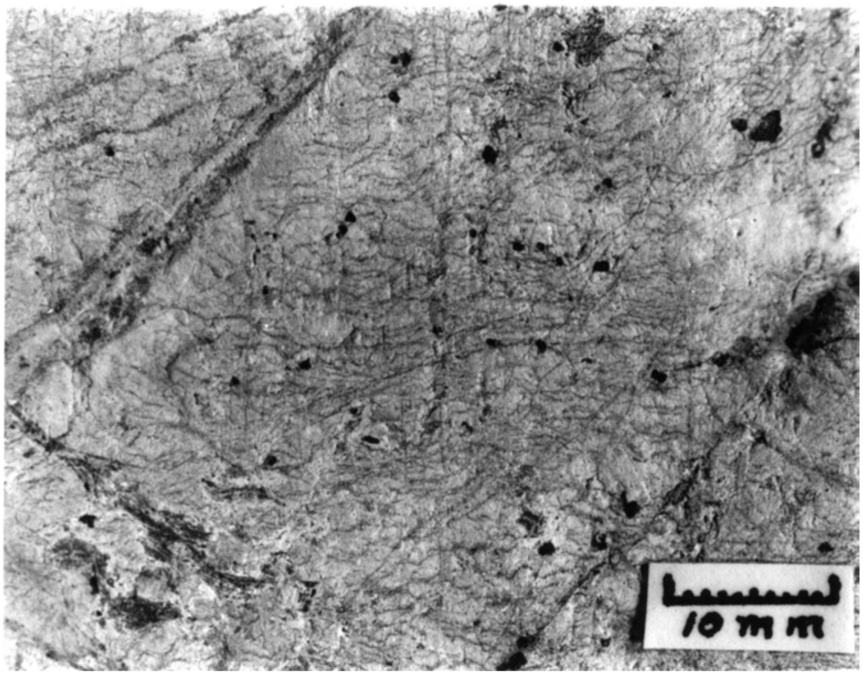


Figure 40 – Typical appearance of mesostructures in the ultramafic core rocks. Most of the early structures are faint but discernable lines; the darker lines are talc-rich late discolorations in the core rocks and appear at about 45º off vertical in the photograph. The vertical lines are S_2 foliations defined by a serpentine mineralization along axial surfaces to early folds, which are visible as weak crenulations in the horizontal S_1 layering. S_1 layering consists of alternating serpentine and altered pyroxene domains and can be best observed in Figures 41 and 42.

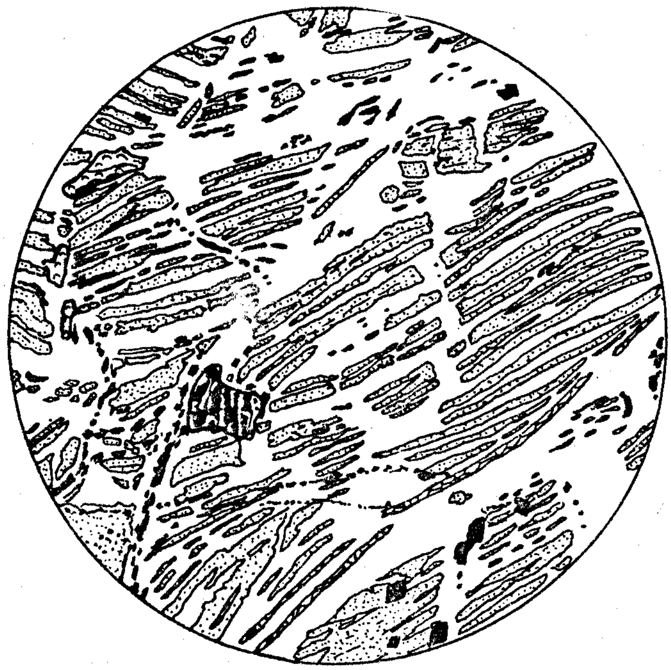


Figure 41 - Typical microscopic appearance of S_1 and S_2 in the ultramafic core rocks, in this case probably a serpentinized harzburgite. S_1 is easily recognized as a strong layering running from lower left to upper right and consists of alternating domains of serpentine minerals (clear) and altered orthopyroxene cleavage fragments (stipple). Slightly below and left of center is a single clinopyroxene grain heavily occluded by opaque minerals. Long trails of olivine crystals are visible in the lower right corner above the thick band of serpentine. S_2 is weakly developed here but is vertically oriented in the hand specimen. Field of view is 16 mm. This rock is number 32-71-S, 129' and a model analysis is given in Table IX.

series of structural elements which may be used to define the internal structure of the core rocks. The earliest mesoscopic element is a foliation (S_1) which consists of serpentine minerals that cut across altered orthopyroxenes and clinopyroxenes. In some cases it appears that S_1 has devloped along cleavage planes in pyroxene grains with an initial preferred orientation (for example in Fig. 41), but in other cases the extinction directions of the alteration products are at high angles to S_1 (for example in Fig. 42 S_1 is nearly horizontal but the grain extinction is along the vertical subgrain boundaries). The subgrain boundaries shown in Figure 42 are of uncertain origin but seem to be cut by S_1. These boundaries form a foliation seen only on a microscopic scale which is referred to as S_0 since it is probably earlier then, or at best synchronous, with S_1. There is not enough evidence at this time to support one alternative over the other.

S_1 layering is disposed in small kink-like early folds on a scale of millimeters, with the kink band boundaries often lying along the original grain boundaries of the pyroxene pseudomorphs. A new serpentine foliation (S_2) usually occupies these boundaries and is often seen as an axial surface foliation to the early folds in S_1 (refer to Fig. 40). This new foliation is not mineralogically distinct from S_1, but while S_1 is folded on this scale, S_2 is essentially undeformed. In the field one can usually observe S_1, S_2 and the early open folds as well. The foliations are very fine and the serpentinite does not tend to cleave well along these surfaces -- thus it is quite difficult to measure orientations of the foliations. The intersections of foliations with horizontal surfaces can be measured with ease, however, and form surface mapping may therefore be performed. Figure 43 illustrates a form surface map for the serpentinite

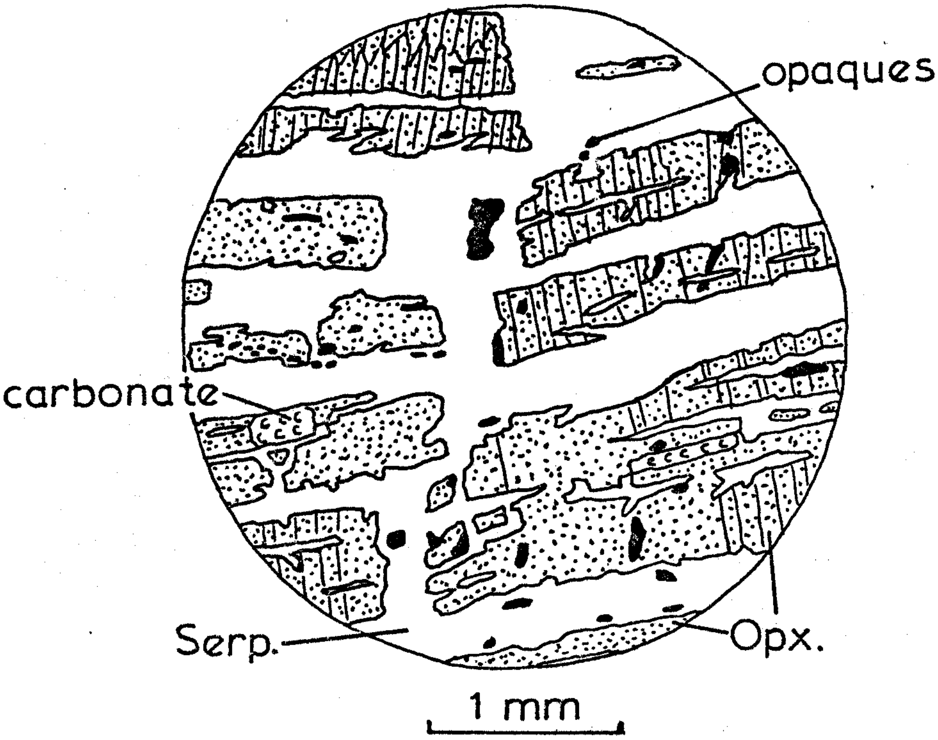


Figure 42 - Detailed view of microstructures in a typical serpentinized harzburgite from the ultramafic core zone. S_1 foliation consists of the nearly horizontal serpentine mineral layers separating altered orthopyroxene cleavage fragments. The nearly vertical dislocation corresponds to a typical axial plane foliation to early "kink-like" folds which deform S_1. The vertical lines inside various areas of altered pyroxene grains are the subgrain boundaries which define S_0.

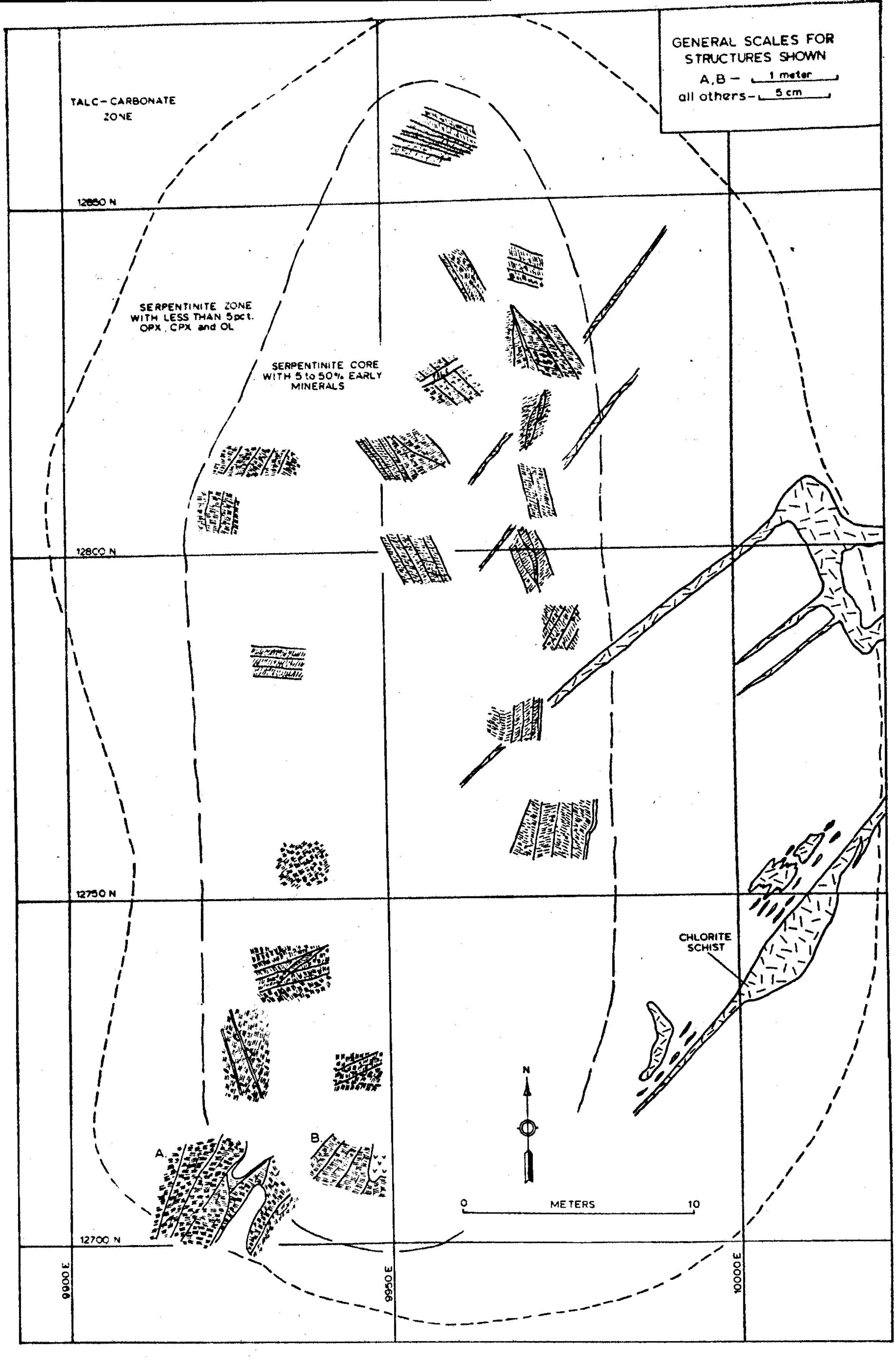


Figure 43 - Mesoscopic structures in the serpentinite at locality "B" (Fig. 32). Solid lines depict S_2 layering, dashes indicate S_1 layering.

core at locality "B" (Fig. 32 and 35). The map indicates the orientation of S_1 and S_2 at various localities in the core as well as the distribution of chlorite schist zones. S_2 is represented by solid black lines and S_1 is depicted as parallel dashes. Sketch "A" in the lower left corner of Figure 43 is actually located in the central core boundary, but is illustrated off the zone for easier observation.

This mapping has shown that S_2 is redistributed by late open folds on a scale of 10 meters in wavelength, which are not usually observed on an outcrop scale. Figure 44 is an interpretation of the distribution of S_1 layering in the core. The folds depicted are early folds in S_1. Figure 45 depicts an interpretation of the distribution of S_2 in the core. The large open folds are the late folds which deform S_1, S_2 and the early folds as well.

It is important to note that the chlorite schists transect all the observed structural elements in the core and probably represent late chloritized intrusions. The fact that the chlorite schists are in turn deformed where they extend into the talc zones indicates an even later deformational episode which had very little effect on the core structures, but which seems to have affected the talc zones and the wallrocks as well. While the case is by no means concluded, this deformational episode appears to be the last event observed in the wallrocks -- that is, the deformation associated with S_3 in the Moretown and S_2 in the Cram Hill. This argument has a certain appeal since on a regional scale the ultramafic zone is deformed by large folds, such as the Chester Dome which correlate well with the late folds on an outcrop scale (B. W. Nisbet, 1975 pers. comm.). Beyond this point little can be said except that there are more structural elements in the

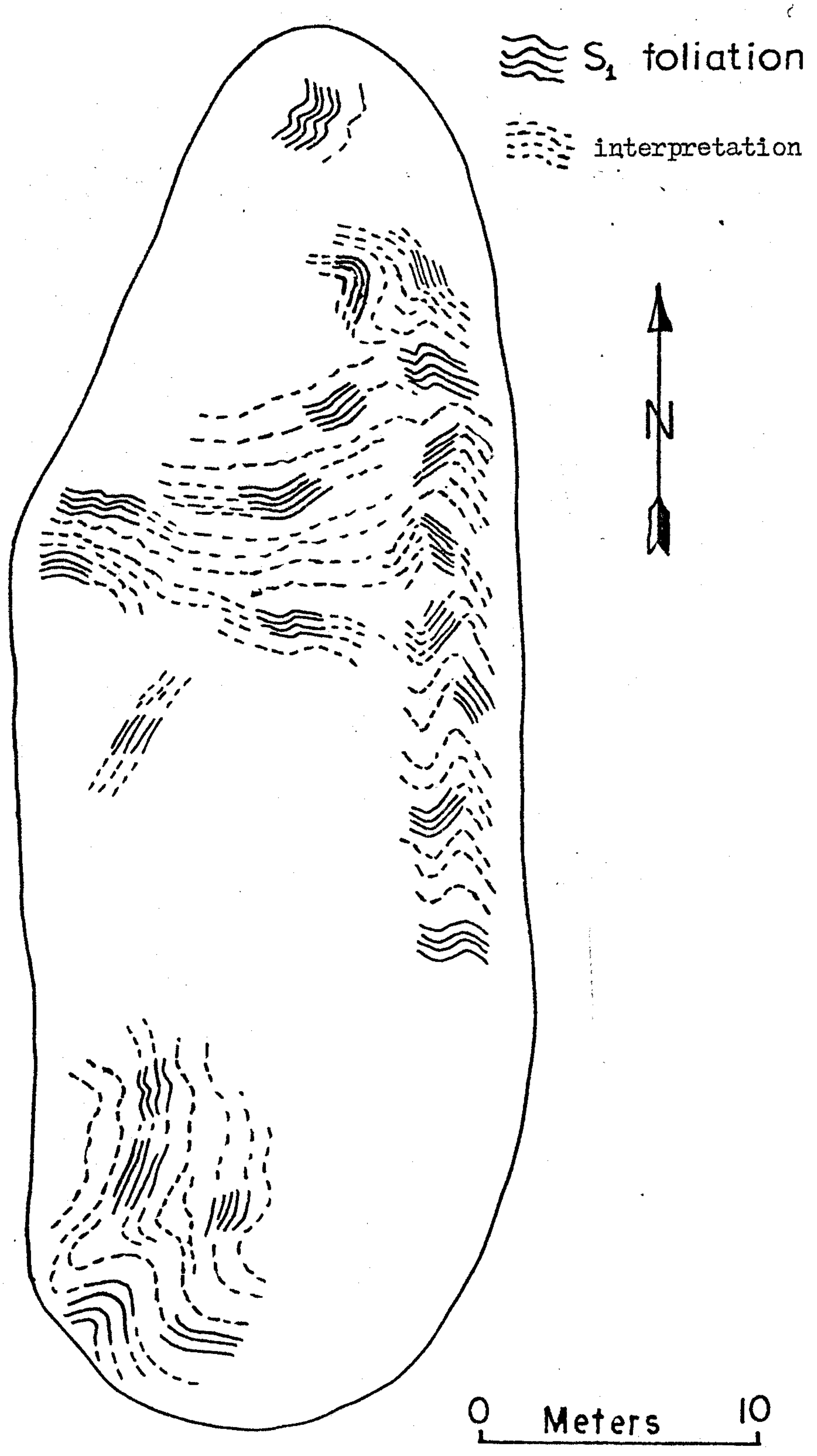
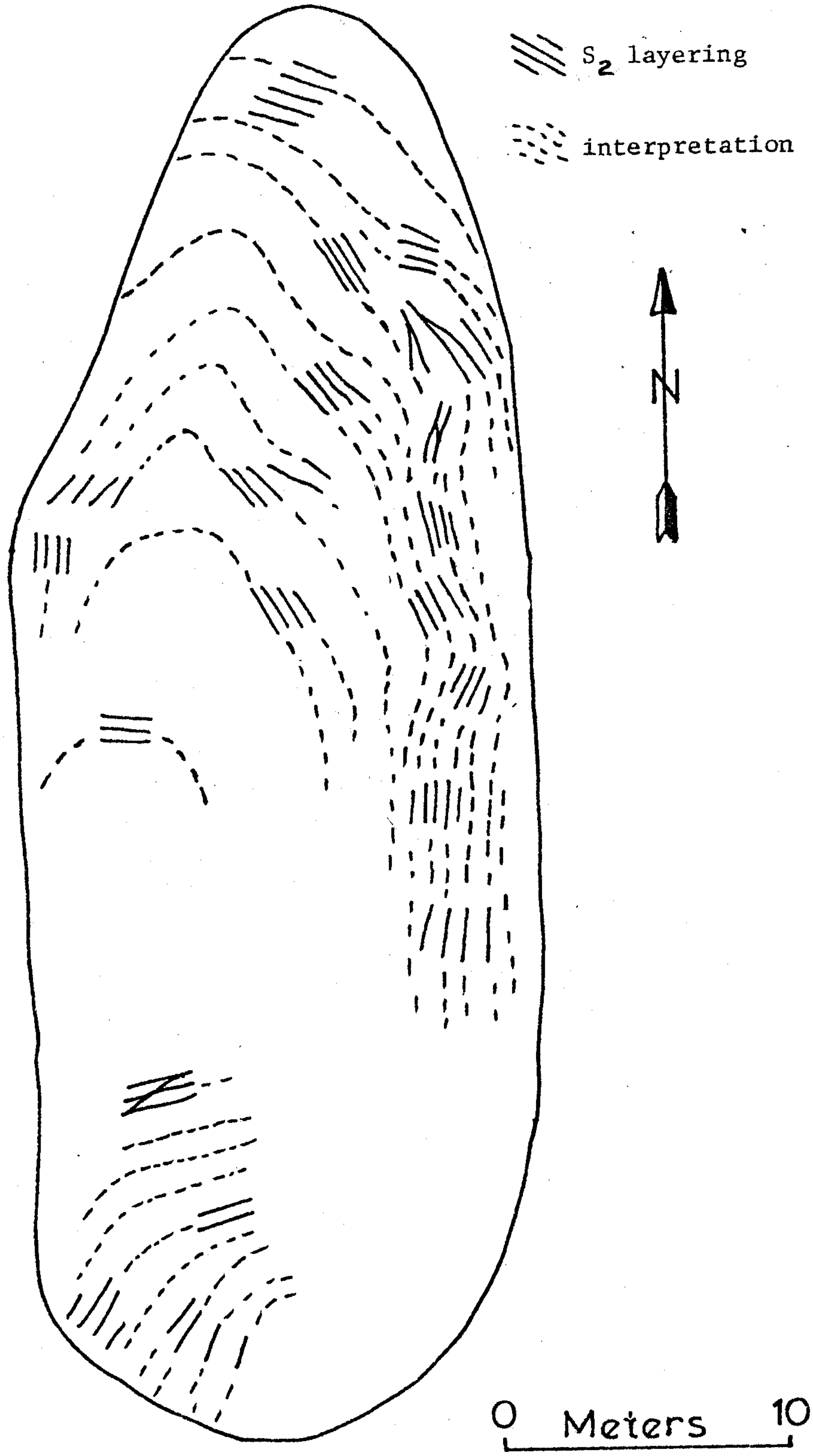


Figure 44 - Possible interpretation of distribution of S_1 in selected areas of Figure 43. Folds are early folds with axial plane foliation S_2 (not depicted). Interpretation is not possible in all areas.



'Figure 45 - Possible interpretation of the distribution of S_2 layering in Figure 43. Folds are late open folds in S_1 and S_2 and are not observed on a mesoscopic scale.

TABLE X - STRUCTURAL ELEMENTS FOR THE ULTRAMAFIC BODY LOCATED AT LOCALITY "A"

	FOLIATIONS	FOLDS
?	none observed	Late macroscopic open folds of S_2
S_2	Mesoscopic serpentine-rich axial plane foliation.	Early kink-like open mesoscopic folds of S_1 layering.
S_1	Mesoscopic serpentine-rich layering cutting altered pyroxenes. Not observed as an axial plane foliation.	none observed.
S_0	Foliation consists of oriented subgrains of unknown pyroxene alteration product.	Early alteration of pyroxene by crystallization of new grains along kink band boundaries? Other explanations for S_0 are feasable.

ultramafics than in the wallrocks and this might indicate that some of the core structures are emplacement related. The author is prejudiced toward the idea that none of the core structures reflect the regional deformation patterns seen in the surrounding wallrocks. In a purely imaginative sense one might expect the core rocks to be protected in some way by the surrounding talc-carbonate sheath, within which most of the regional deformation would occur. Thus it is possible that the structural elements in the core are syn-emplacement or pre-emplacement structures which have been preserved by the surrounding talc sheath.

Table X is a summary of the structural elements which have been observed in the core rocks. The structural history can be outlined as follows:

(1) Early deformation of fresh ultramafic rocks (harzburgites and lherzolites) with alteration of orthopyroxenes to an unknown phase (amphibole?) and low-angle subgrain boundaries forming at high angles to cleavage traces and extinction positions in pyroxene pseudomorphs. A new extinction position is produced in many grains, defined by a parallel orientation of subgrains (S_0) throughout the pseudomorph which retains the old cleavages traces.

(2) S_1 serpentine foliation is formed parallel to the old cleavage traces in the pyroxene pseudomorphs. Primary olivines are altered and the pseudomorphs are tectonized. The final rock structure is a new layering of alternating serpentine and altered pyroxene layers, with old pyroxene grain boundaries still visible.

(3) S_1 layering is deformed by early, kink-like folds visible in hand specimens. A new foliation (S_2) is formed by serpentinization

along axial plane surfaces to the folds.

(4) S_0, S_1 and S_2 are finally deformed by late, open folds on a large scale -- not usually visible in outcrop. No foliations have been correlated with this folding.

(5) A series of minor, apparently random talc-rich dislocations transect all early features on all scales. The zones are usually 1-5 mm in width.

CHAPTER IV

DEFORMATIONAL HISTORY OF THE STUDY AREA AND REGIONAL STRUCTURAL OBSERVATIONS

Introduction

While the assignment of structures to Style Groups within the homogeneous lithologic assemblages is based on observable features and is a relatively non-speculative step in structural analysis, the correlation of Style Groups throughout the various rock units in an area, and the resulting assignment of structures to Generations, is a much more speculative step. Here the orientations of structural elements may play an important part and other geologic relationships are also used. In this area, the geometric relationships, and inferred age relationship, of the ultramafic rocks to the surrounding metasediments are especially important in the comparison of structures between the various rock types.

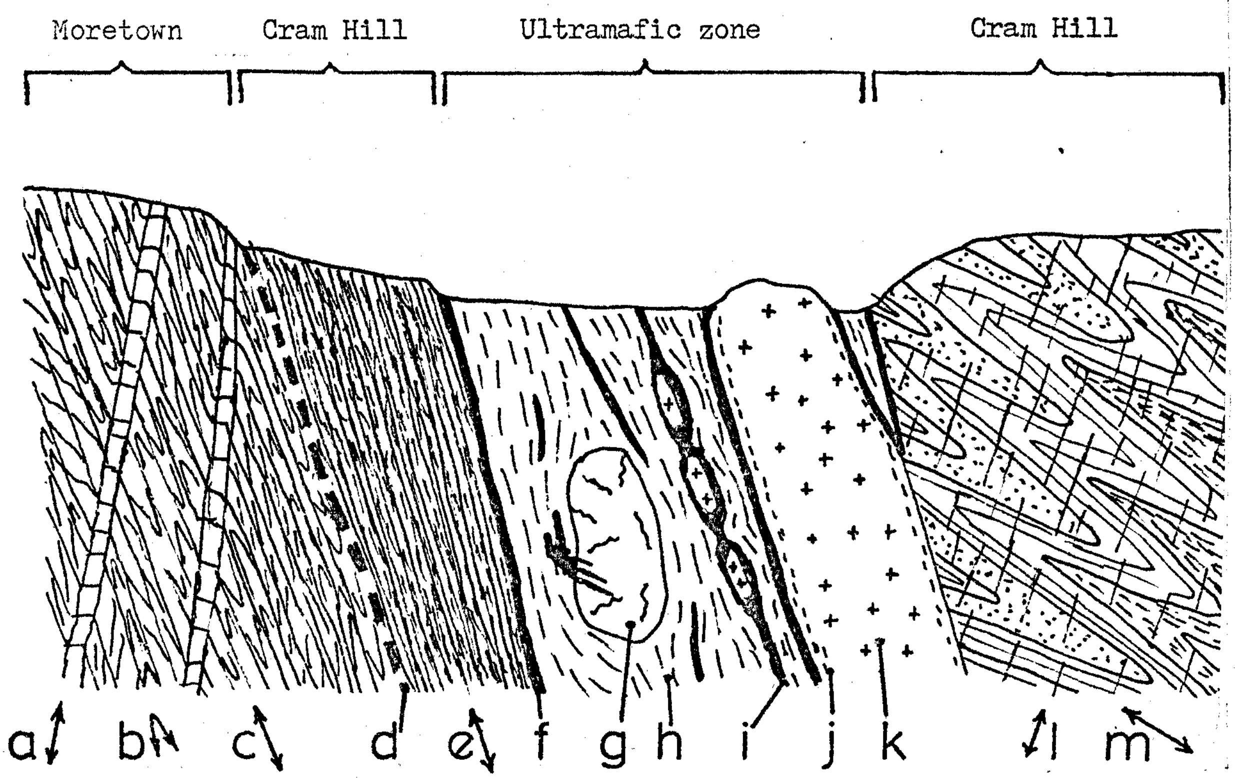
Correlation of Style Groups Between the Moretown and Cram Hill

Table XI is a compilation of style group elements in the Moretown and Cram Hill rocks. The principal arguments for the correlations are based upon the relationship between the ultramafic zone and the surrounding wall-rock structures. The structures listed under the B_2 generation are most closely associated with the emplacement of the ultramafics. Style Group M-2 elements in the Moretown exhibit structural concordancy with the ultramafic zone; for example, S_2 in the Moretown is subparallel to the strike of the ultramafic zone throughout the region and the possibility of some structural control of the zone by B_2 structures may exist.

Within the ultramafic zone itself the only style group elements observed in the talc-carbonate rocks are those of the last event -- in some

GENERATION		B_3	GRANITIC INTRUSION	ULTRAMAFIC EMPLACEMENT	B_2	B_1
CRAM HILL PHYLLITES	STYLE GROUP	C-2			C-1	(undeformed)
	FOLD STYLES	Open-tight			Isoclinal	
	FOLIATIONS	S_2			S_1	
	LINEATIONS	L_2			L_1	
	PORPHYROBLASTS	----			Garnet	
MORETOWN GNEISS	STYLE GROUP	M-3			M-2	?
	FOLD STYLES	Open			Isoclinal	?
	FOLIATIONS	S_3			S_2	S_1
	LINEATIONS	L_3			L_2	--
	PORPHYROBLASTS	----			----	Garnet

TABLE XI -- Correlation chart for various style groups in the Cram Hill phyllites and the Moretown gneisses. Note that the Cram Hill is deformed by only two generations, while the Moretown is deformed by an additional B_1 early deformation. B_3 corresponds to the regional "dome" folding associated with the Chester Dome and other such antiformal structures.



·Figure 46 - Generalized relationships of various sturctural elements to the ultramafic zone: (a) late S_3 foliation in Moretown gneiss, (b) early S_1 layering in the Moretown gneiss, (c) S_2 layering in the Moretown as axial plane foliation to isoclinal folds, (d) fault contact between Moretown gneiss and Cram Hill phyllite with 6" vein quartz, (e) strong B_2 foliation in Cram Hill phyllite equivalent to S_2 in the Moretown, (f) blackwall contact of chlorite and biotite schists, (g) serpentinite core, (h) talc-carbonate rocks, (i) chlorite schist contact zone around granitic boudins, (j) weakly foliated tonalite near contacts, (k) massive, unfoliated tonalite body, (1) S_2 crenulation cleavage or trace of crenulation axial plane to B_3 folds, (m) B_2 folds in the Cram Hill quartzites and phyllites (style group C-1).

cases it is possible to trace small-scale Style Group C-2 folds in the Cram Hill into the ultramafic zone. Thus, only one deformational event has been recognized in the zone after steatization, and the Style Groups in the metasediments which overprint B_2 structures are classified as the B_3 generation. On a more regional scale, B_3 structures seem to deform the ultramafics and may dispose the zone in large regional structures such as the Chester Dome. Large-scale B_2 structures, however, have not been observed in the relatively simple map pattern of the ultramafic zone throughout the State of Vermont.

Deformational History of the Ultramafic and Surrounding Rocks

The earliest deformation (B_1) is seen only in the Moretown gneiss on the western boundary of the ultramafic zone, and consists of garnet grade metamorphism accompanied by strong deformation and the development of a pronounced early secondary layering (S_1) not reported by previous workers. Textural evidence from garnet microstructures suggests garnet growth late in the development of S_1.

B_2 deformation can be observed in both the Moretown gneiss and Cram Hill phyllites. In the Moretown, B_2 structures overprint B_1 structures, but in the Cram Hill no earlier deformational structures have been seen. The ultramafic zone is generally concordant to the strike of S_2 layering in the rocks along the western boundary as are the apparently fault controlled contacts between the various metasedimentary rocks in the central Vermont area. It seems likely that the emplacement of the ultramafics occurred late in the B_2 deformation. Subsequent to ultramafic intrusion, small granitic bodies were intruded into the ultramafic zone, generally parallel to the strike of the zone. These stocks are weakly foliated along the margins and are boudinaged within the talc-carbonate rocks where chlorite reaction rims are present. The granitic rocks appear to have been weakly deformed by B_3 deformation but evidence is restricted by poor field exposure. The boudinaged dikes in the talc zone may have been produced by B_3 deformation of the granitic rocks.

Radiometric dating of the unfoliated tonalite granites within the larger stocks may resolve a number of problems in the regional geology, including the placement of upper age limits for B_2 folding and a minimum emplacement age for the ultramafic rocks; however, more detailed information concerning field relationships is needed.

Additional checks on the ages of various deformational periods have been proposed (Rickard, 1965) by dating recrystallized micas taken from the foliations associated with each style group. The writer has seen no absolute proof that the mica-rich domains associated with any of the secondary layering have formed by recrystallization of layer silicates in the domains, rather than by selective removal of quartz from these zones of high strain. Although the micas must have crystallized at some point in time, we know little about where the crystallization occurred with respect to the foliations.

B_3 deformation affected the Moretown, Cram Hill, and ultramafic zone rocks. The effects of B_3 are weaker in the Moretown gneisses than in the Cram Hill phyllites, possibly due to the respective characters of the rocks. In the Moretown the effects are limited to rare open folds and weak S_3 layering (refer to Fig. 7). B_3 structures are strongly developed in the less competent Cram Hill phyllites with ubiquitous crenulations and crenulation cleavage and fairly common mesoscopic folds. In the ultramafic zone only the talc-carbonate rocks and chlorite schist inclusions were affected by B_3 folding with occasional crenulation cleavage as an axial plane structure to open folds. Although the serpentinites are characterized by a polyphase deformational history, they were unaffected by B_3 folding on a mesoscopic scale, possibly due to the ductility contrast between these rocks and the surrounding talc-carbonate schists and metasediments. However, the ultramafic zone is folded on a regional scale by large B_3 folds, such as the Chester Dome.

Orientation Diagrams for Various Structural Elements

Figure 47 represents the attitude of B_2 folds in the Moretown gneisses and is compiled from 63 L_2 lineations and 42 S_2 foliations dispersed uniformly throughout the area. Measurements from subarea "B" are excluded from Figure 47 since B_2 is strongly overprinted by B_3 in the subarea. The attitudes of S_2 in the area are shown in Figure 48 where poles to S_2 are contoured with 2%, 5%, 10%, and 20% contours. The restricted scatter and strong point maximums in the L_2 and S_2 plots indicates minimal redistribution of B_2 structures by B_3 deformation in the Moretown gneisses outside area "B." The great circle to the poles to S_2 layering is employed as an approximation of the axial surface for B_2 folding in the area, while the L_2 intersection lineations are used to approximate B_2 fold axes in the area. B_2 folds are not directly measureable in the Moretown because of the tendency of S_2 to obliterate early structures. L_2 plots on a strong point maximum at 41 degrees to N26E and S_2 strikes due north and dips 63E as determined from the maxima plotted in Figure 48.

Figure 49 illustrates the effects of B_3 folding in subarea "B" in the

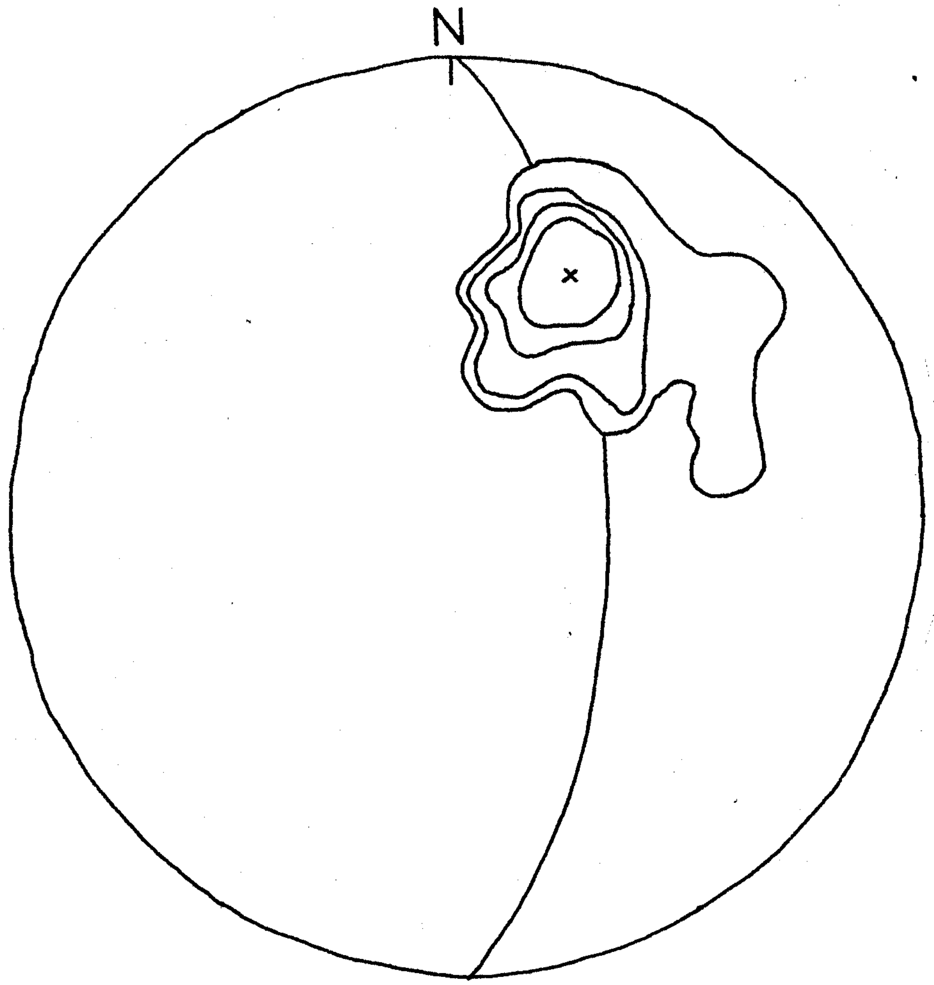


Figure 47 - Equal-area projection of structural elements associated with B_2 deformation in the Moretown gneiss. Sixty-three recordings of L_2 are contoured at 2, 5, 10, and 20 percent. The great circle is constructed from 42 poles to S_2 in the Moretown as in Figure 47.

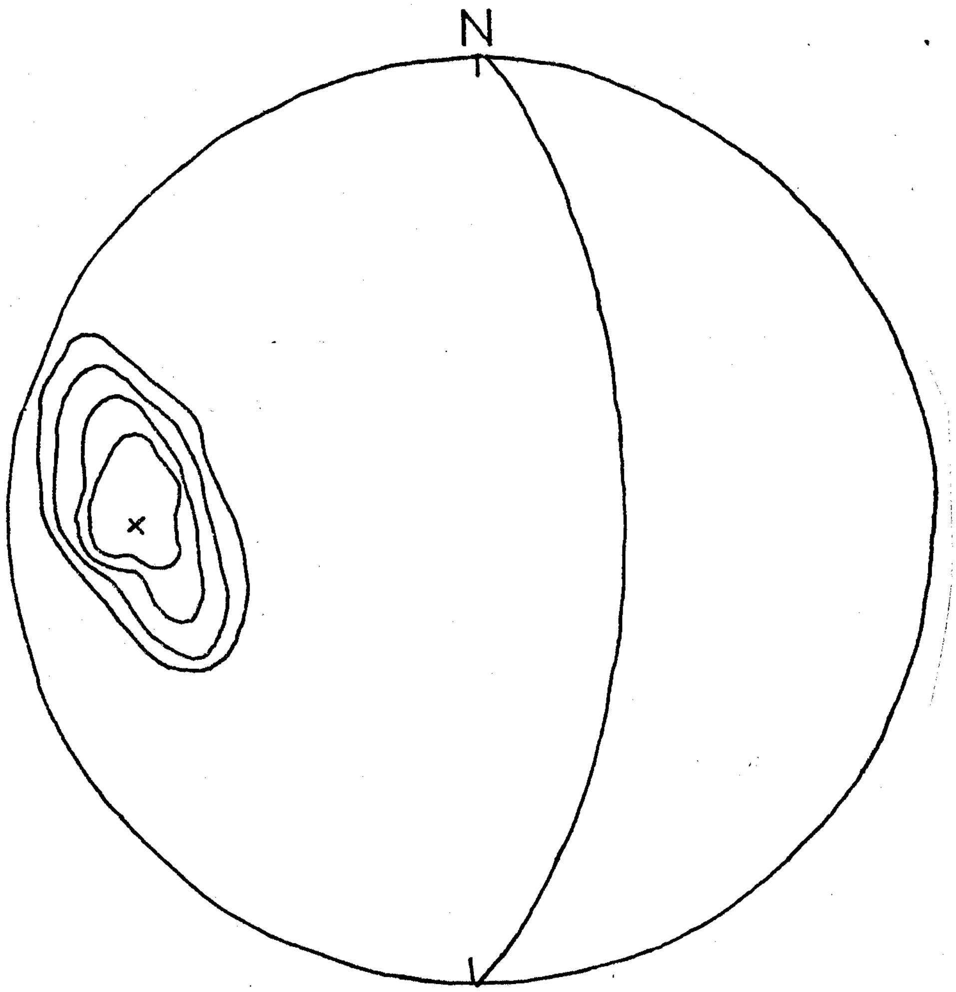


Figure 48 - Poles to S_2 in the Moretown, taken throughout the area excluding subarea "B." Fourty-two poles are contoured to 2, 5, 10 and 20 percent. Great circle is constructed from point maximum.

Moretown. B_3 folds in S_2 layering are directly measureable and have vertical axial planes striking N35E and axes plunging 10 to 20 degrees to N35E.

In the Cram Hill, B_3 folds are common and directly measureable with axial planes striking N16E (Fig. 50) and dipping approximately 80 degrees to east with fold axes plunging about a maximum at 25 degrees to N24E.

B_2 folds in the Moretown generally plunge more steeply than B_3 folds in the Moretown and Cram Hill, but B_2 and B_3 folds may often be coaxial and nearly coplanar when comparisons are made between the generations from differing rock types. This point clearly demonstrates the ineffectiveness of grouping fold types on the basis of orientation alone during analysis of the structure of adjacent rock units of differing lithologies.

Implications for Regional Structure

Structural work in Vermont has been limited largely to the recognition of "early" and "late" folds or associated mesostructures by previous workers, with relatively few workers attempting structural analyses as outlined in Turner and Weiss (1963) (Woodland, 1965, is an exception). Considerably more advanced work has been carried out in the Sutton Anticline and Stokes Range areas in southern Quebec throughout the last two decades; this work is briefly summarized in Beland (1967). Farther north in the Gaspé Peninsula additional work has been done by workers attempting less detailed studies over very large area (Sikander and Fyson, 1969; Carrara and Fyson, 1973, for example). Much of the original work has been revised in the Gaspe area, by reinterpretation of previous data in the light of new arguments based on stratigraphic principles. For example, Carrara and Fyson (1973) reinterpreted F_2 folds in the Matane-Matapedia area as Taconic in age (thus producing two phases of Taconic deformation) although the original

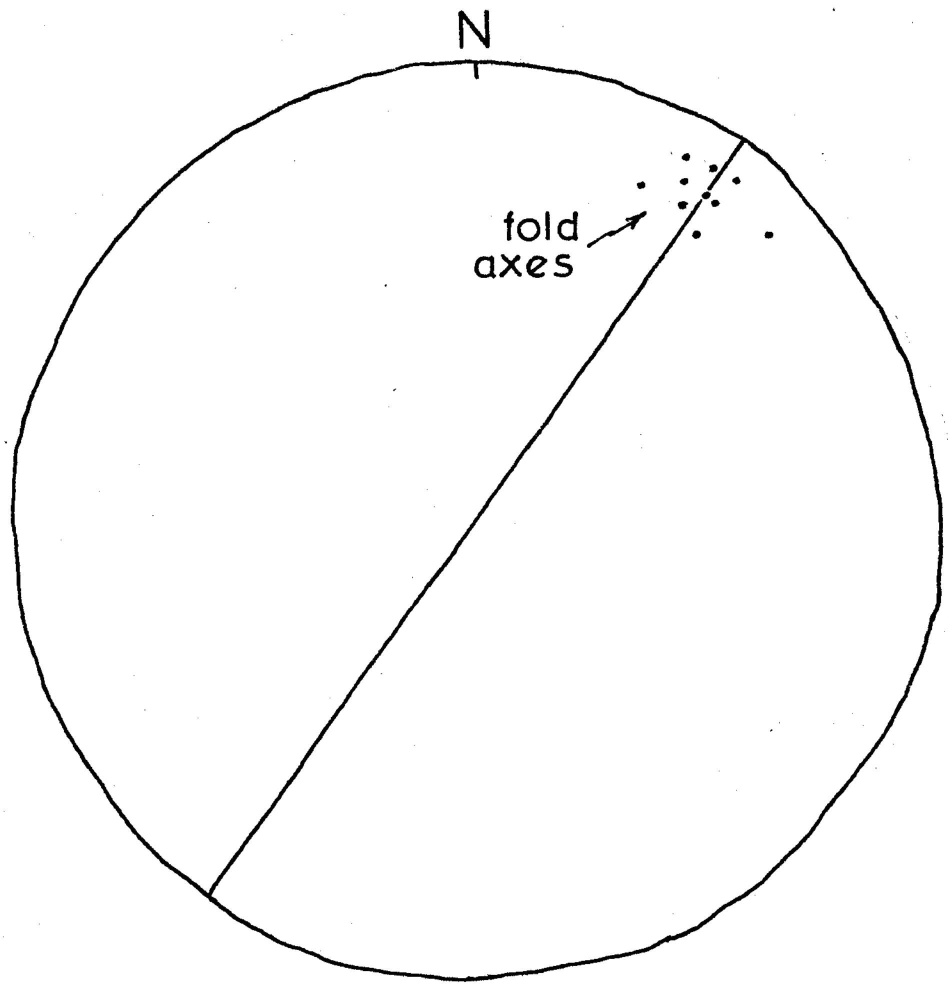


Figure 49 - Equal area projection of structural elements associated with B_3 deformation in the Moretown from sub-area "B." Ten mesoscopic Group M-3 folds are plotted. Great circle is plotted for fold axial planes.

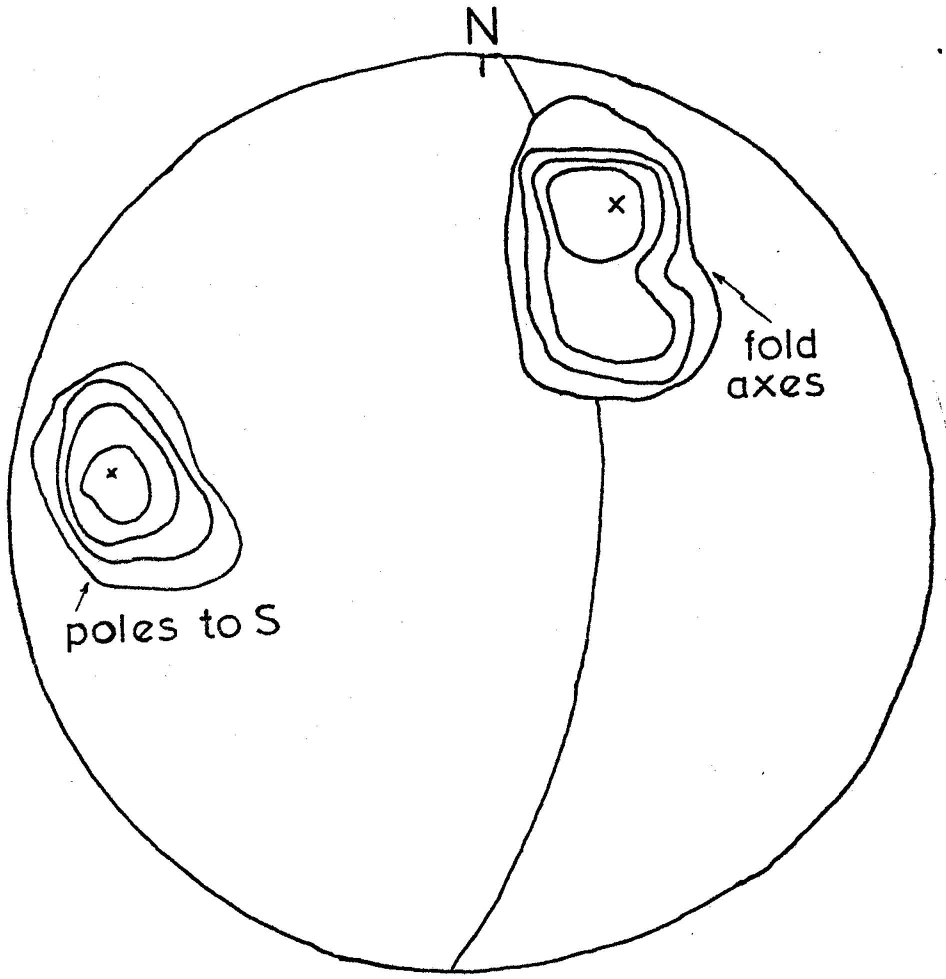


Figure 50 - Equal area projection of structural elements associated with B_3 deformation in the Cram Hill phyllites. Fifty-one Style Group C-2 fold axes are contoured to 2, 5, 10 and 20 percent. The great circle is plotted from the poles to S maximum.

work performed by Sikander and Fyson (1969) suggested an Acadian age on the basis of structural relationships between fold generations in Cambro-Ordovician and Siluro-Devonian rocks. This reinterpretation was carried out on the assumption that unconformities between Ordovician and Silurian rocks seem to cut the F_2 folds in the Ordovician rocks. The writers admit that the evidence is tenuous (the actual contacts are not exposed) but do not consider the possibility that the contact may be tectonic. Thus, it would seem that stratigraphic notions easily wrest control of the deformational history framework from structural analyses even in the Canadian Appalachians, regardless of the strength of the arguments.

Within the Sutton-Green Mountain anticlinorium DeRomer (1961) and Rickard (1965) classify all folds as Taconic, even though two generations were recognized by both workers. An additional earlier generation was postulated by Rickard (1965) based upon generalized relationships between nappe structures in the Taconics and the area investigated by Rickard. The consensus of opinion in the Sutton area seems to be that the Cambro-Ordovician rocks were unaffected by Acadian deformation, and that the two generations observed in these rocks represent two phases in the Taconic deformation. Acadian folds are usually recognized only in the Siluro-Devonian rocks for example, in the Stokes Mountain area.

In the northern and western Gaspé areas a similar deformational scheme is now recognized (Carrara and Fyson, 1973), with two phases of Taconic deformation in the Shickshock and Quebec groups of Ordovician age and an Acadian fold generation observed exclusively in the Siluro-Devonian rocks. This writer, however, is unconvinced by the arguments set forth by Carrara and Fyson (1973) for reasons explained in the previous section and considers that the second generation in the Ordovician rocks may actually be Acadian, as originally stated by Sikander and Fyson (1969). There is still much to be desired on the level of field studies in all of these areas and, although the problems have been approached in a more systematic manner than those in central Vermont, detailed studies in small areas are still lacking.

Although it is inappropriate to draw direct comparisons between the Ludlow area and the Sutton and Stokes Mountain areas, the deformational histories may have a few similar points. B_3 deformation in this region corresponds to the late folding associated with the Chester Dome and other similar features in eastern Vermont. In central Vermont, rocks classified as Siluro-Devonian in age are folded about the Chester Dome, thus B_3 folding is probably Acadian in age. If this age is accepted then the B_2 and B_1 deformations might correlate with the two "pulses" of the Taconic orogeny observed in the Sutton area. The writer considers that there is no compelling evidence for a two-pulse Taconic orogeny in the central Vermont area, hence the earliest generation (B_1) may correspond to an earlier orogeny in the Moretown rocks. The question of the age of the Moretown rocks will remain a problem for some time. The complexity of structure and gneissic composition of the Moretown stand out in contrast to the neighboring rocks, and many possibilities exist concerning its origin. For example, it is even worth considering that the Moretown could be a fault-bounded slice of basement, remobilized along with the ultramafic intrusion, although no concrete evidence exists to prove or deny this possibility.

The existence of a two-phase Taconic orogeny has been supported by Rickard's work on radiometric dating of recrystallized micas in cleavage domains associated with different fold generations (Rickard, 1965).

Rickard, however, does not adequately justify his most important assumption; i.e., that the cleavage formed by recrystallization of micas in the cleavage planes. It may well be that the cleavage formed by selective removal of quartz within the cleavage planes, leaving behind a residuum of early micas which rotate into parallelism with the cleavage surface. Given such an alternative mechanism, the date obtained from the micas would always be older than the date of the orogeny in which the cleavage formation occurred; in some cases the date may be the same as the age for the original crystallization of mica in the earliest deformation. Rickard found nearly identical dates for the two generations when the mica-rich domains were dated, but he interpreted this as proof that the two deformations were of the same age. Alternatively, the results may be interpreted as proof that cleavage may form by selective removal of quartz from cleavage planes, without mica recrystallization. Thus, the writer considers that there is insufficient evidence to rule out the effects of Acadian deformation on the Cambro-Ordovician rocks in the Northern Appalachians; and that the two-pulse Taconic orogeny has not yet been adequately demonstrated.

BIBLIOGRAPHY

Albee, A. L., 1972, Stratigraphic and structural relationships across the Green Mountain anticlinorium in northcentral Vermont: New England Intercoll. Geol. Conf., Guidebook for 64th Ann. Mtg., p. 179-194.

Beland, J., 1967, Contributions from systematic studies of minor structures in the southern Quebec Appalachians, in Appalachian Tectonics (T.H. Clark, ed.): Roy. Soc. Canada, Spec. Publ. 10, p. 48-56.

Billings, M. P., 1950, Stratigraphy and the study of metamorphic rocks: Geol. Soc. America Bull., v. 61, p. 435-448.

Bishop, D. G., 1972, Transposition structures associated with cleavage formation in the Otago schists: New Zealand Jour. Geol. and Geophys., v. 15, p. 360-371.

Blackburn, W. H., and Dennen, W. H., 1968, Flattened garnets in strongly foliated gneisses from the Grenville series of the Gananoque area, Ontario: Am. Mineralogist, v. 53, p. 1386-1393.

Braun, G., 1969, Computer calculated nets for petrofabric and structural analysis: Neues Jarhbuch fur Min. (Monats.) Heftll, p. 469-476.

Brown, E. H., 1968, Metamorphic structures in part of the eastern Otago schists: New Zealand Jour. Geol. and Geophys., v. 11, p. 41-65.

Cady, W. M., 1960, Stratigraphic and geotectonic relationships in northern Vermont and southern Quebec: Geol. Soc. America Bull., v. 71, p. 531-576.

________, 1961, Excursion across Green Mountains, Hineburg to Montpelier: New England Intercoll. Geol. Conf., Guidebook for 53rd Ann. Mtg., section A-1.

________, 1967, Geosynclinal setting of the Appalachian Mountains in southeastern Quebec and northwestern New England, in Appalachian Tectonics (T. H. Clark, ed.): Roy. Soc. Canada, Spec. Publ. 10, p. 57-68.

Carrara, A., and Fyson, W. K., 1973, Taconic and Acadian folds in northern and western Gaspé Peninsula, Quebec: Canadian Jour. Earth Sci., v. 10, p. 498-509.

Chang, P. H., Ern, E. H., and Thompson, J. B., 1965, Bedrock geology of the Woodstock Quadrangle, Vermont: Vermont Geol. Surv. Bull. 29, 65 pp.

Clark, T. H., 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. America Bull., v. 45, p. 1-20.

Cox, F. C., 1969, Inclusions in garnet: discussion and suggested mechanism of growth for syntectonic garnets: Geol. Mag., v. 106, p. 57-62.

Currier, L. W., and Jahns, R. H., 1941, Ordovician stratigraphy of central Vermont: Geol. Soc. America Bull., v. 52, p. 1487-1512.

de Wit, M. J., 1972, The geology around Bear Cove, Eastern White Bay, Newfoundland: Unpublished Ph.D. Thesis, University of Cambridge, England.

________, 1974, Metamorphic textures and deformation: a new mechanism for the development of syntectonic porphyroblasts and their inclusion fabrics: in press.

Dalziel, I. W. D., and Bailey, S.W., 1968, Deformed garnets in a mylonitic rock from the Grenville front and their tectonic significance: Am. Jour. Sci., v. 266, p. 542-562.

Engel, A. E. J., 1949, Studies of cleavage in the metasedimentary rocks of the northwest Adirondack Mountains, New York: Trans. Amer. Geophys. Union, v. 30, p. 767-784.

Ern, E. H., 1963, Bedrock geology of the Randolph Quadrangle, Vermont: Vermont Geol. Surv. Bull. 21, 96 pp.

Eskola, P. E., 1949, The problem of mantled gneiss domes: Quart. Jour. Geol. Soc. London, v. 104, Part 4, pp. 461-476.

Gregg, W. J., 1975, Tabular garnets in metamorphic rocks from east-central Vermont: Abstract, Geol. Soc. America, Northeastern Section Ann. Mtg., p.

________, and Nisbet, B. W., 1975, Secondary and transposed layering in metamorphic rocks with examples from east-central Vermont: in preparation.

Gresens, R. J., 1966, Dimensional and compositional control of garnet growth by mineralogical environment: Amer. Mineralogist, v. 51, p. 524-528.

Hawkins, J. B., 1962, Helicitic textures in garnets from the Maine rocks of Moidant: Trans. Edinb. Geol. Soc., v. 18, p. 315-324.

Heim, A., 1900, Gneissfaltelung in alpinem Centralmassiv, ein Beitrag zur Kenntis der Stauungsmetamorphose: Naturf. Gesellshaft in Zurich, Vierteljahrschrift, v. 45, p. 205-226.

Hepburn, J. G., 1972, Geology of the Guilford Dome area, southeastern Vermont: New England Intercoll. Geol. Conf., Guidebook for 64th Ann. Mtg., p. 231-243.

Hobbs, B. E., 1965, Structural analysis of the rocks between the Wyangala batholith and the Copperhannia thrust, New South Wales: J. Geol. Soc. Australia, v. 12, p. 1-24.

________, 1966, The structural environment of the northern part of the Broken Hill orebody: J. Geol. Soc. Australia, v. 13, p. 315-338.

Hobbs, B. E., Means, W. D., and Williams, P. F., 1976, An Outline of Structural Geology: John Wiley and Sons, Inc., 572 pp.

Jahns, R. H., 1967, Serpentinites of the Roxbury District, Vermont, in Ultramafic and Related Rocks (P. J. Wyllie, ed.): John Wiley and Sons, Inc., New York, 464 pp.

Knopf, E. B., 1931, Retrogressive metamorphism and phyllonitization: Amer. J. Sci., v. 21, p. 1-27.

Means, W. D., 1963, Mesoscopic structures and multiple deformation in the Otago schist: New Zealand Jour. of Geol. and Geophysics, v. 6, p. 801-816.

________, 1966, A macroscopic recumbent fold in schist near Alexandria, Central Otago: New Zealand Jour. Geol. and Geophys., v. 9, p. 173-194.

________, 1976, Stress and Strain: Basic Concepts of Continuum Mechanics for Geologists: Springer-Verlag, New York, in press.

Nicholson, R., 1966, Metamorphic differentiation in crenulated schists: Nature, v. 209, p. 68-69.

Osberg, P. H., 1965, Structural geology of the Knowlton-Richmond area, Quebec: Geol. Soc. America Bull., v. 76, p. 223-250.

Phillips, A. H., and Hess, H. H., 1936, Metamorphic differentiation at contacts between serpentinite and siliceous country rocks: Am. Mineralogist, v. 21, p. 333-362.

Powell, D., and Treagus, J. E., 1967, On the geometry of S-shaped inclusion trails in garnet porphyroblasts: Mineral. Mag., v. 36, p. 453-456.

Ransom, D. M., 1968, The relationship of Lode Shape to wall-rock structure in the southern half of the Broken Hill orebody: J. Geol. Soc. Australia, v. 15, p. 57-64.

Ramsay, J. G., 1962, The geometry and mechanics of formation of "similar" type folds: J. Geol., v. 70, p. 309-327.

Richardson, C. H., 1928, The geology and petrography of Reading, Cavendish, Baltimore and Chester, Vt.: Vermont State Geologist 16th Report, p. 208-248.

Rickard, M. J., 1965, Taconic orogeny in the western Appalachians: experimental application of microtextural studies to isotopic dating: Geol. Soc. America Bull., v. 76, p. 523-536.

Rosenfeld, J. L., 1970, Rotated garnets in metamorphic rocks: Geol. Soc. America Spec. Paper 129, p. 1-105.

Ross, J. V., 1973, Mylonitic rocks and flattened garnets in the southern Okanagan of British Columbia: Canadian Jour. Earth Sci., v. 10, p. 1-16.

St. Julien, P., 1967, Tectonics of part of the Appalachian region of southeastern Quebec (southwest of the Chaudiere River), in Appalachian Tectonics (T. H. Clark, ed.): Roy. Soc. Canada Spec. Publ. 10, p. 41-47.

Sander, B., 1911, Über Zusammenhange zwischen Teilbewegung und GeFüge in Gesteinen: Tschermaks Mineralog. Petrog. Mitt., v. 38, p. 281-317.

Sikander, A. H., and Fyson, W. K., 1969, Structural development of the Paleozoic rocks of western Gaspé, Quebec: Canadian Jour. Earth Sci., v. 6, p. 1113-1127.

Skehan, J. W., 1961, The Green Mountain anticlinorium in the vicinity of Wilmington and Woodford, Vermont: Bull. 17, Vt. Geol. Surv., 159 pp.

________, and Hepburn, J. C., 1972, Stratigraphy of the east flank of the Green Mountain anticlinorium, southern Vermont: New England Intercoll. Geol. Conf., Guidebook for 64th Ann. Mtg., p. 3-26.

Spry, A., 1963, The origin and significance of snowball structure in garnet: J. Petrol., v. 4, p. 211-222.

________, 1969, Metamorphic Textures: Pergamon Press, London, 350 pp.

Stauffer, M. R., 1970, Deformation textures in tectonites: Canadian Jour. Earth Sci., v. 7, p. 498-511.

Talbot, J. L., and Hobbs, B. E., 1968, The relationship of metamorphic differentiation to other structural features at three localities: Jour. Geol., v. 76, p. 581-587.

Thompson, J. B., 1950, A mantled gneiss dome in southeastern Vermont: Unpublished Ph.D. Thesis, Massachusetts Institute of Technology.

________, 1952, Southern Vermont: Geol. Soc. America 55th Ann. Mtg., Guidebook for Field Trips in New England, p. 14-23.

________, 1972, Lower Paleozoic rocks flanking the Green Mountain anticlinorium: New England Intercoll. Geol. Conf., Guidebook for 64th Ann. Mtg., p. 215-229.

Turner, F. J., 1941, The development of pseudo-stratification by metamorphic differentiation in the schists of Otago, New Zealand: Amer. J. Sci., v. 239, p. 1-16.

________, F. J., and Weiss, L. E., 1963, Structural Analysis of Metamorphic Tectonites, New York: McGraw-Hill Book Company, 545 pp.

White, W. S., 1949, Cleavage in east-central Vermont: Trans. Am. Geophys. Union, v. 30, p. 587-594.

________, and Jahns, R. H., 1950, Structure of central and east-central Vermont: Geol. Soc. America Bull., v. 58, p. 179-220.

________, and Dall, C. G., 1961, Excursion from Barre to Strafford via Bradford: New England Intercoll. Geol. Conf., Guidebook for 53rd Ann. Mtg., section A-2.

Williams, P. F., 1967, Structural analysis of the Little Broken Hill area, New South Wales: J. Geol. Soc. Australia, v. 14, p. 317-331.

________, P. F., 1972, Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagin, Australia: Am. J. Sci., v. 272, p. 1-47.

Wilson, M. R., 1971, On syntectonic porphyroblast growth: Tectonophysics, v. 11, p. 239-260.

Wood, B. L., 1963, Structure of the Otago schists: New Zealand Jour. Geol. and Geophys., v. 6, p. 641-680.

Woodland, B. G., 1965, The geology of the Burke Quadrangle, Vermont: Vermont Geol. Surv. Bull. 28, 151 pp.