Some Aspects of Deformation Fabrics along the Highland/Lowland Boundary, Northwest Adirondacks, New York State

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 Geology

School of Science and Mathematics
Department of Geological Sciences

Peter C. Hall

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ABSTRACT - The Geology of the

Whippoorwill Corners Area, NW Adirondacks, NYS

In the northwestern Adirondack Mountains of New York State lies a 110 km long, 0 - 5 km wide topographic and lithologic boundary which divides the Adirondack Highlands from the Northwest Lowlands. Structurally this boundary is apparently a highly strained zone of intensely foliated and lineated rocks. Within the zone mineral grains are highly recrystallized and of fine grain size with respect to rocks outside the zone, quartz becomes undulose, and mineral assemblages frequently exhibit retrograde characteristics. Despite the apparent high strain no demonstrable offset may be seen across the zone. Furthermore, finite strain indicators such as lineation and foliation orientations show no variation inside and outside the zone. Likewise, sense of shear indicators, which include rotated augen and multiple foliations, and two heretofore undescribed indicators: hornblende fabrics and oblique secondary quartz ribbons, rarely show asymmetry. The rare asymmetries which can be found are inconclusive, with 67% showing a northwest side down, or normal fault motion. This decided lack of the standard evidence for a simple shear zone suggests that the entire region has suffered a very high strain such that changes in finite strain can not be accurately measured, or, perhaps more probably, that the region has undergone a late progressive coaxial deformation.

ABSTRACT - Grain Boundary Bands

Grain Boundary Bands (GBB) are small, elongate rims which may be observed on feldspar grains in many northwest Adirondack quartzofeldspathic rocks. These albitic bands exhibit a strong preferred orientation within a given rock, although the significance of the orientation is enigmatic. They are probably a late feature, as they are unstrained and free of the dirt and inclusions which are common in their host grains. They appear in optical continuity with the host plagioclase or potassium feldspar and are found only in contact with other feldspars. A sharp change in composition exists across the boundary of the band and its host, and the boundaries themselves generally appear correspondingly sharp. Occasionally, however, the boundary may be cuspate and less distinct, and it may exhibit faint trails at a high angle to the walls. It appears that the GBB are produced by a dilational mechanism, and it is thought likely that grain boundary sliding and diffusive mass transfer play important roles. This indicates that grain boundary sliding may be active in coarser grain rocks (.15 to .2 mm diameter) than previously thought possible. Furthermore, this is the first direct observation of a microstructure which might be attributed to grain boundary sliding in rocks.

ACKNOWLEDGEMENTS

I am greatly indebted to all of the people who helped me out during this project, both at SUNY and at Colgate University. Thanks are especially due to Dr. Win Means for his critical mind and his logical thought processes, some of which I hope I have acquired. Special thanks are also due to Drs. William Bosworth and James McLelland for getting me started in the Adirondacks, and for their overwhelming confidence in me.

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State University of New York at Albany

School of Science and Mathematics

Department of Geological Sciences

The thesis for the master's degree submitted by

Peter Chapman Hall

under the title

Some Aspects of Deformation Fabrics

along the

Highland/Lowland Boundary,

Northwest Adirondacks, New York State

has been read by the undersigned. It is hereby recommended for acceptance by the Faculty with credit to the amount of six semester hours.

W. D. Means 11/30/84

(signed) (date)

George W. Putman 11/30/84

(signed) (date)

W. S. Fyfe 30/11/84

(signed) (date)

Recommendation accepted by the Dean of Graduate Studies for the Graduate Academic Council.

(signed) (date)

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PART I

The Geology of the Whippoorwill Corners Area
Northwest Adirondacks, New York State

CHAPTER ONE - INTRODUCTION

The Adirondack Mountain region of northeastern New York State is a roughly circular uplift of Grenvillian gneisses which is surrounded by a cover of Cambrian and Ordovician sedimentary rocks. Covering approximately 25000 square kilometers (10000 square miles), the intensely deformed and metamorphosed rocks rise sharply along fault boundaries in the east, reaching a maximum elevation of 5344′ above sea level, and gradually decrease in relief and elevation westward to disappear unconformably under the younger sediments.

Slightly to the east of the center of the Adirondack uplift are a number of anorthosite massifs which form the highest peaks in the area (Fig. I-1). Surrounding these are a variety of metamorphosed charnockitic, mangeritic, granitic, and gabbroic rocks, as well as diverse metasedimentary units. The foliation in the rocks tends to circle around the borders of the anorthosite, and on a very large scale is believed to form hundred kilometer amplitude isoclinal folds (McLelland & Isachsen, 1980). Further development of structural patterns in the Adirondacks is thought to result from interference of several fold

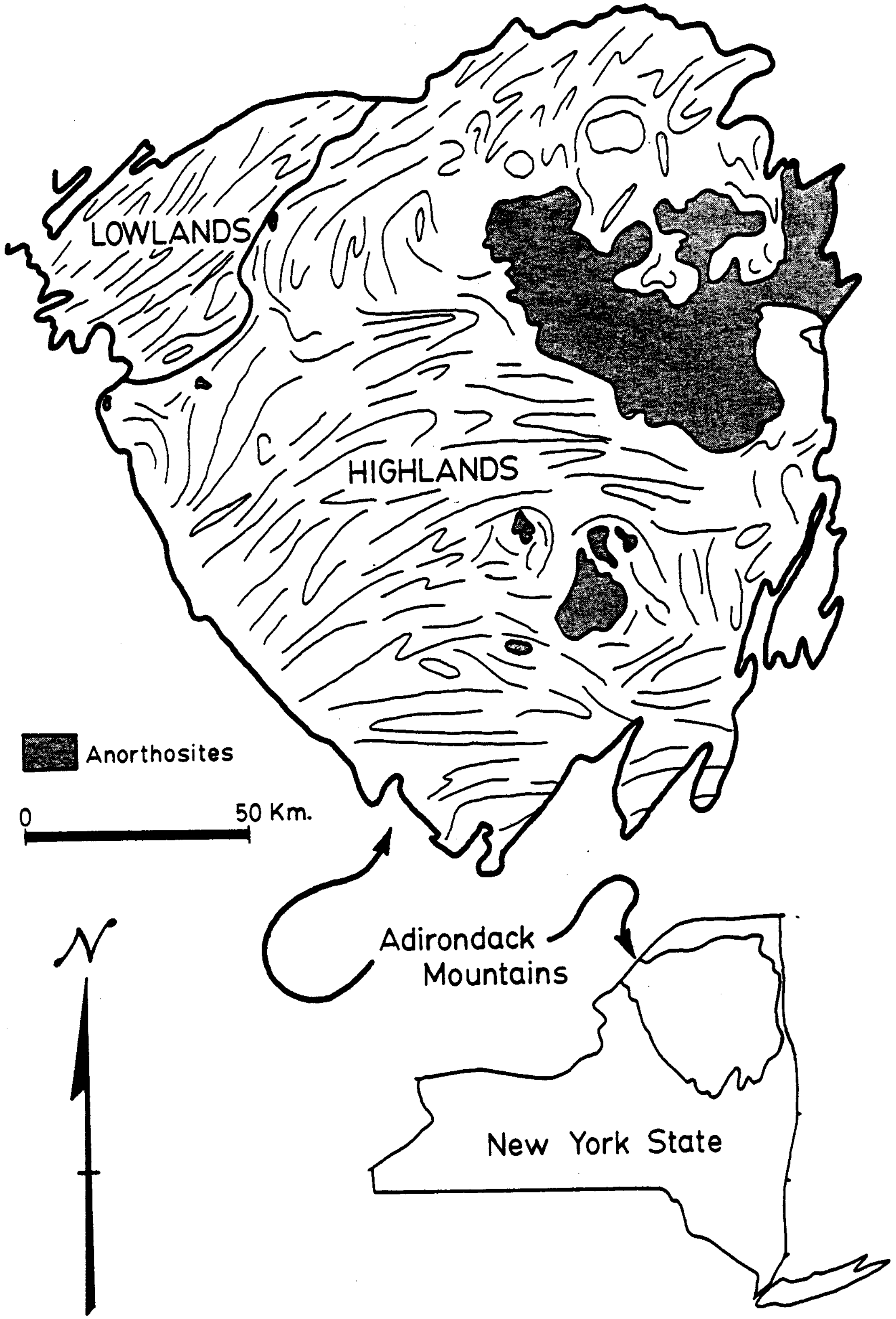


Figure I-1. Location of and foliation trends within the Adirondack Mountains, New York State (After Whitney, 1983).

generations (Foose & Carl, 1980; McLelland & Isachsen, 1980).

This deformation, and the corresponding metamorphism of the Adirondack region is thought to have occurred around 1100 Ma ago. Recently, Grant and others (1984) have suggested that three pulses of metamorphism actually occurred, one at about 1250 Ma, one around 1100 Ma, and a final pulse around 950 Ma. The metamorphic environment at that time has been likened to that of Tibet today (Dewey & Burke, 1973) due to the mineral assemblages present which represent pressures of seven or eight kilobars (Jaffe et al., 1978), and temperatures of 550^{o}C - 800^{o}C. The temperatures tend to decrease away from the main anorthosite massif, and concentric isograds may be drawn about it (Bohlen et al., 1980).

It is possible that a high temperature, low pressure metamorphic event also occurred. In many localities wollastonite is found, and ackermanite and monticellite have also been observed. However, these may also be explained by a locally very high fluid pressure (Valley & Essene, 1980).

The 650^{o}C isotherm in the northwestern Adirondacks runs roughly along a marked topographic boundary. This is the border between the Adirondack Highlands and the Northwest Lowlands (Fig. I-2). The Lowlands are a zone about 80 kilometers (50 miles) long and 40 - 50 kilometers (25 - 30 miles) wide of predominantly metasedimentary rocks known as the "Grenville Series". These consist of assorted marbles,

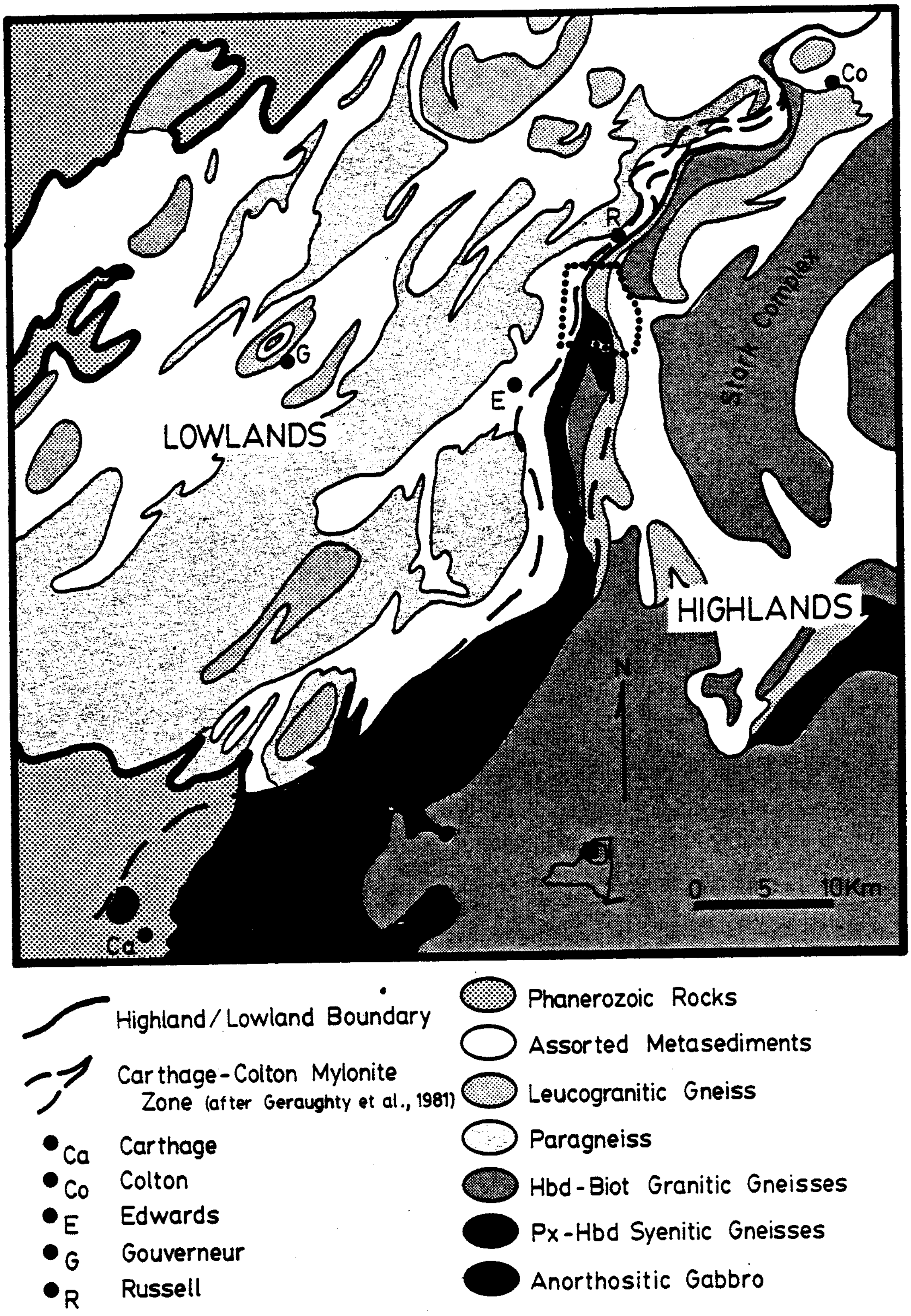


Figure I-2. Generalized geology of the Northwest Adirondacks. Dotted line borders the Whippoorwill Corners study area.

calcsilicates, and granitic gneisses, and have been extensively examined due to their economic value (Engel & Engel, 1953; Brown & Engel, 1956; DeLorraine & Dill, 1982). The Highlands, or Adirondack Core, is made up of what are primarily considered to be metaigneous rocks which include the anorthosite massifs and associated charnockites and mangerites, and a later series of granites. The rocks of the Grenville Series are also evident, especially to the south, but they are considerably less common than in the Northwest Lowlands.

Topographically, the Lowlands are characterized by relatively low relief, rolling hills, and swampy land, whereas the Highlands are hilly to mountainous, with deep, abrupt valleys and swamps.

Due to the topographic and lithologic differences across it, the Highland/Lowland boundary has long been suspected of being of fault related origin. The first reference to this was probably made by C. H. Smyth in 1899 when he referred to a "cataclastic structure" near Harrisville, New York (Geraghty et al., 1981).

In the early and middle 20th century, Buddington extended this description to include a zone 70 miles long and 8 - 10 miles wide with moderately to steeply dipping foliation, a down dip lineation, and in which the Highlands rocks have an intense "cataclastic" structure. He suggested that the zone may have been the result of thrusting of the

Lowlands "syncline" over the Highlands (Buddington, 1939).

Brown & Engel (1956) agreed that the boundary was a zone of discontinuity and suggested dextral strike slip displacement on the order of three miles based on asymmetry of what they interpreted to be drag fold limbs.

The 1961 edition of the New York State map labelled the area as the Highland-Lowland Boundary Fault (Broughton et al., 1961), and this was described by Buddington and Leonard (1962) as a zone of repeated movement characterized by slickensided surfaces, fractures, mylonites, and retrograde metamorphism.

In 1963 Buddington noted that, while the Highland/Lowland Boundary includes areas of extensive mylonitization, it also includes zones in which primary igneous textures remain, and that this variation often seems to be related to the presence of water. Generally, the more OH^- rich assemblages are more intensely foliated.

In more detailed mapping of the boundary region, Hargraves (1968) noted that the increase in mylonitization near the boundary was somewhat related to the rock type, with the more granitic rocks tending to be more intensely foliated.

Leitzke (1974) also found a lithological correlation with degree of deformation and suggested a slight but sudden change in lineation directions within Highland versus Lowland rocks. He suggested that the discontinuity was due to competency contrasts between the metasedimentary and metaigneous rocks.

A reconnaissance survey was published for the Nuclear Regulatory Commission (Geraghty et al., 1981) which described the "Carthage-Colton Mylonite Zone" as being 110 km long and 0 to 5 km wide. For ease of field recognition, the mylonite was defined as a zone of reduced grainsize and strong planar fabric. The study found a correlation with rock type with the best evidence for mylonitization occurring in granitic rocks, and evidence decreasing respectively in anorthositic gabbros, calc-silicates, metapelites, and amphibolites. Almost no mylonitic fabric was seen in the marbles.

The zone was seen to be essentially parallel to foliations except in the north where it narrows and disappears. When it reappears, it does so further to the northeast, and it appears that it must crosscut several lithologies.

Based on this possible crosscutting relationship and on inferred high strain rate, Geraghty et al. suggest a ductile fault origin for the zone, and have suggested formation as an eastward directed thrust associated with contemporaneous folding.

Weiner (1983) mapped a small section of Lowlands rocks which abuts the mylonite zone in the Harrisville-Pitcairn area and discovered that xenoliths in the Diana Complex, a large intrusive body entirely within the mapped mylonite zone, correlate with the adjacent metasediments. He also found that there is little or no simple shear movement on the boundary since folding. At the same time, Weiner believes that the mylonitic foliation is axial planar to what he determines to be second phase isoclinal folds, and therefore is synchronous with them. Weiner concludes that the mylonitic fabric originated from intense flattening with a small simple shear component.

A 1980 COCORP profile across the zone failed to shed much light on the problem. A possible eastward dipping reflector may indicate some discontinuity in the vicinity of the Carthage-Colton zone, but this is highly speculative (Brown et al., 1983).

Clearly a great deal of confusion exists as to the nature of the boundary between the Adirondack Highlands and the Northwest Lowlands. It has been called a simple topographic boundary, a brittle fault zone, a ductile simple shear zone, and a zone of ductile flattening. High strains have been attributed to dextral thrusting (Buddington, 1939; Geraghty et al., 1981), sinistral thrusting on a now folded fault plane (Whitney, 1983), strike-slip faulting (Brown & Engel, 1956; Leitzke, 1974), and axial planar shortening (Weiner, 1983).

To attempt to help sort out this confusion, detailed mapping and sense-of-shear analysis was undertaken in the Whippoorwill Corners area between Russell and Edwards, New York (Fig. I-2). Just south of Whippoorwill Corners, the Diana Complex narrows and disappears, and the high strain zone also narrows to a width of a few hundred meters. This narrowing makes the area ideal for a detailed study. The high strain zone is of an easily mappable size, the relative effects of lithic contrasts between the Diana and Lowland metasediments can be observed, and the difference between Highland and Lowland lithologies can be determined. Progressive development of high strain fabric can also be seen in both directions, and sense of shear indicators may be attributed to the formational event.

Work in the study area was first reported by C. H. Smyth in 1896 in a paper dealing with the petrography of a small anorthositic gabbro body in the area. This body was reexamined in 1921 by W. Miller.

Miller in 1926 submitted a manuscript and map of the Russell 15′ Quadrangle to the New York State Museum and Science Service, however it was never published. The first published report on the quadrangle was a reconnaissance survey by Nelson Dale (1934). Dale was primarily concerned with the general features of the area and their economic significance.

Buddington (1963, 1969) discussed many of the granites as well as the anorthositic gabbro and their deformation characteristics, mineralogy, and chemistry.

Leitzke (1974) mapped a strip in the south half of the Hermon 7 1/2′ Quadrangle which traversed the Highland/Lowland Boundary, and Parodi (1978) investigated the geochemistry of the anorthositic gabbro body.

CHAPTER TWO - LITHOLOGIES

Eight lithologies are evident in the area, several of which have more than one mappable facies. These are herein described in sequence approximately from east to west, or roughly from the bottom to the top of the structural section. All units have undergone extensive metamorphism; however, some may retain relict igneous characteristics, or may bear resemblance to igneous rocks. These are accordingly referred to on occasion by the names of their igneous counterparts. It should be noted that while many of the rocks resemble igneous rock, there are no unmetamorphosed igneous rocks in the area.

Noduliferous Microcline Granitic Gneiss:

The lowermost structural unit in the area is a microcline rich granitic gneiss which commonly contains nodules of quartz plus sillimanite.

In the field it occurs as a medium to coarse grained rock which is noticably rich in potassium feldspar, and accordingly displays a deep red colour. Small round opaques may be present, and these often weather to red hematite, or show iron-weathering haloes. The rock often displays a high content of biotite, muscovite, and sillimanite, and these often define a good foliation, and may form rather thick selvedges between quartz-feldspar layers.

The rock is commonly riddled with pegmatitic veins of either potassium feldspar, quartz, or both. Additionally, muscovite is not uncommon in these pegmatites. A gradation may be seen from fairly homogeneous rock, to a veined pegmatitic gneiss, and finally to an unusual noduliferous gneiss. These nodules occur as oblate ellipsoids and usually consist of quartz plus sillimanite, with occasional muscovite. The long axes of the nodules may be up to seven centimeters in length, and the ratio of major and minor axes is generally about 3:1. The intermediate axis is usually about equal to the major axis, producing circular discs when viewed normal to the foliation of the rock (Fig. I-3b).

Microscopically, the host rock proves to be extraordinarily rich in microcline, with up to 50% of the groundmass may be this type of feldspar, the rest being predominantly quartz. The quartz and microcline generally occur in roughly equal abundances. Lesser perthite and plagioclase may also be seen. Biotite, magnetite, and ilmenite, and their alteration products may be observed in varying amounts. Hematite is usually the most abundant opaque mineral. Biotite, when present, usually shows a light tan to green pleochroism, although dark green and brown biotites were also seen. Sillimanite is almost inevitably present as tabular grains, and in rare instances it displays

(a)

(b)

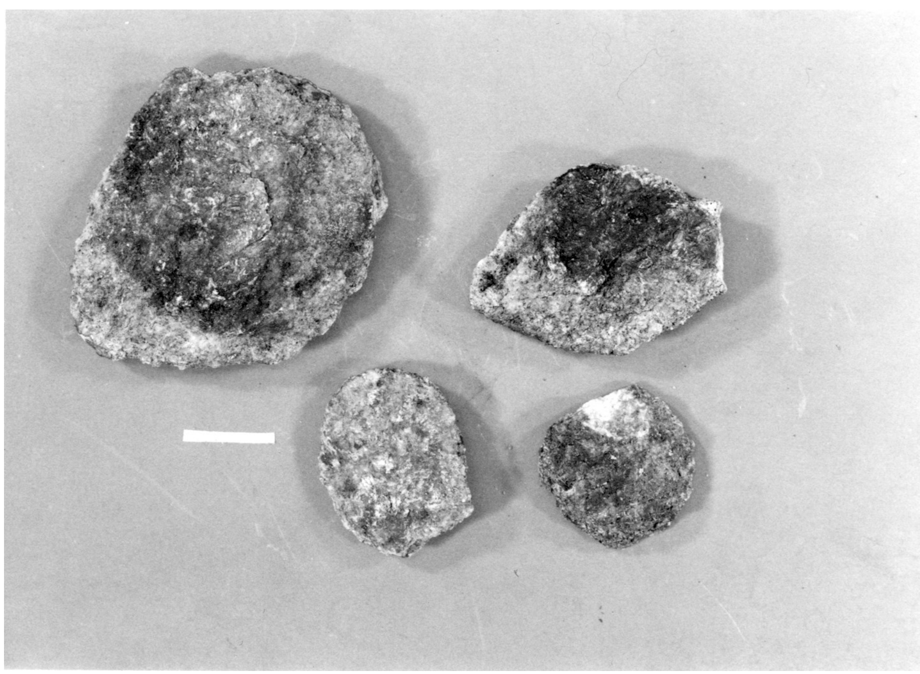


Figure I-3. Noduliferous Microcline Granitic Gneiss. (a) Outcrop photo from locality #258. (b) General shape of nodules. Scale bar is 2 cm.. Thickness of nodules averages 1 cm..

a vermicular relationship with quartz. It also may appear in fibrolitic lensoid masses with associated quartz or tabular sillimanite (Fig. I-4). These lenses are the nodules seen in outcrop.

Similar noduliferous sillimanite plus quartz gneisses have been reported to occur in other high grade terrains throughout the world, and according to Buddington (1957) result from metasomatism of biotite-quartz-plagioclase gneisses, such as that which occurs abundantly in the Northwest Lowlands. This particular gneiss has been described in great detail by Engel and Engel (1958, 1960, 1963).

The spatial gradational sequence from homogeneous to pegmatitic to noduliferous gneiss might be thought to represent a genetic sequence. Areas may be observed where veins and nodules coexist, and it may easily be imagined that the veins were stretched and boudinaged to form the nodules (Fig. I-5). Furthermore, the sillimanite-quartz-muscovite-microcline assemblage of the rock unit may suggest an origin related to the second sillimanite isograd. It is possible that as muscovite and quartz reacted to form sillimanite plus potassium feldspar plus water, the excess water created an anatectic situation in which crystallization differentiation of the sillimanite and feldspar could rapidly occur. Thus migmatitic sillimanite-quartz veins could exist in a feldspthic matrix. Deformation could subsequently disrupt

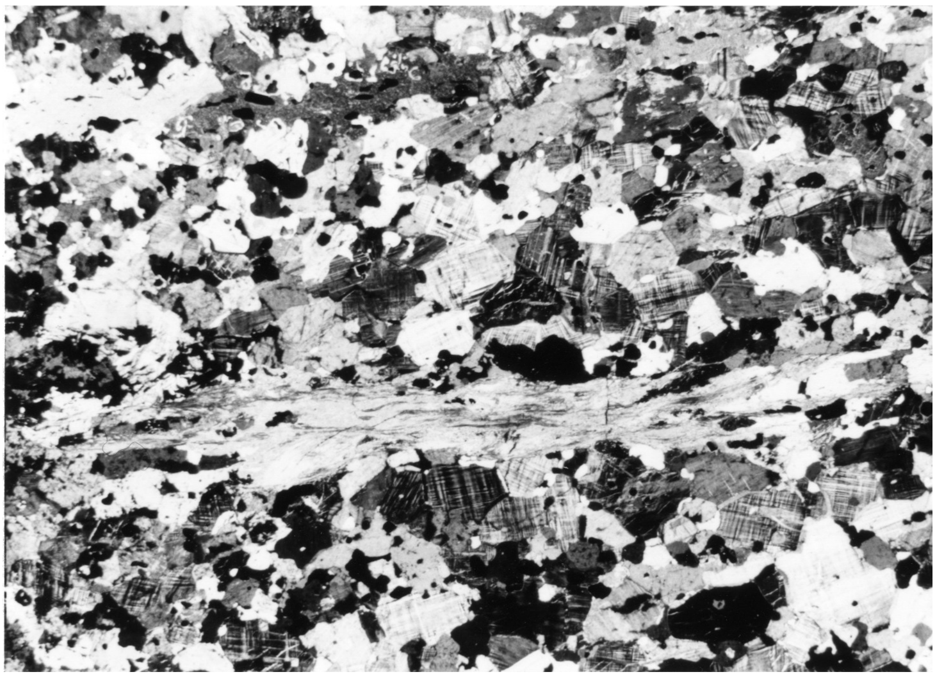


Figure I-4. Photomicrograph of Noduliferous Microcline Granitic Gneiss. Elongate mass in center is small quartz-sillimanite nodule. Long dimension of photo is 1.75 cm..

(a)

(b)

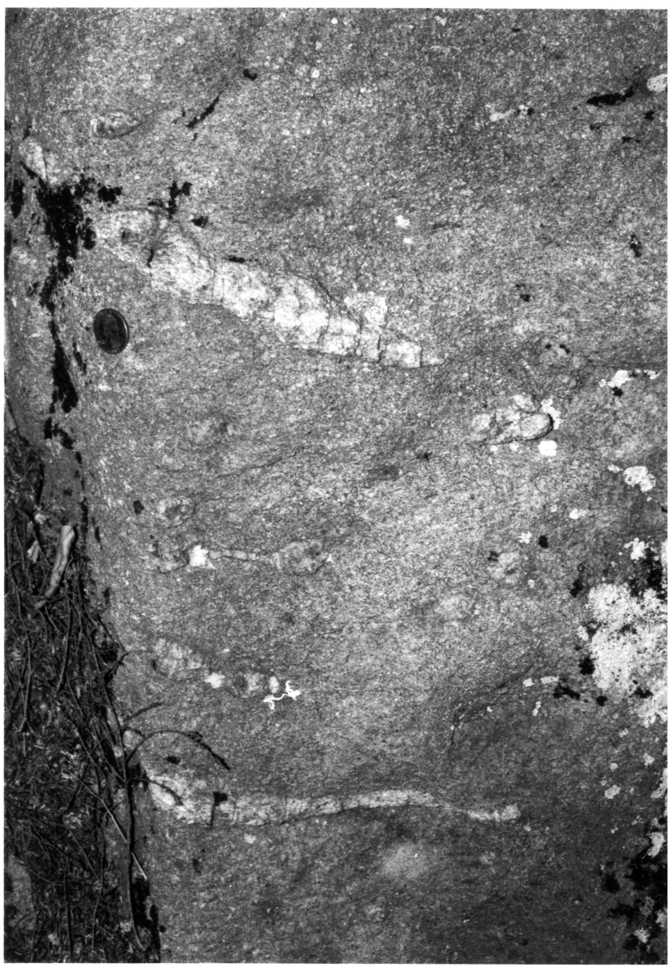


Figure I-5.
(a) Sketch of outcrop #281 showing pegmatitic quartz-sillimanite veins which are folded and drawn out into nodules.
(b) Photo of nodules with vein-like tails. Outcrop from locality #258.

these pegmatites to form the nodules so prominently displayed.

The noduliferous microcline granitic gneiss frequently includes meters-thick layers of blocky quartzite which may be white or faint green in colour.

<u>Eastern Granitic Gneiss</u>:

<u>Main Facies</u>: The easternmost granitic gneiss is equivalent to what Buddington (1939) described as the "Lowville-St. Regis Batholith" and classified as a hornblende granite. In this facies, however, hornblende is absent. Rather, biotite occurs as a sparse mafic constituent.

Field recognition is generally based on a low mafic content of the unit. The granite appears as a pink to deep red, and occasionally almost purple coloured rock on fresh faces, and weathers pinkish, sometimes with a black surface tint. It is usually coarse grained and may or may not have a quartz ribbon foliation. Those outcrops in which such a foliation does exist may weather to a finely striped pink and white gneiss, or to a finely corrugated face in which some layers weather away (Fig. I-6).

Outcrop faces, especially horizontal ones, often have interspersed quartz veins, several inches thick, which roughly parallel the foliation. Some purple fluorite grains may be found on fresh surfaces.

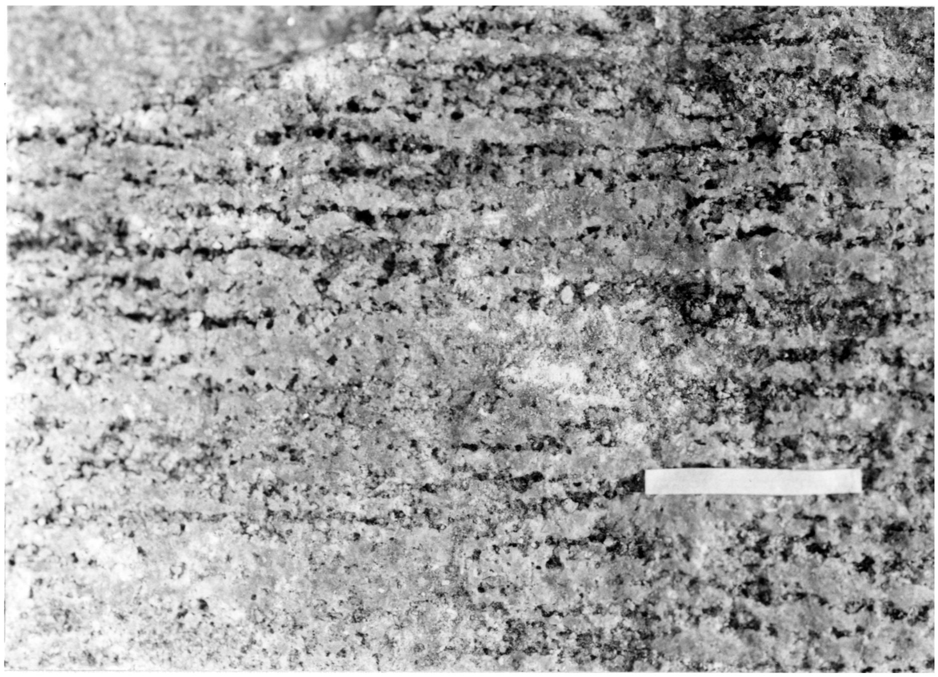


Figure I-6. Eastern Granitic Gneiss wherein weathering has created a corrugated accent of the foliation. Scale bar is 2 cm..

In thin section the feldspar in the granitic gneiss is seen to consist of somewhat more plagioclase than microcline, with some remnant perthite in less deformed samples. Feldspar content is generally high with respect to quartz, although quartz is plentiful. Rare mafics occur as light coloured tan to green pleochroic biotite, with scattered magnetite which is substantially altered to sphene, and less frequently to hematite.

Mafic Syenite: Near the contact of the eastern granitic gneiss with the structurally higher anorthositic gabbro, it is common to find an increase in mafics and a concurrent decrease in quartz. This mafic syenite consists of plagioclase and microcline with rare perthite, some quartz, hornblende, and biotite.

The hornblende and biotite appear as different generations of minerals. Blue-green hornblende is abundant and defines a fairly clear foliation in the rock. The biotite, pleochroic in shades of brown, may also lie in the foliation, but commonly crosscuts both the foliation and individual hornblende grains, as seen in figure I-7.

Anorthositic Gabbro:

Perhaps historically the most studied, and yet the most confusing unit in the study area is a body of mafic origin known variously as the Russell Gabbro (Smyth,1896; Dale,

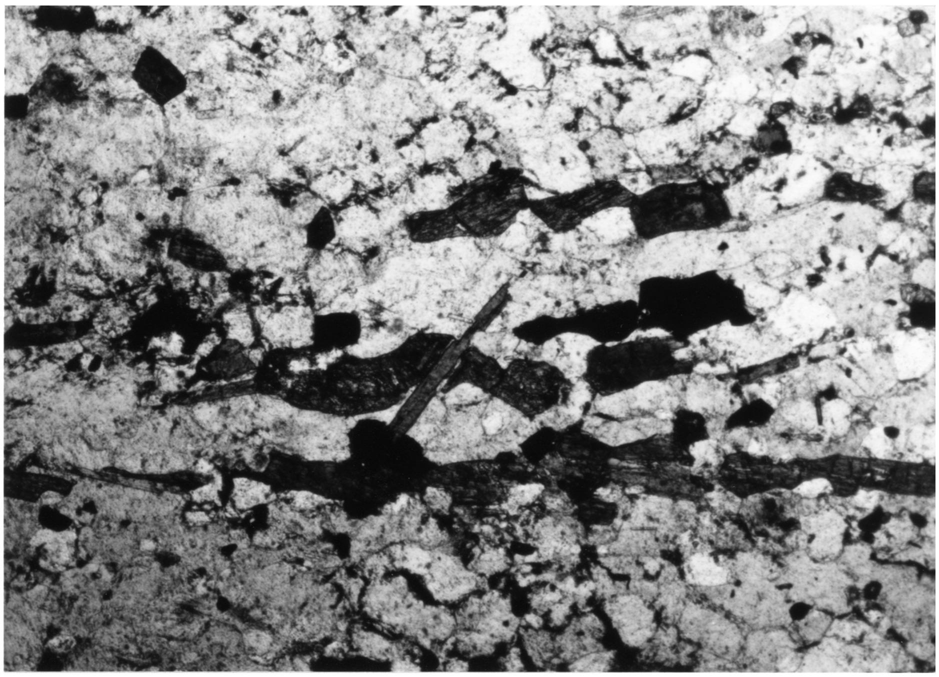


Figure I-7. Photomicrograph of Mafic Syenitic Gneiss. Note biotite which lies within hornblende foliation as well as crosscuts it. Long dimension of photo is 3.8 mm..

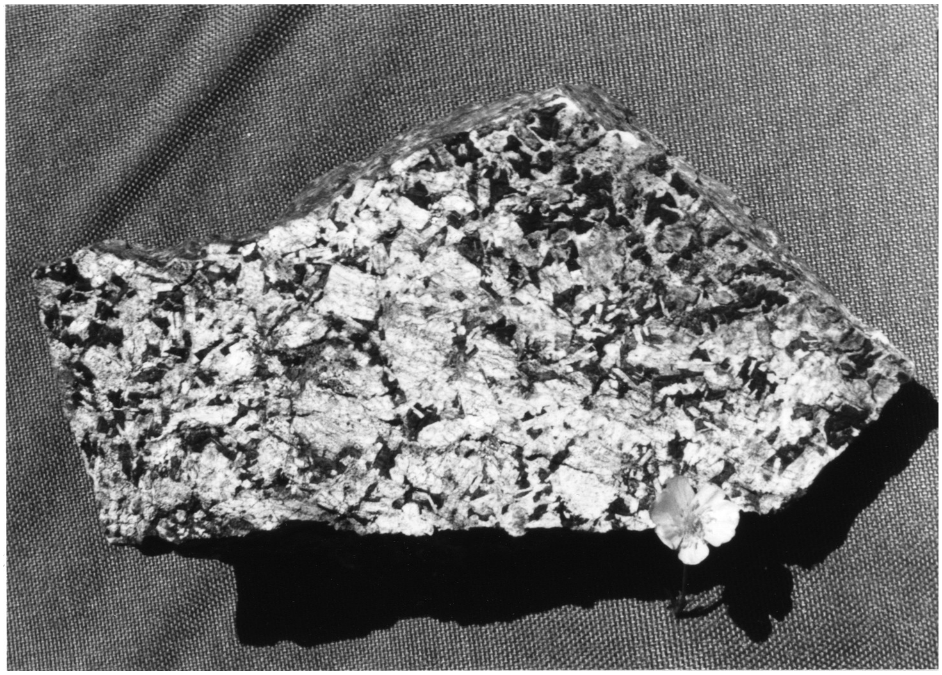
1934), the Russell Feldspathic Gabbro (Buddington, 1939), and the Dana Hill Metagabbro (Leitzke, 1974; Parodi, 1978). Miller (1921a, 1921b) acknowledged that the body ranges in composition from true gabbro to true anorthosite, and hence called it the Russell Anorthosite-Gabbro. Generally it appears that the more gabbroic members occur towards the south, and the more anorthositic to the north.

The body is apparently one in a series which stretches along the eastern border of the Diana Complex from Lowville to Russell (Buddington, 1963).

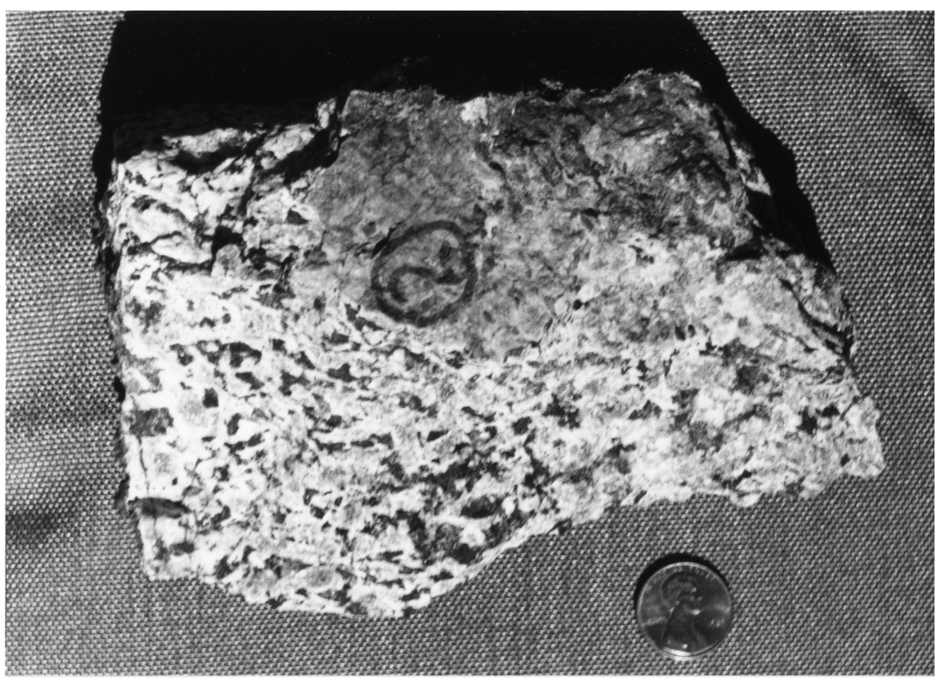
Ophitic Anorthositic Gabbro: Outcrop exposure of the undeformed rock is usually an ophitic anorthositic gabbro consisting of euhedral plagioclase laths up to eight centimeters in length, but more commonly reaching only one or two contimeters (Fig. I-8a). These are surrounded by intercumulate pyroxene and/or hornblende which ranges in color from black to bright copper green. Weathered faces may show a white and black portrait of this structure, or may appear as a uniformly dark brown or black color. In the latter event, fresh faces generally are also dark coloured, and plagioclase is seen only by reflection of light off the large cleavage faces. The weathered faces in either case often have a mottled appearance which is the result of the greater resistance of hornblende versus plagioclase to the elements.

In thin sections of the least deformed anorthositic

Figure I-8. Development of foliation in anorthositic gabbro. Samples from hill just south of Route 87.
(a) Ophitic rock. Note angular plagioclase, intercumulus pyroxene/hornblende.
(b) Sub-ophitic texture. Plagioclase is more rounded and grain boundaries are mesoscopically less distinct.
(c) Foliated and lineated gneiss. No igneous texture remains.
(d) Highly deformed fine grained sample. Some plagioclase porphyroblasts remain.



(a)



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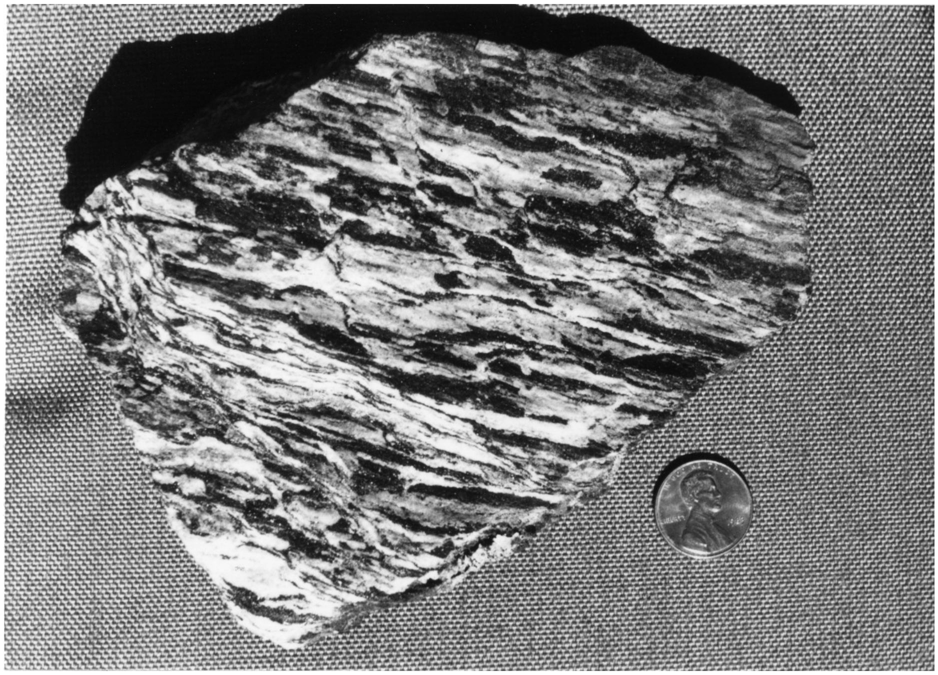

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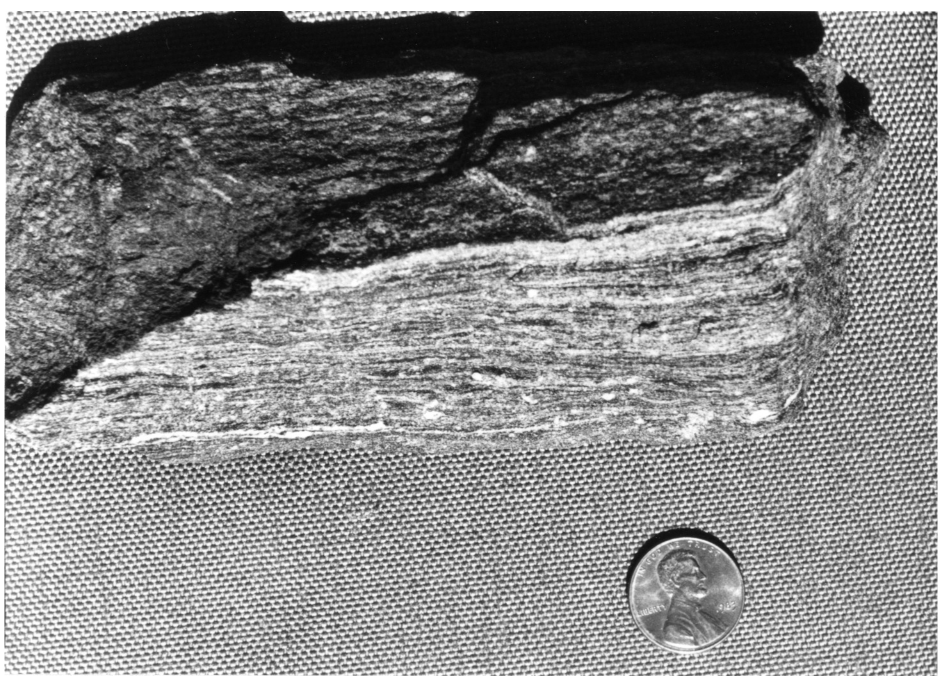


(d)

gabbro (Fig. I-9), the plagioclase is revealed to be labradorite. This is surrounded by augite which is colorless or light green in plane light. Minor magnetite occurs and is usually surrounded by biotite replacing clinopyroxene, either as radial needles or as a large mass.

In almost all cases the rock is altered to some extent from its original igneous form. Most commonly some of the pyroxene has been retrograded to a hornblende of a deep green colour. This hornblende may replace the pyroxene in whole or in part, and may occur as rims around pyroxene grains or as intergrowths of hornblende plus quartz, surrounded by more pure hornblende.

The plagioclase is usually fractured to some extent and is often recrystallized along these fractures and along the edges, especially at the corners. Some replacement by scapolite may occur, especially in these recrystallized grains. The plagioclase is also subject to alteration to a fine grain product.

<u>Border Gabbro</u>: Along the western margin of the anorthositic gabbro body, the rock assumes a more gabbroic character. Outcrops of this border facies appear as a slightly rusty black rock, rather fine grained, and unfoliated. Very narrow (<1 mm) parallel black fractures occupy positions of high relief on otherwise smooth rock faces.

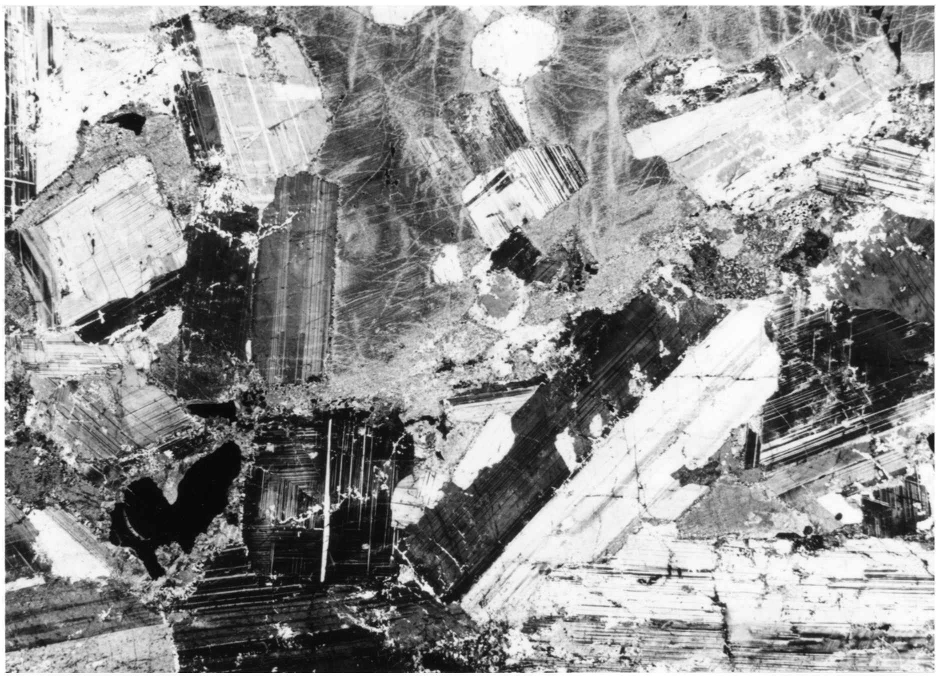


Figure I-9. Photomicrograph of ophitic anorthositic gabbro. Plagioclase is slightly fractured and recrystallized, and some pyroxene has altered to hornblende plus quartz (arrow). Long dimension of photograph is 2.8 cm..

Thin section observations reveal that the gabbro consists of highly strained plagioclase, often completely transformed from the original ophitic texture, together with large amounts of hornblende and quartz, and lesser clinopyroxene. The mineral assemblage is usually well mixed, with no trace of either an ophitic or a foliated fabric. Occasionally, right at the western contact, the border facies exhibits a mottled texture. This resembles the texture produced by high fluid flow along fractures seen elsewhere in the anorthositic gabbro (see Chapter Two, Dykes). This is in agreement with Buddington´s (1939) hypothesis suggesting fluid flow along this contact.

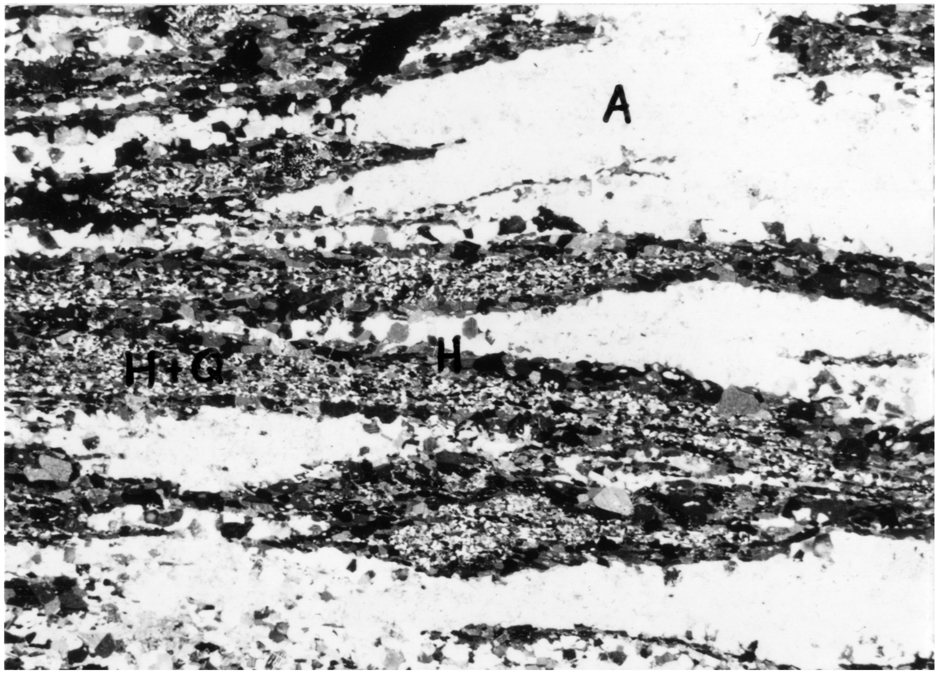
Foliated Anorthositic Gabbro: Limited areas in the southern half, and most of the northern half of the anorthositic gabbro massif show strong foliation and lineation (Fig. I-8b - d). This sheared fabric begins as a drawing out of mafic minerals and retrogression of pyroxene to blue-green hornblende; and some recrystallization of the plagioclase around the edges to form more rounded porphyroclasts. The recrystallization of the plagioclase is usually accompanied to some extent by transformation to scapolite, and it also yields a more albitic plagioclase than that in the original ophitic rock. Retrograde formation of the hornblende is usually nearly complete, and in places retrogression may be so thorough as to produce clear to reddish brown biotite. In these biotite rich areas, magnetite and ilmenite are usually replaced by sphene. Garnet may also appear in the sheared gabbroic anorthosites, and is usually associated with the opaque minerals.

With continued deformation, feldspar and mafics continue to be reduced in grain size. However, it seems that the mineralogy remains fairly constant through continued deformation. An exception is that altered plagioclase recrystallizes to clean new grains (Fig. I-10). Also, hornblende-quartz aggregates seem to lose quartz with more intense deformation.

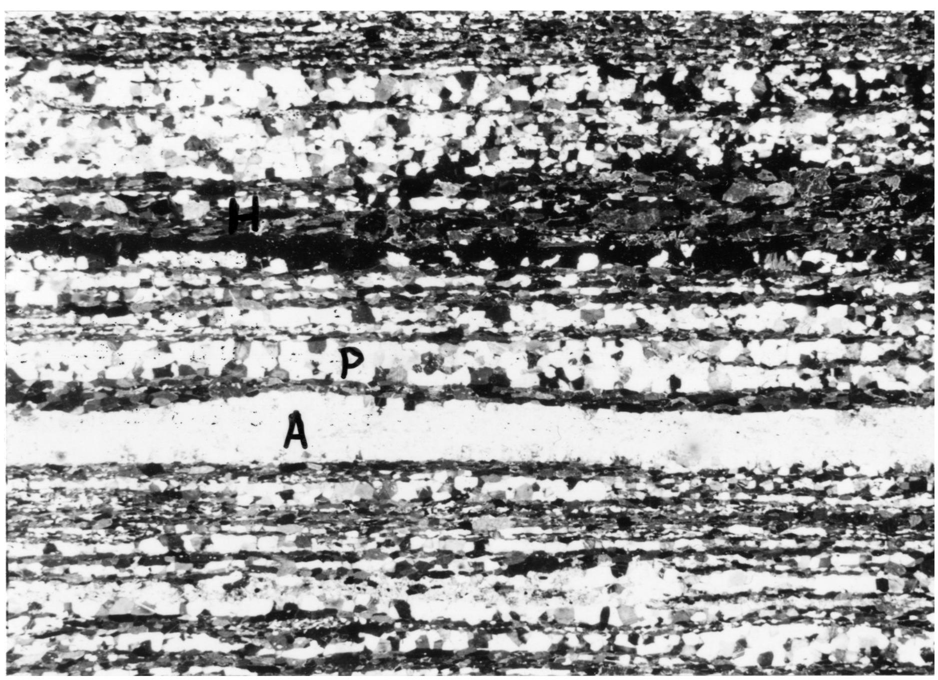
The foliated anorthositic gabbro often encloses lozenges of nearly undeformed rock, and transformations from mildly deformed rock to ultramylonite may occur within ten centimeters.

Additionally, a later brittle faulting event seems to have occurred, leaving very slightly rotated blocks across sharp discontinuities of high mafic content. An example of this may be observed at outcrop #587.

Crystalline Anorthositic Gabbro: In both the foliated and unfoliated parts of the anorthositic gabbro massif, areas occur in which the grain size is highly reduced and alteration of the plagioclase to scapolite is extensive. This has the effect of making the rock look almost crystalline in hand sample. This is especially true in the more feldspathic members, and the rock can assume a purplish,



(a)



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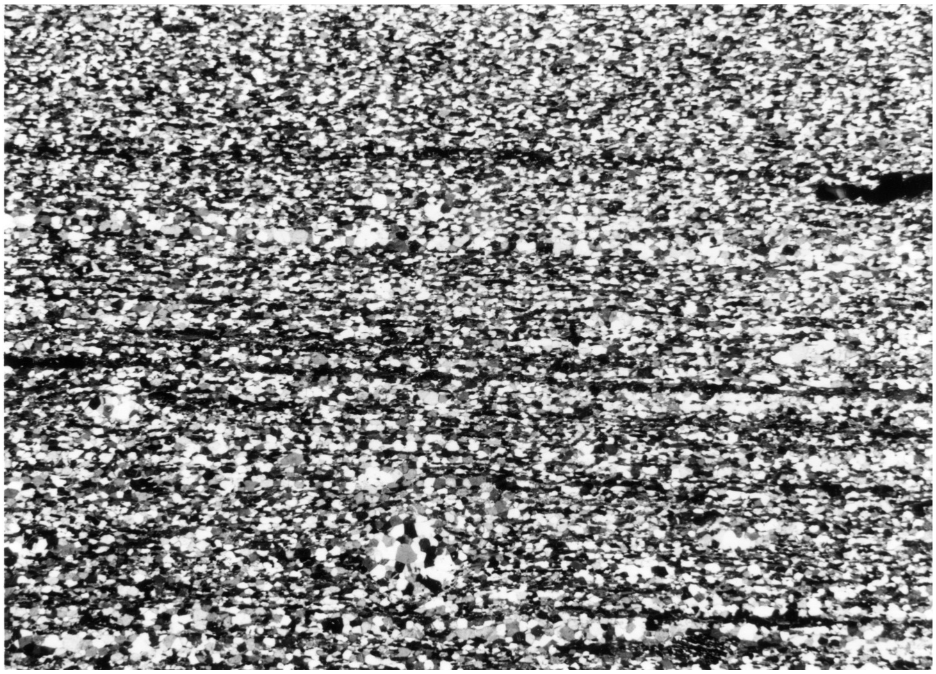


(c)

Figure I-10. Development of foliation in anorthositic gabbro, Outcrop #467.
(a) Layers of altered plagioclase (A), hornblende (H), and hornblende plus quartz (H+Q).
(b) More deformed. Layers of pure hornblende versus layers of either altered plagioclase or clean recrystallized plagioclase (P).
(c) Most deformed. Significant grain size reduction. All plagioclase is clean and recrystallized and layering is less evident than in (a) and (b).
All photos have long dimensions equal to 2.3 cm..

greenish, or very white colour which looks most unlike an anorthosite-related rock.

Potassic Anorthositic Gabbro: Toward the north end of the anorthositic gabbro body, extensive replacement of plagioclase by potassium feldspar and injection by syenitic rock may occur. The potassic replacement imparts an orangy color to the rock without noticably changing the structure. It is often associated with green pyroxene. Fractures may be seen with the orangy rock in the immediate viscinity, and it seems highly likely that the rock results from alteration by a potassium rich fluid.

Dykes: In addition to the above mentioned syenite, numerous other types of dyke-like features cut the anorthositic gabbro. These may be seen in either the ophitic or foliated rock. Where they crosscut foliation, they may be mildly distorted.

The syenite is perhaps the most prolific injection material, and may form large blobs as well as narrow dykes. It bears a striking resemblance to the mafic syenite when it appears as blobs, and these may actually result from interfingering during deformation rather than from injection.

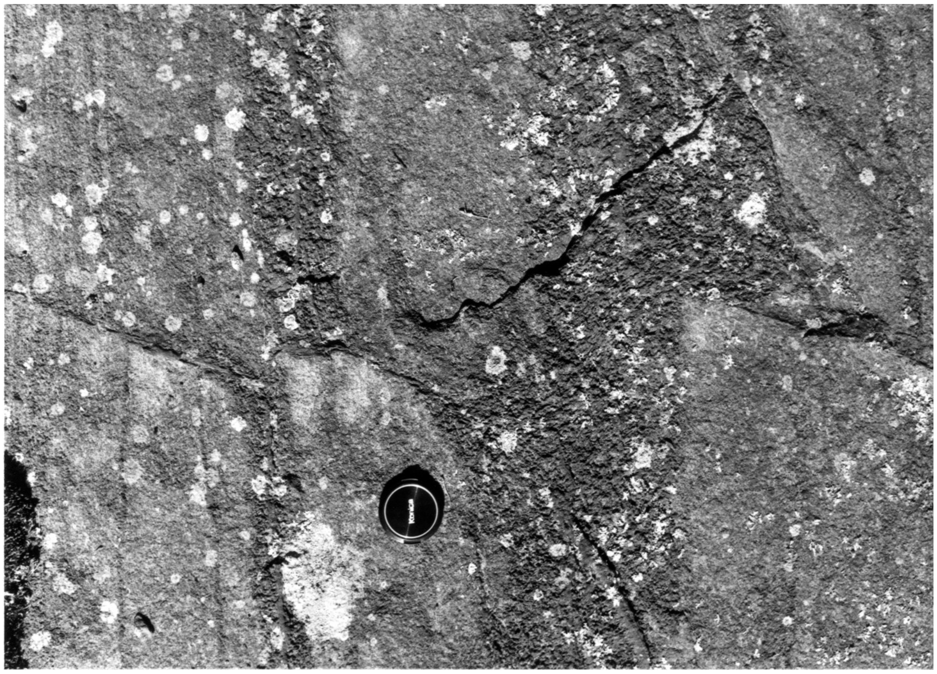
Also abundant in the anorthositic gabbro are pure white dykes consisting of either scapolite or plagioclase. In a few localities, these fill en echelon fractures.

Quartz or quartz plus plagioclase, granite, green pyroxenite, and black hornblendite also fill fractures in the area. While most of these fracture-like or dyke-like features show no chilled margins, fibers, or other features which might indicate a mechanism of formation, the hornblendite veins often exhibit bent country rock foliations, suggestive of simple shear, along their margins. Also, hornblendite veins may sometimes be observed to offset each other, as in figure I-11a. They additionally posess flanking zones where the country rock assumes a mottled texture (Fig. I-11b). This seems to suggest that the features were fractures through which a high temperature hornblendic fluid flowed and crystallized, altering the wall rock as it passed. Whether the zone developed as a minor shear zone and allowed increased flow of fluid, or whether it developed as a fracture along which fluid weakened the rock to allow ductile deformation, is enigmatic.

Diana Complex:

The Diana Complex was a primary consideration in Buddington´s work (1939,1957). He believed it to be of plutonic origin, and based on differentiation trends suggested that it may be an overturned gravity stratified sill. Subsequent work has shown that it does, indeed, appear to be of plutonic origin (Eckelman, 1980), and that crystallization differentiation features do exist, although it may be an overturned antiform (Hargraves, 1968).

(a)



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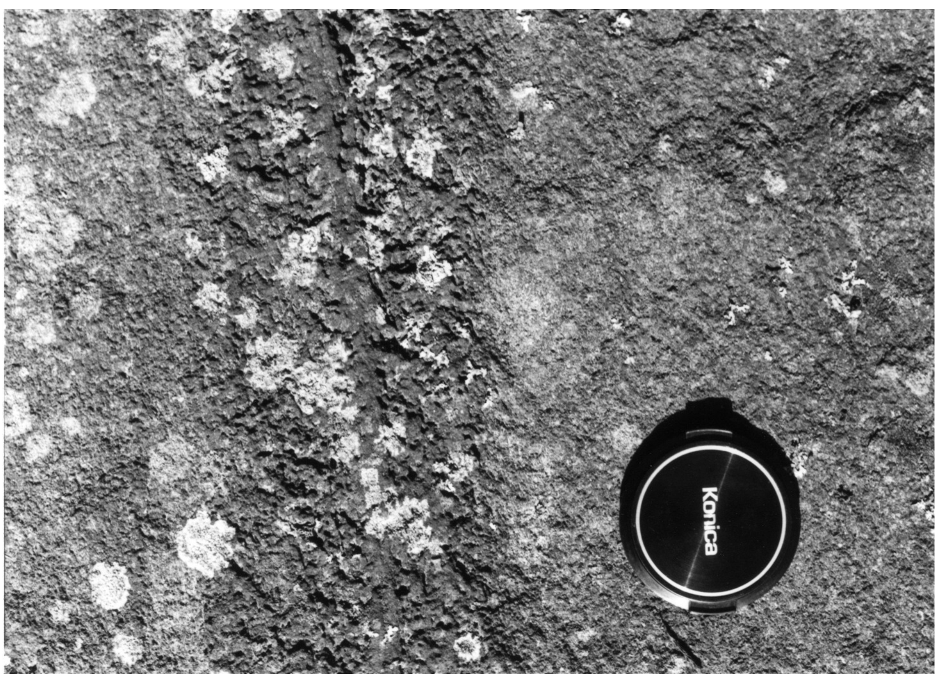

Figure I-11. Outcrop #412. (a) Offset hornblende filled fractures. (b) Closeup showing fracture and mottled diffusion zone.

In the study area, the Diana Complex appears as a porphyroclastic augen gneiss, as it does along most of the northwestern margin. Elsewhere, to the southeast, this grades into a flaser gneiss, and finally to a completely recrystallized granoblastic gneiss. In this transition, it is seen that the perthite-rimmed plagioclase augen recrystallize to microcline plus plagioclase. In the final granoblastic gneiss, no augen are left. Due to the textures involved, the entire Diana Complex is placed within the mylonite zone of Geraughty et al. (1981), and accordingly no protolith is known.

Compositionally, the Diana Complex within the field area is a granitic to quartz monzonitic gneiss. Porphyroclasts are somewhat fractured and recrystallized, and are generally perthitic, but commonly show the core and rim structure of plagioclase and perthite (Fig. I-12). The groundmass consists of medium to fine grain recrystallized plagioclase and microcline. The gneisses generally contain fairly prominant quartz ribbons, which often wrap around the porphyroclasts. Moderate (about 5%) amounts of dark green or blue-green hornblende and a light tan to brown pleochroic biotite are also present, with additional darker biotite. Magnetite occurs and in many places it has altered significantly to hematite, causing a deep red or purplish colour on fresh surfaces. Alteration of opaques to sphene is also common.

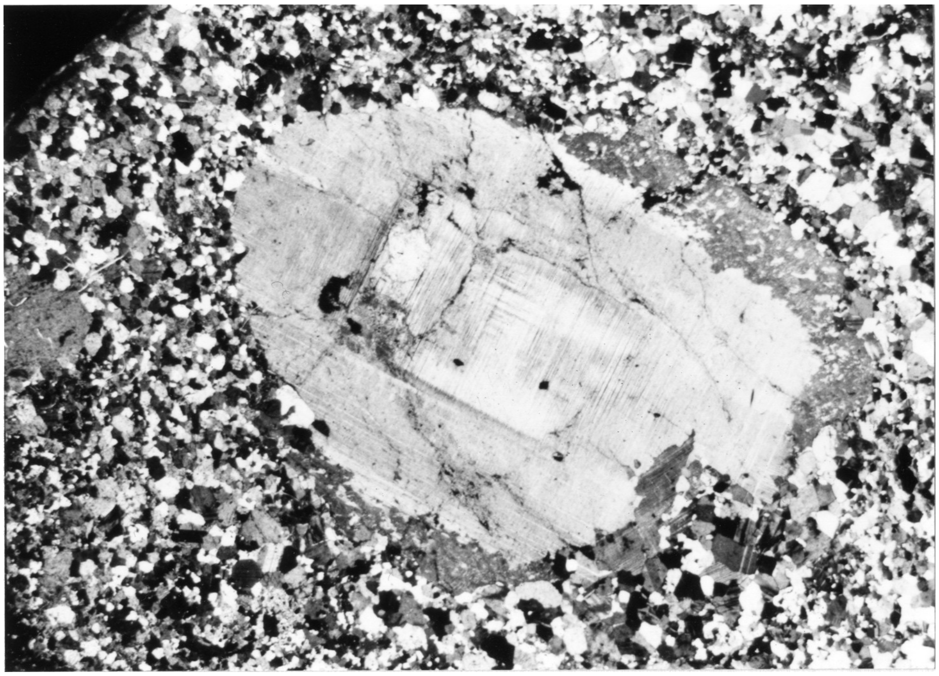

Figure I-12. Diana Complex auge showing plagioclase core, perhaps with some zoning present, and perthite rim. Auge diameter about 1.4 cm..

In outcrop the augen gneisses may weather such that the augen, or quartz ribbons, or both, occupy positions of high relief. This gives the outcrop faces a very distinctive rough appearance. The more eastern rocks tend to show a more pronounced quartz ribbon relief, and this is probably the result of a slightly higher quartz content than their western counterparts.

To the north, the rocks appear to be more drawn out, with narrower quartz ribbons and smoother outcrop faces. Augen are less frequent, and thin sections indicate that the feldspars occur in packets of clean, coarse, polygonal grains of a single feldspar type. The packets are surrounded by quartz, and each respective packet shows a fairly uniform crystallographic orientation of the feldspars, indicating recrystallization and plastic deformation within it.

The Diana Complex rocks around Whippoorwill Corners frequently contain fine grained dark greenish grey areas of roughly equal amounts of augite, microcline, and scapolite. Large amounts of quartz may also be present. These dark zones crosscut the foliation in an irregular, haphazard manner. The borders may be quite irregular as well, or they may be extremely sharp. The origin of these zones is unclear.

Outcrop #195 contains a layer of fine felsite of a composition much like that of the country rock. Fine quartz ribbons were present, and the layer probably parallels the foliation.

Diana Border: Along the western border of the Diana Complex in the study area, the rock has a more biotitic, finer grained, and more foliated appearance. The porphyroclasts are smaller and rarer, although they still exhibit the core and rim structure. Quartz ribbons are also smaller than the average, and quartz ganerally seems less evident in the border zone than in the rest of the Whippoorwill Corners Diana. The biotite has a light tan to brown pleochroism, and grungy dark hornblende is occasionally seen. Magnetite is considerably altered to sphene. The boundary between this fine grained biotitic border zone and the larger porphyroclastic Diana rocks is a gradational one, and the smaller grainsize and larger percent of recrystallized groundmass in the border zone suggest the change is due to increased strain in the rocks near the western margin of the Complex.

Calcsilicates:

Mixed Calc-Silicates: Most of the calcsilicates in the area appear as layered rocks of variable composition. Layers are usually one to five centimeters in thickness, and weathered surfaces show a corrugation corresponding to the varying degrees of resistance to weathering shown by adjacent layers. The layers are usually continuous and planar, but some pinching of layers and minor folds may occasionally be observed (Fig. I-13).

By far the most common layers consist of combinations of one or more of potassium feldspar, plagioclase, biotite, and clinopyroxene. If potassium feldspar is present, quartz may also appear. Hornblende is rare, but may appear if clinopyroxene is present, and it is probably a retrograde product. Which minerals are included in a given layer, and their relative amounts, grain size, and grain relationships seem to follow no clear pattern.

Perhaps the most distinctive, and fortunately also a fairly common member of this layered belt is a fine grained dark greenish to purplish rock. It is most common in the northwest portion of the field area, and consists of pale to dark brown pleochroic biotite and olive to blue-green hornblende, mixed with untwinned potassium feldspar and lesser amounts of microcline and plagioclase. It is more resistant to weathering than many of the other layers, and is easily noticed at outcrops.

Within this general calcsilicate unit, there are also included rare layers of scapolite plus clinopyroxene, amphibolite, pyroxenite, impure quartzites, and various marbles which usually contain some diopside or augite. Additionally, larger, discontinuous zones of quartzite or marble appear as hills scattered throughout the calcsilicate belts.

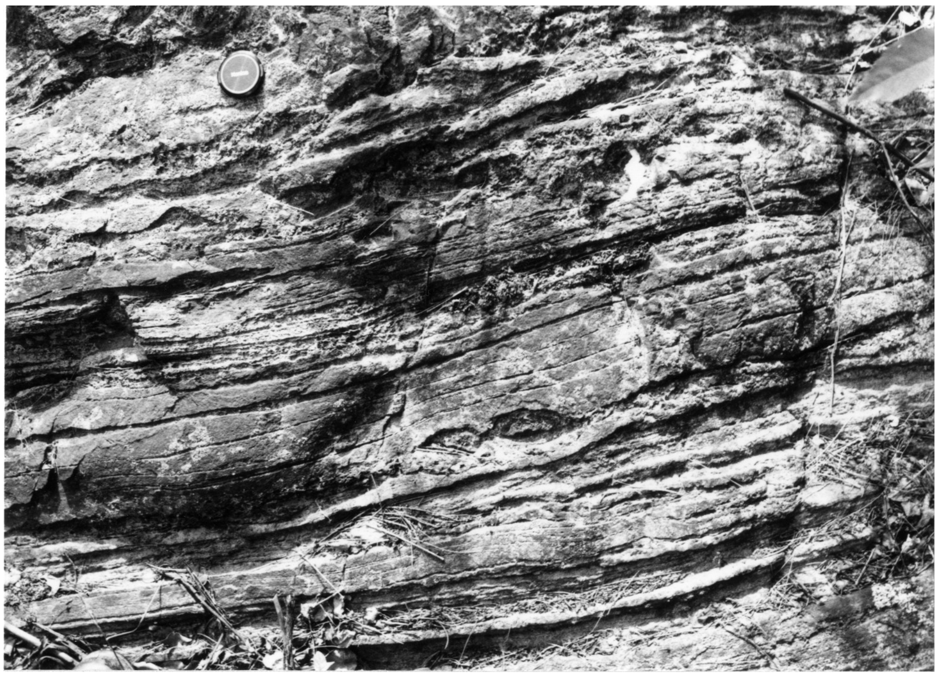

Figure I-13. Mixed Calcsilicates, Outcrop #18.

Rusty Calcsilicates: Another distinctive unit in the calcsilicates is a rusty weathering gneiss. Indeed, the appearance of this unit and its resistant nature caused early geologists to call the entire calcsilicate formation the "Rusty Gneisses" (Miller, 1921b; Dale, 1934).

The rock may weather to either a black or rust colour and contains small pyrite or magnetite grains, the latter being invariably rimmed by hematite, and either of which may produce the weathering colouring.

Several types of rocks are actually included in this unit. Perhaps the oddest looking of these is a rock which, when weathered, is full of apparent vesicles such that it takes on a volcanic appearance. It is made up of augite and quartz layers which pinch and swell on a scale of centimeters, effectively enclosing augite packets in envelopes of quartz. The rock sometimes has minor scapolite or calcite, and the vessicles are caused by weathering out of pyrite.

A variation of this contains very large biotite grains which themselves show the pinch and swell structure on a microscopic scale. As in the quartz-augite rock, the quartz forms envelopes about the mafic minerals.

Another common rock type in the rusty calcsilicate unit is a syenitic gneiss. This is light pink in colour and contains roughly equal amounts of plagioclase and potassium feldspar. The latter includes both orthoclase and microcline. Very minor quartz may be present, and some clinopyroxene is also evident. Small magnetite and hematite grains impart the rusty weathering.

Feldspathic Quartizte: The final mappable member of the calcsilicates is a feldspathic quartzite. This appears layered in outcrop, and this is probably due to changes in grainsize rather than in composition. The rock is more than 50% quartz and the predominant feldspar is potassic. Miller (1921b) reported that several percent of the rock was pyrite, however, hematite is more abundant. Biotite and sphene may also occur, and a bit of garnet, pale green in plane light, was seen in one thin section.

The very fine grain size and high quartz content cause conchoidal fracture when struck. Coarser grained parts of the unit, however, weather to a crumbly white to orange rock, even on fresh surfaces (Fig. I-14). These coarser rocks often have a relatively high mafic content, and the opaque minerals weather out to leave small round voids in the rock.

Belleville Monzonitic Gneiss:

The westernmost portion of the study area is occupied by a gneiss of monzonitic to quartz monzonitic composition. Plagioclase makes up slightly more than 50% of the total feldspar, and it is accompanied by orthoclase with minor microcline and perthite grains. Quartz is rare, but when it

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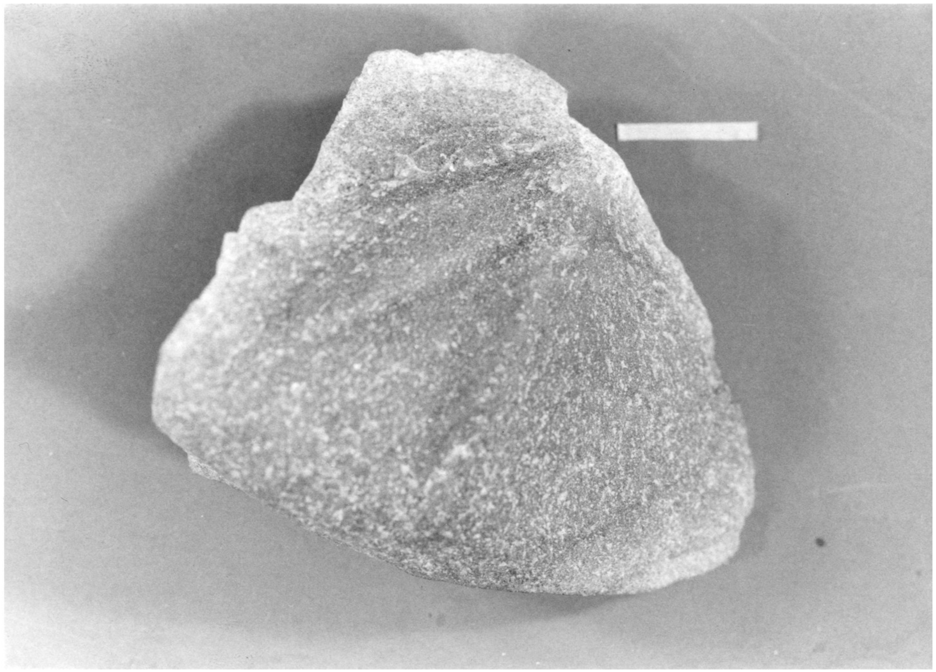

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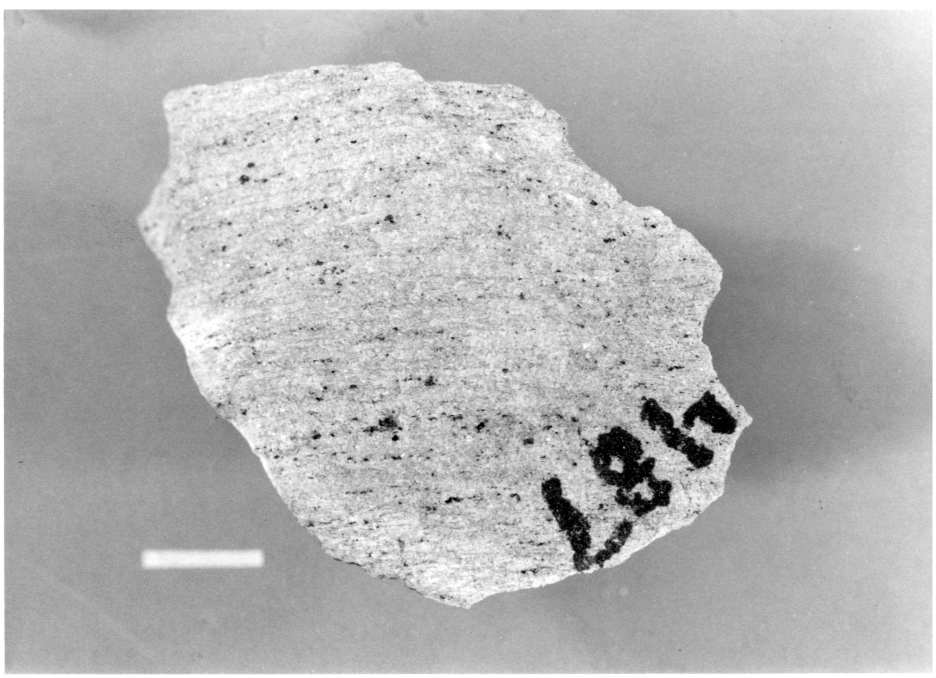

Figure I-14. Feldspathic Quartzite. Scale bars are 2 cm.. (a) Fine grained to crystalline facies. (b) Coarser facies. Dark spots are from weathering out of opaque minerals.

occurs it is generally found as fine grained recrystallized ribbons. Varying amounts of biotite occur, usually characterized by a nearly clear to tan pleochroism, but with darker varieties as well. The biotite seems to control the grainsize of the rock, with larger amounts of mafics in finer grained rocks. Accordingly, both grain size and colour vary considerably. Magnetite or ilmenite is often present and is generally substantially altered to sphene. It may also weather to hematite, producing round rust spots on fresh surfaces, and a black or rust colour on weathered surfaces. This weathering appears more commonly in more mafic rocks, and it gives the outcrop an uneven garbled appearance. In mafic poor varieties, the rock has a white to light pink weathering reflecting the high plagioclase content. In these cases, the weathered outcrops have a smooth, columnar appearance.

Some purple fluorite may be found in these rocks, prompting Buddington and Leonard (1962) to correlate them with the eastern granitic gneiss. While this is possible, the generally higher mafic content and lower quartz in the Belleville would seem to allow separation of them as mappable units.

The Belleville is often associated with a rusty pyritic quartzite or a pyritic quartz-feldspar rock. These frequently have a high biotite content, and may include some calcite. They are generally fine to medium grained and are not areally extensive, nor are they continuous .

Marble:

The western border of the map area roughly follows the easternmost contact of a calcareous marble. This frequently has a substantial amount of sphene within it. Also associated with it is a slightly calcareous plagioclase-minor potassium feldspar-biotite rock. Miller (1921b) described the unit as a medium grain white to grey limestone with interbedded layers of pyroxenite or quartzite. Areas rich in serpentine are also found. The serpentine in these places weathers to form whitish lumps on the outcrop surfaces. According to Buddington and Leonard (1962), disseminated graphite, pyroxene, and scapolite are also found in this unit elsewhere.

CHAPTER THREE - STRUCTURE

Field scale structures in the Whippoorwill Corners area are most obviously characterized by a northerly striking, northwest dipping foliation of varying intensity, and a down dip lineation. The foliation parallels the lithologic contacts and in the viscinity of the Highland/Lowland Boundary the intensity of foliation and lineation increases.

Foliation and Folding:

Closer scrutiny reveals that the foliation reflects several fold events. It has been suggested that the foliation itself is the result of isoclinal folding (Weiner, 1983) and transposition (Granath & Barstow, 1980), probably associated with substantial metasomatism (Engel & Engel, 1958, 1960, 1963; Buddington, 1963). It is also possible that in places layering is primary, as evidenced by stromatolites in the Lowlands (Isachsen & Landing, 1983) and primary igneous textures in the Highlands.

This transposed regional foliation is said to be itself isoclinally folded (Weiner, 1983). Several isoclinal folds having axial planes parallel to the foliation were observed on outcrop scale in the calcsilicate belts (Fig. I-15). Additionally, the calcsilicate belt in the center of the noduliferous microcline granitic gneiss and the one in the Diana Complex may be thought to result from such an isoclinal fold event. Likewise, the rusty calcsilicates, flanked by mafic calcsilicates may mark the hinge of such a fold.

Cross sections of the area suggest a later asymmetric upright to overturned tight folding event (Plate 2). Axial planes of these folds, like the axial planes of the isoclinal folds, strike north-northwest to north-northeast and dip moderately to steeply to the northwest.

Outcrop scale examples of these folds were observed in two locations. Both small scale tight to open folds show axial planes as described above, and hinge lines plunging moderately to the northwest (Fig. I-16).

A final fold generation is observed on a large scale on the map, and is reflected by a stereoplot of foliations (Fig. I-17). This open, upright fold warps all previous foliations and folds into the weak girdle seen on the stereoplot, and it may also be responsible for a slight spread in the composite lineation point maximum. The axial plane of this fold appears to be roughly upright striking to the northwest, and the hinge line appears to

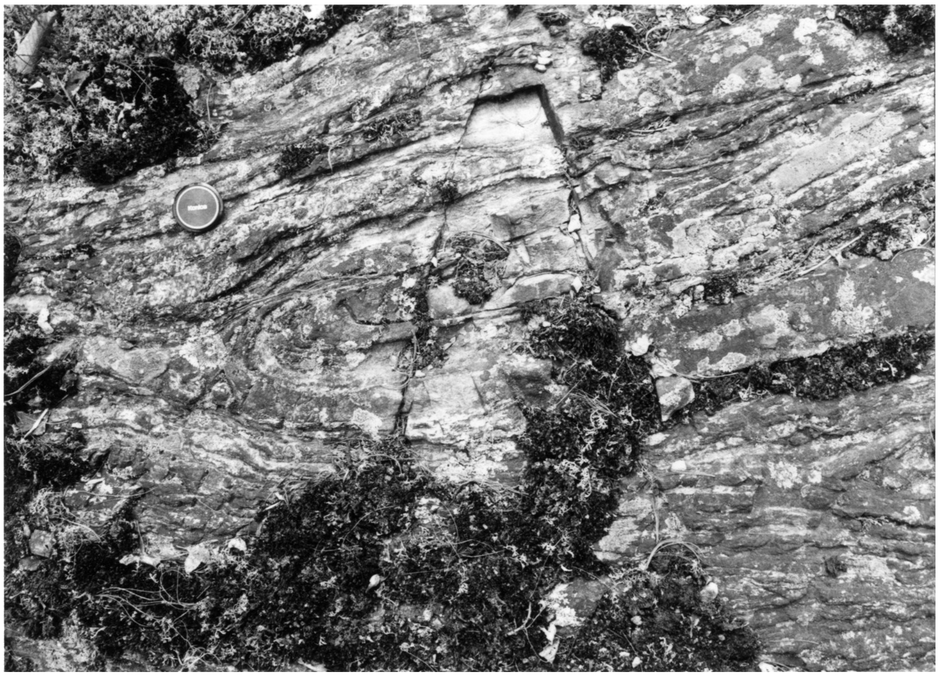

Figure I-15. Isoclinal folds in Calcsilicates, outcrop #18.

Figure I-16. Tight folds in Calcsilicates from outcrop #979.

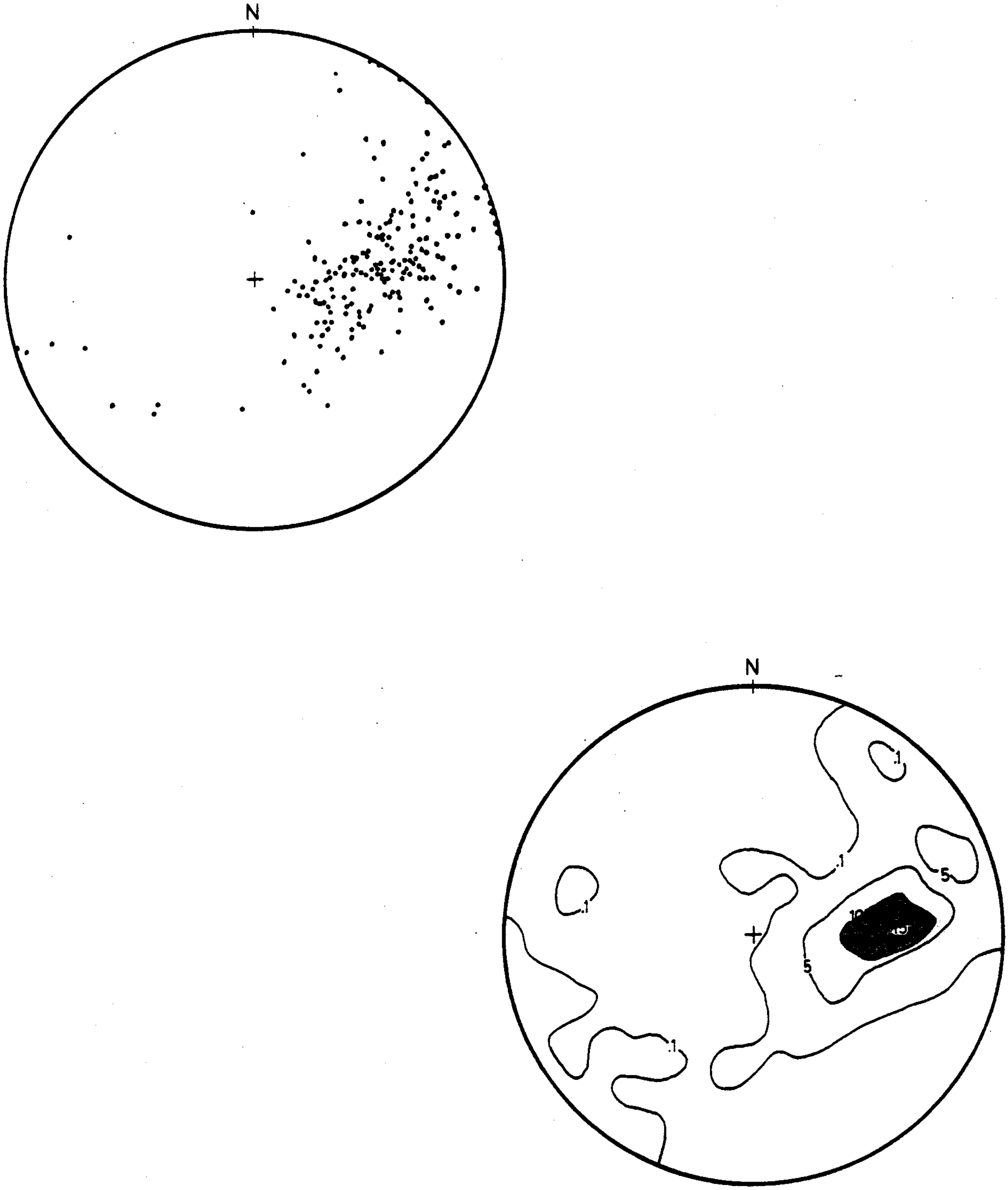

Figure I-17. 182 poles to foliation. Contours are .1, 5, 10, 15% per 1% area.

plunge moderately to the northwest.

Lineations:

Stereoplots of the hinge lines of major and minor folds in the Whippoorwill Corners area show a remarkable parallelism of these lineations. In addition, it is observed that stretching lineations, platy mineral long axis orientations, and the long axes of ellipsoidal nodules in the noduliferous microcline granitic gneiss, all tend to plunge at roughly 30 to 40 degrees to the northwest (Figs. I-18, I-19).

Furthermore, the orientation of these lineations is nearly constant at a given location, regardless of the degree of foliation development, and hence, regardless of the amount of deformation suffered by the rock. This constant orientation on a small scale seems to mimic the larger scale patterns as well. Just as in outcrop, increase in deformation intensity toward the Highland/Lowland Boundary is not accompanied by a change in lineation orientation.

This unvarying lineation orientation is contrary to standard models of high strain zones, as is the unchanging foliation (i.e. Ramsay, 1980). It would seem to indicate either a very high strain throughout the area or a coaxial deformation. The presence of ophitic textures in the anorthositic gabbro might argue, at least locally, against

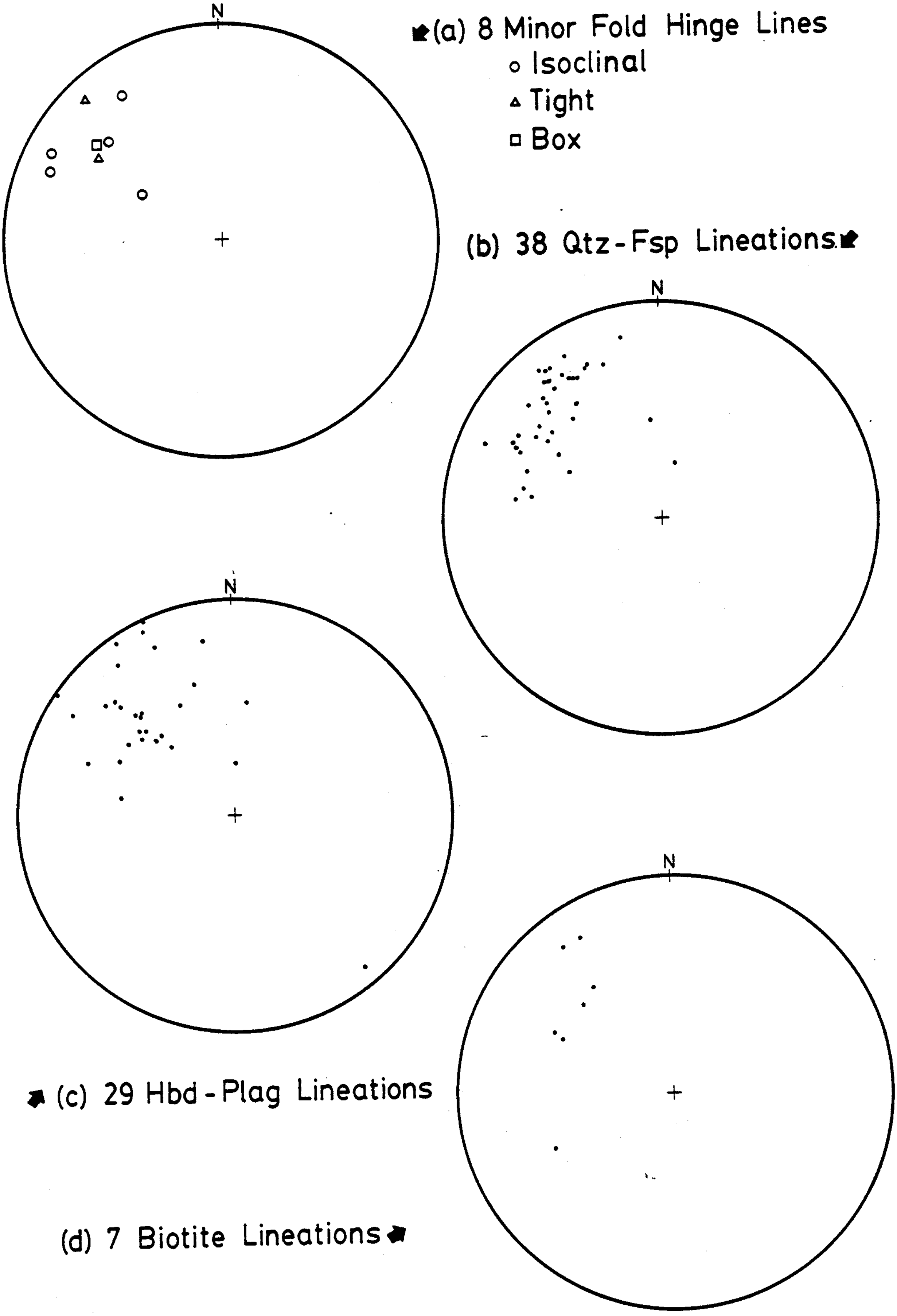

Figure I-18. Lineations in the Whippoorwill Corners area.

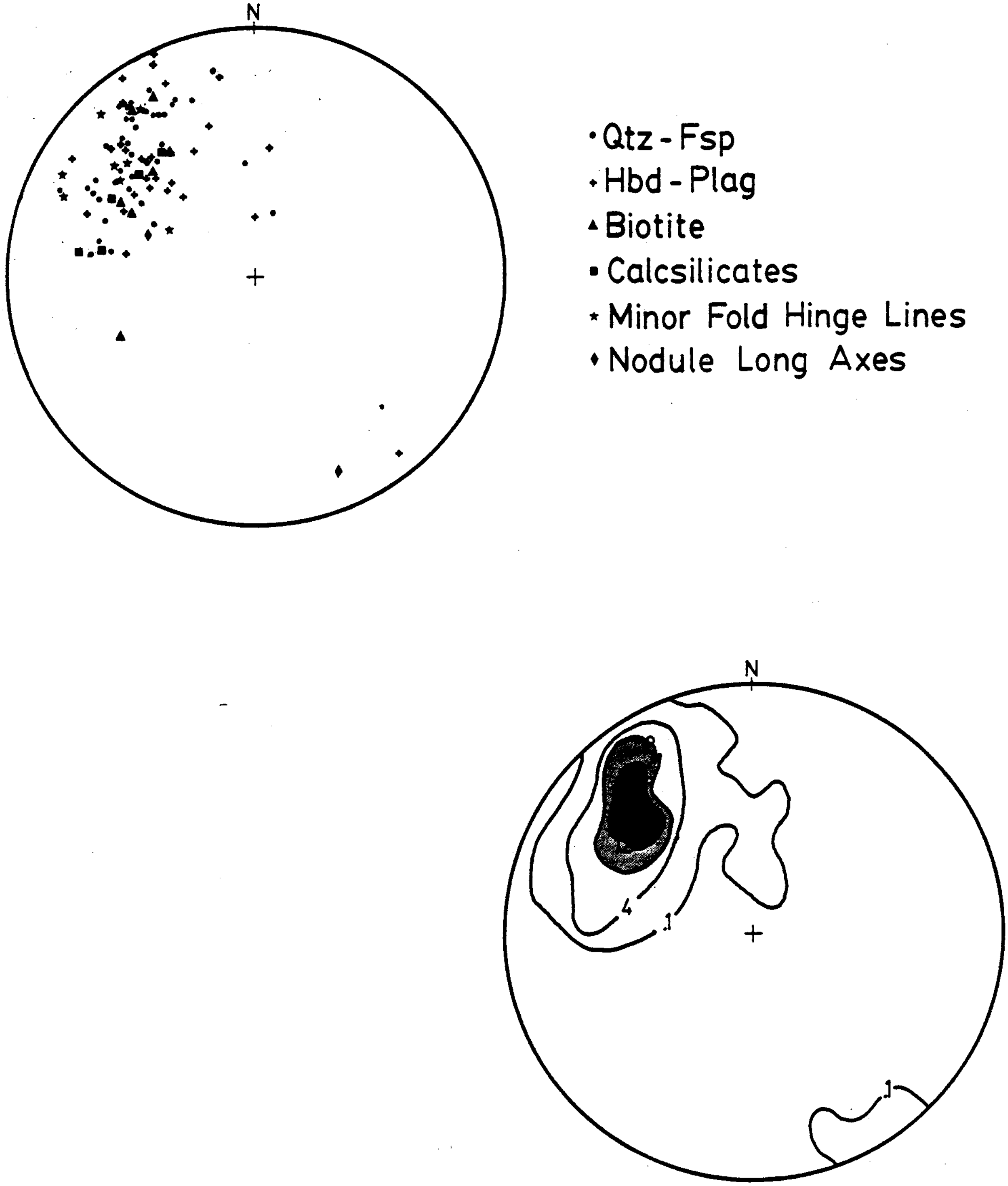

Figure I-19. Composite of all lineations in the Whippoorwill Corners area. Contours are .1, 4, 8, 12% per 1% area.

the former.

Fabrics:

A clear increase in intensity of deformation as the Boundary is approached is evidenced by both the increase in the intensity of development of foliation and lineation, and by concurrent changes in microfabrics, however. As figure I-20 illustrates, the pattern of fabrics may be somewhat more complex than that mapped by Geraghty et al., but their reconnaissance estimates were essentially correct. It appears that a zone exists in the center of the field area in which the rocks commonly display a more apparent foliation and lineation, and a finer grain size and more recrystallized fabric than their counterparts farther from the Highland/Lowland Boundary. The minerals within this central zone also often display a moderate to strong preferred crystallographic orientation. Quartz is commonly more undulose, and it usually forms ribbons rather than disseminated grains.

Within this more deformed zone, however, there exist lozenges of less deformed rock in which the fabrics are roughly analagous to those in the rocks surrounding the central high strain zone. These lozenges range in size from centimeters to meters in outcrop, and the map pattern of fabric intensity suggest they also occur on kilometer scale.

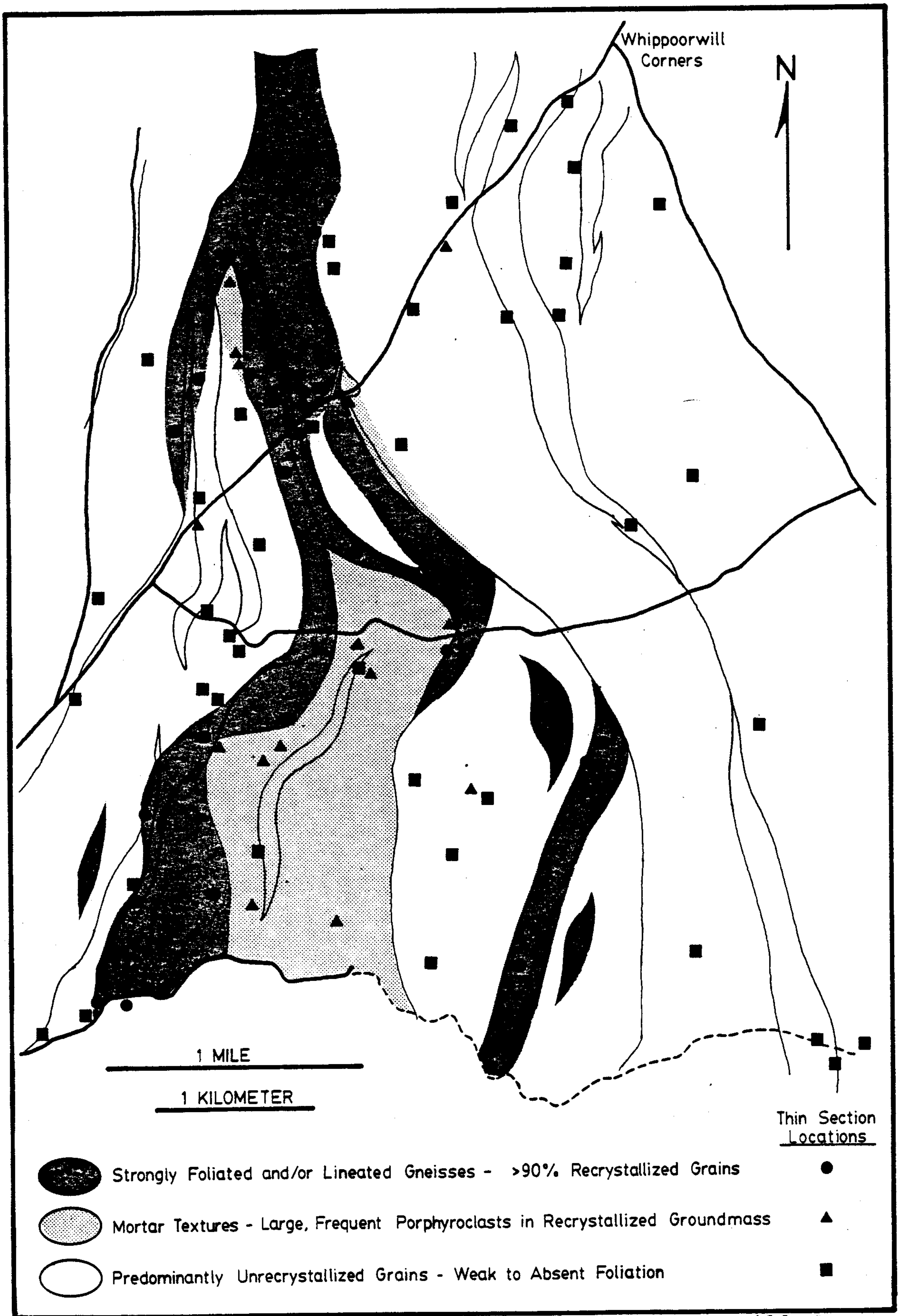

Figure I-20. Map of fabric types in the Whippoorwill Corners area. Data is primarily from thin sections, but is refined by field observation of foliation and lineation.

The boundaries of these lozenges, on outcrop scale, are sharp. Transition from relatively unfoliated to highly foliated and lineated rock is extremely abrupt and correspondingly no sense of asymmetry of foliation is readily apparent. Likewise, the lozenges do not seem to have a strong shape asymmetry.

<u>Brittle-Ductile</u> <u>Fractures</u>:

Apparently associated with the ductile deformation of the Whippoorwill Corners area is an odd, curved fracture system. These fractures occur most frequently in the highly foliated portions at the northern end of the anorthositic gabbro massif. In some places in these finely laminated, grain size reduced areas, the outcrop face is covered with evenly spaced (5 - 20 cm spacing), smoothly curving fractures (Fig. I-21). These fractures are oriented roughly normal to the regional and outcrop lineation direction, and were a primary topic of consideration in Miller´s 1921 investigation of the mafic body. As Miller observed, they occur primarily in the lighter coloured, more anorthositic members of the massif. Moreover, they are invariably in the finer grained portions of these rocks. Miller also observed that the cracks were filled with either fine grained plagioclase or hornblende. It actually appears that scapolite is more common than plagioclase, and that when present, the hornblende usually plates the sides of scapolite

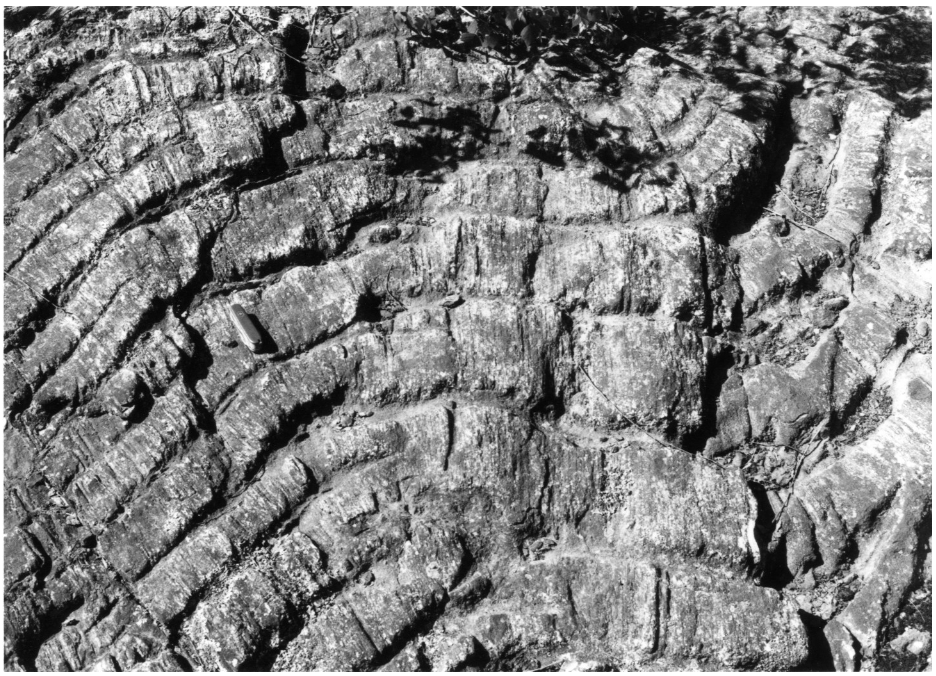

Figure I-21. Curved fractures in the Anorthositic Gabbro, outcrop #448.

cored fractures. This causes low relief on the outcrop faces as the scapolite is easily weathered, but upon sampling, hornblende bordered rocks are frequently retrieved. Buddington (1939) also looked at these odd fractures and found that the hornblende is oriented normal to the walls of the cracks. This, he suggested, is analagous to fibrous vein fillings of lower grade rocks. The normal relationship to the walls, however, also happens to parallel the lineation in the country rock, and may therefore be a mimicing of a preexisting fabric.

These fractures are not exclusively related to the anorthositic gabbro. In an outcrop of syenite just across the northern border of the mafic unit (Outcrop #506), curved fractures occur. These were also investigated by Buddington (1939) and by Miller (1921a) and were found to consist of augite. This, Buddington points out, suggests brittle-ductile deformation in the granulite facies, or a high temperature hydrothermal fluid filling the fractures.

Rare curved fractures have also been observed in the calcsilicate belts just west of the anorthositic gabbro and in the southeastern part of the field area.

The significance and formational mechanism of these fractures remains unexplained. Their occurrence in only highly strained rocks, and the general similarity of their mineralogy to that of these strained rocks, suggests that they originated synchronously with the foliation development. On the other hand, the relatively undeformed nature of the fractures would seem to indicate that they occurred very late in, or after, the shearing event. It is likely that the fractures did exist during the shearing, as in some instances the country rock foliation may be observed to bend into the fractures in a manner resembling simple shear zones. This shear appears to be of opposite sense on opposite limbs of folds which bend the fractures, and the shear senses resemble what might be produced by flexural slip during fold development. Such fold development would require slip along foliation planes.

It seems likely that a shear-related extensional mechanism played a role in the fracture formation. At the top of the hill just north of Route 87, a small set of fractures may be observed which, at their ends, form a 45 degree angle with the foliation. These fractures curve to an orientation normal to the foliation (Fig. I-22), and further along the zone, the central part of the fractures is oblique to the foliation, in a sense opposite to that at their ends. This morphology is quite similar to quartz filled extension fractures observed in lower grade rocks. If such brittle fracturing ocurred at the high temperatures at which deformation ocurred, an extremely high pore fluid pressure might be suggested.

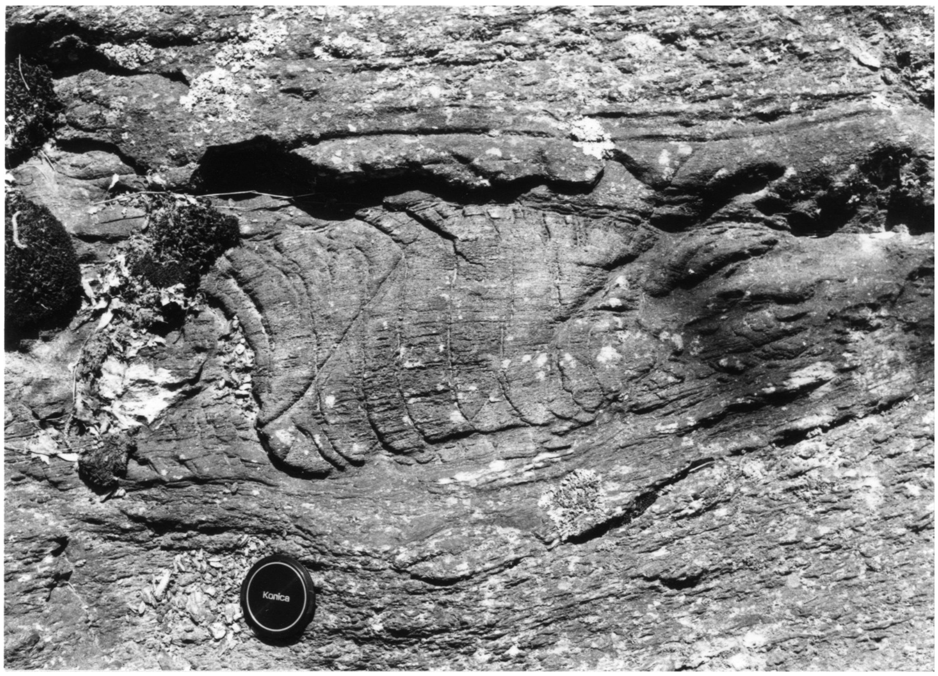

Figure I-22. Extension fracture-like fractures in the anorthositic gabbro, outcrop #448. Note change in orientation of middle section of the fractures laterally along the zone.

CHAPTER FOUR - SENSE OF SHEAR

The origin of the high strain fabric, as mentioned previously, is unclear. Most authors since Buddington and Leonard (1962) have agreed that differing competencies of adjacent lithologies has played an important role in concentration of strain, however, they have disagreed on what this role is. Buddington and Leonard envisioned isoclinal folds growing around and upon rigid units in the Adirondacks, and that the foliation and steeply plunging lineations reflected movement of the fold against the rigid bodies. In the case of the Highland/Lowland Boundary, the quartz-syenitic gneisses of the Diana complex would, by this hypothesis, behave as a more rigid unit than the Lowlands metasediments. Thus, Buddington and Leonard do not discuss the zone as a major structural discontinuity, but rather as a rheologic and topographic boundary, the detailed nature of which is unspecified. In 1939, however, Buddington had discussed the boundary area as being a thrust of Lowlands over Highlands rocks.

Leitzke (1973) and Weiner (1981) also imagine the high strain zone to be the effect of competency contrasts. Leitzke, however, claims the more deformed zone marks either a strike-slip or thrust zone, whereas Weiner states that continuity of folds across the boundary and correlation of xenoliths within the Diana Complex with adjacent metasediments negate the possibility of substantial offset. Rather, Weiner believes the foliation to reflect shortening normal to the axial planes developed during large scale isoclinal folding. It should be noted, however, that Weiner´s mapped area includes only a narrow section of the edge of Geraghty et al.´s mylonite zone. The zone itself is several kilometers wide in the Harrisville-Pitcairn area, and strains reflected at the boundary could be negligible while strains across the whole zone could be tremendous.

Like Weiner, Geraghty et al. (1981) suggested that the foliation was related to the isoclinal folding in the area, but they hypothesized that it might be related to simple shear along the limb of a southeast directed isoclinal fold nappe.

Finally, Whitney speculated that the Carthage-Colton zone and other Adirondack high strain zones might be the present expression of plate boundaries from a Grenvillian Tibetan style orogeny. In this scenario, the Lowlands would have been thrust over the Highlands from the east, such that the westward dipping zone which is presently exposed is tilted from its original orientation and appears as a northwest side down normal fault.

Thus, proposals suggest that the region should show northwest up or northwest down motion, or strike-slip movement, or no movement sense at all. No definitive proof of any type has been shown using meso- or macroscale features of the Boundary zone. However, stretching lineations generally plunge to the northwest, suggesting that the motion was not strike-slip, as this would require north to northeast lineations. To attempt to constrain the possibilities of type and sense of shear, observation of microscopic structural features was undertaken with the hope of recognizing microstructural asymmetries such as those described by Simpson and Schmid (1983).

Eighty-eight thin sections were examined, each cut normal to foliation and parallel to lineation, and taken from rocks within and near the high strain zone. Additionally, polished serial slabs were analyzed from twenty of the more highly deformed and recrystallized rocks. In all, ninety rocks were cut and studied, and eighteen showed some sense of asymmetry. Perhaps the most striking feature of the suite, however, was the remarkable symmetry which was displayed in nearly all samples, and which often dominated even those rocks which are ranked as showing asymmetry.

Augen, for example, are nearly always symmetric in thin section (Fig. I-23a). Only two of hundreds observed had clearly asymmetric tails. Slabs more frequently yield asymmetric augen, probably by virtue of the greater number

(a)

(b)

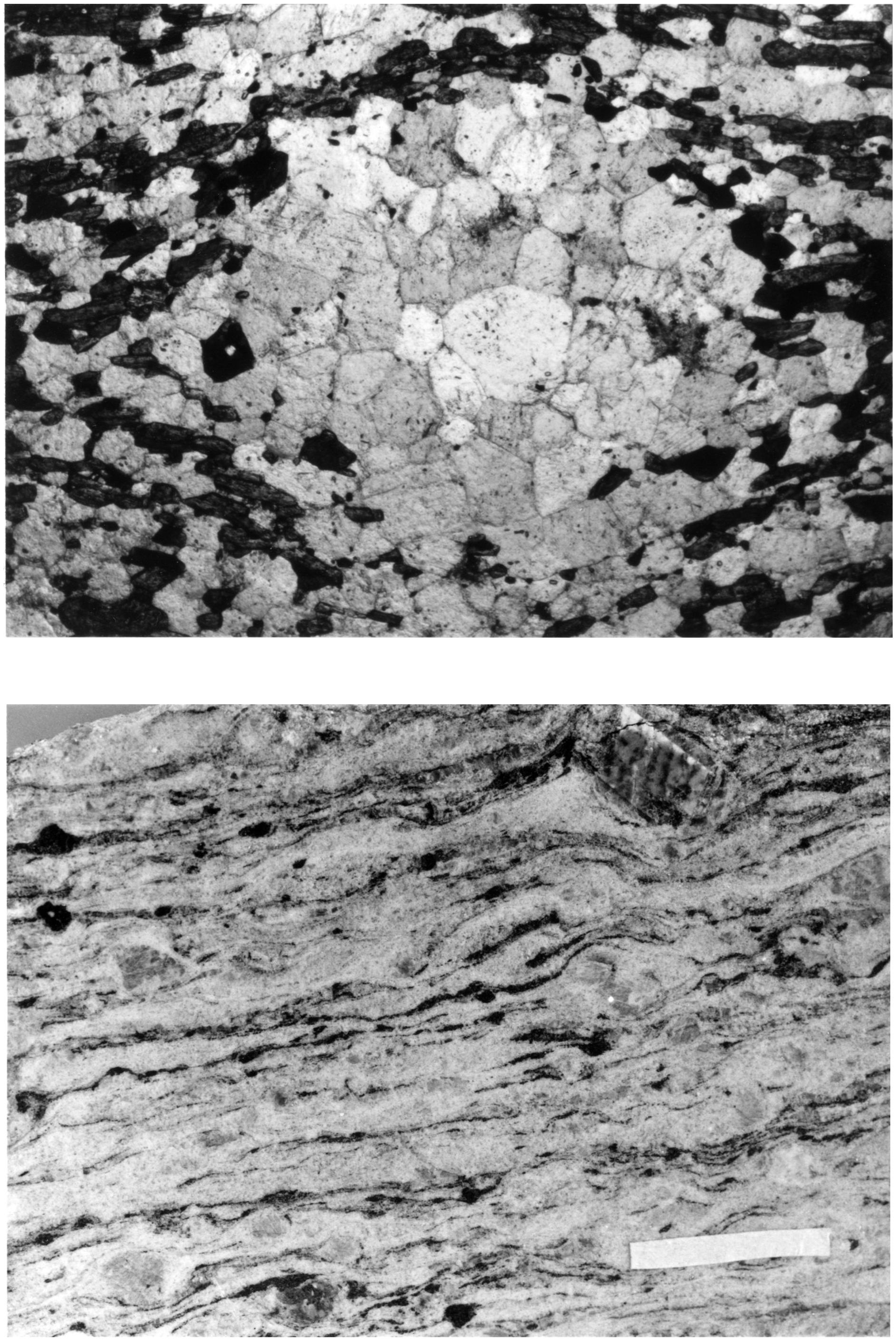

Figure I-23. (a) Thin section of nearly symmetric auge in anorthositic gabbro. Long dimension of photo is 3.8 mm.. (b) Asymmetric porphyroblasts in slab of the Diana Complex. Scale bar is 2 cm..

observable (Fig. I-23b). An estimated 2 - 3% of the augen in slabs show recognizable asymmetry. Of these there is about a 68% consistency in shear sense within a given rock. Despite this low frequency and questionable reliability, augen asymmetries are the most abundant, most easily obtained, and perhaps the most reliable data obtainable in the Whippoorwill Corners rocks.

The only other proven sense of shear indicators were oblique foliations observed in two thin sections. Although not terribly clear, it appears that some small hornblende grains lie at a very low angle to the predominant hornblende foliation (Fig. I-24). These two foliations are assumed to be analogous to the S and C foliations of Berthé et al. (1980). This assumption is based upon a correlation with a shear sense suggested by augen asymmetries seen in slabs corresponding to these thin sections. In both cases the foliation and augen asymetries suggest northwest side up motion.

Two additional, heretofore undescribed, possible shear sense indicators were observed. The first of these, illustrated in Figure I-25, involves asymmetry of hornblende grains. In this instance, although grain shape orientations appear to be invariant within a thin section, it seems that a crystallographic preferred orientation may exist. When a thin section which is cut normal to the foliation and parallel to the lineation is placed under a

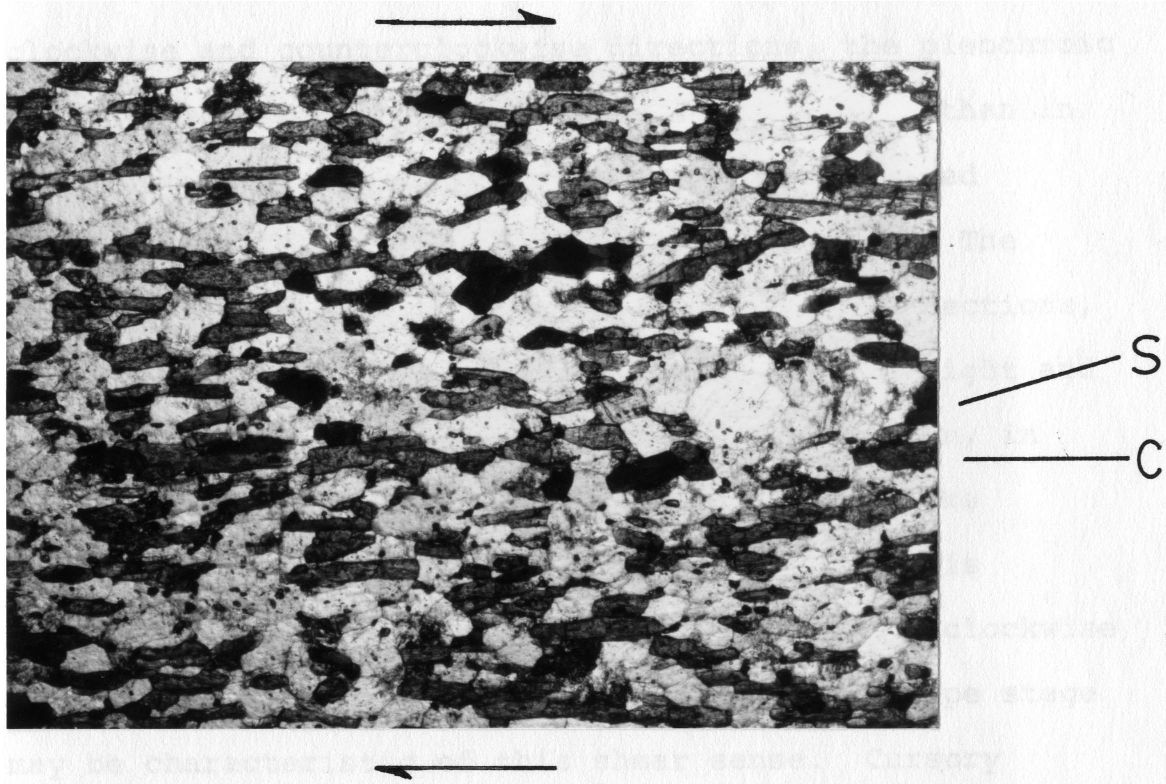

Figure I-24. Two hornblende foliations in Anorthositic Gabbro. The oblique foliations are considered to be analagous to C and S fabrics of Berthé et al. (1980). Photo suggests dextral shearing. Long dimension of photo is 3.8 mm..

single polarizer such that the plane of the foliation is parallel to the plane of polarization, and the thin section is subsequently rotated equal amounts in both clockwise and counterclockwise directions, the pleochroic field assumes a lighter colour in one direction than in the other. This would seem to suggest a preferred orientation of all of the axes of the crystals. The pleochroic asymmetry was observed in five thin sections, and four of these showed a consistency between light and dark pleochroism, and thus hornblende orientation, in geographic space. Augen in several of these rocks indicate a probable dextral shear sense, and it is therefore suggested that extra brightening upon clockwise versus counterclockwise rotation of the microscope stage may be characteristic of this shear sense. Cursory universal stage observations suggest that slip on the [100] (001) system occurs, and that the system slips preferentially such that the acute angles formed by the intersection of the <c> and <a> axes is opposite the shear sense (Fig. I-25).

Alternatively, of course, a preexisting preferred orientation may be responsible for the feature, but no such preferred orientation is readily apparent in undeformed rocks, such as the anorthositic gabbro, in which the feature commonly occurs.

The second previously undescribed feature which may

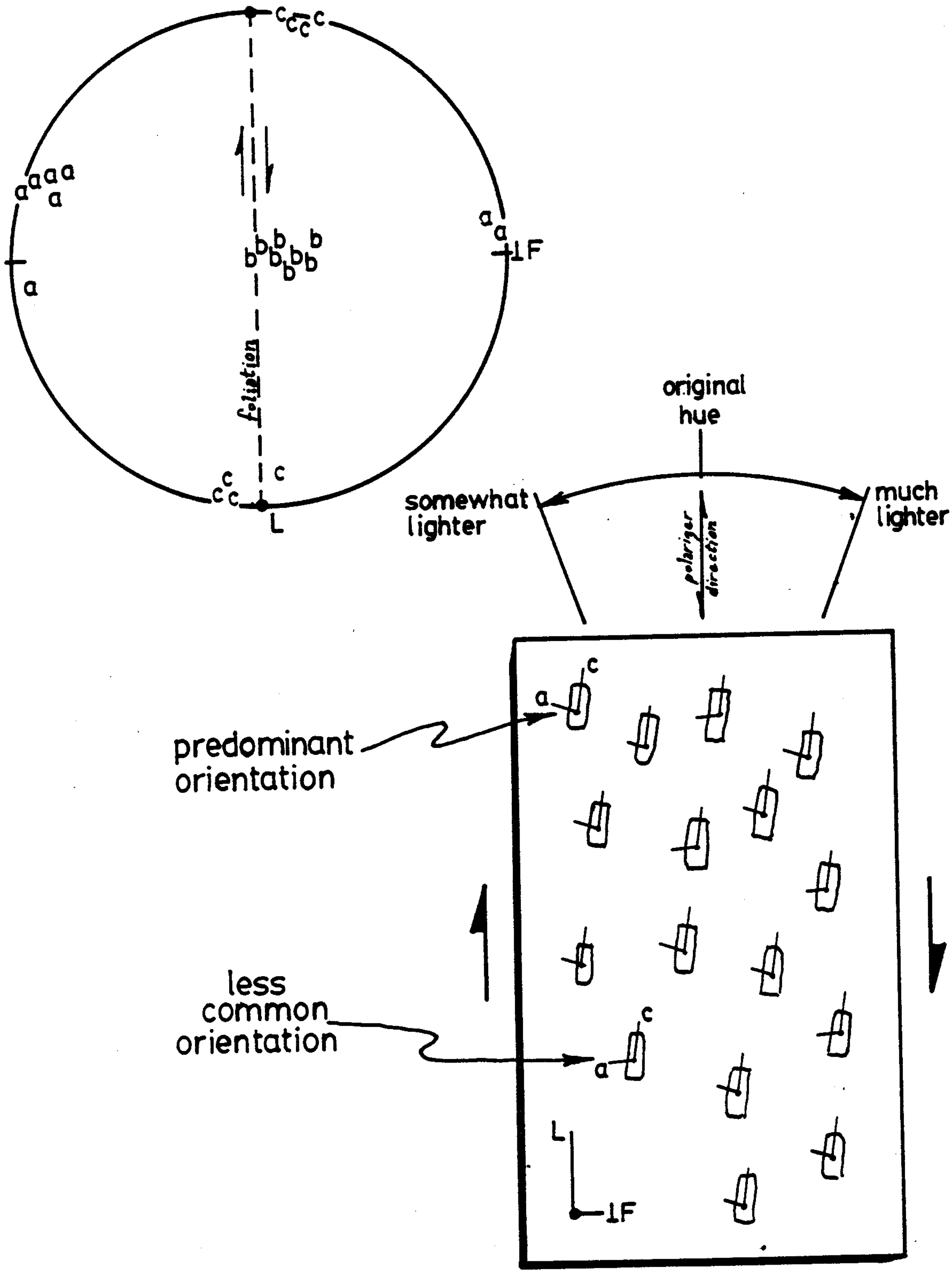

Figure I-25. Preferred crystallographic orientations of hornblende. Stereonet is representative lower hemisphere projection illustrating orientations as represented in the schematic thin section. L is the lineation direction, ⊥F is the pole to foliation. Rotation of thin section produces different degrees of pleochroic changes as a result of the preferred orientations.

reflect the sense of shear occurs in quartz ribbons, usually in granitic gneisses. Many of these gneisses contain a well developed ribbon lineation made up of coarse apparently unstrained quartz grains. In some samples, however, later elongate subgrains or new grains may form within the original ribbons. These new subgrains and grains are oriented obliquely to the original ribbon, and although the shape of the original ribbon is preserved, the new grains are essentially new quartz ribbons.

Within a thin section, the direction of obliquity of these secondary ribbons is remarkably consistent, and comparison with augen in one sample exhibiting the new ribbons suggests that the obtuse angles between the primary and secondary ribbons mark the shear sense (Fig. I-26).

The oblique secondary quartz ribbons just described nicely illustrate the profound effect of the last stage of shearing on sense of shear indicators and on the microstructure. It appears that quartz ribbons had already developed before the secondary ribbons formed. The deformation event which produced the primary ribbons was the event responsible for other lineations and foliations in the area, and the secondary lineations overprint this. Thus, they may reflect continued deformation, or different strain or strain rate than the

Figure I-26. Oblique secondary quartz ribbons. Foliation of rock is parallel to the quartz layers. The new subgrains/grains are oblique to this. Inferred sense of shear is dextral. Long axis of photo is 3.8 mm..

surrounding rocks, or they may be completely unrelated to the deformation responsible for the predominant features in the area.

In summary, then, of the ninety rocks studied, fifteen showed some asymmetry which was apparently produced by deformation with some simple shear component. Ten of these rocks indicated a northwest down shear sense, and the remaining five suggested the movement was northwest side up (Table I). On the average, a 68% consistency was observed within a given rock with respect to sense of shear indicators such as rotated augen. More pervasive indicators such as foliations were more consistent within a rock but were more difficult to identify and interpret independently of other indicators.

Generally, then, the vast majority of rocks showed no evidence of non-coaxial straining. Those that did showed a slight bias toward normal fault motion along the boundary. This suggests that either the rocks are all so highly strained that fabrics have lost their asymmetry, or that they have undergone a basically coaxial deformation, such that no asymmetries developed. The latter interpretation also explains why both dextral and sinistral shear senses were observed. It is easy to imagine an environment of shortening in which slip occurred slightly obliquely to the extension direction as the result of minor fabric asymmetries. It is much more difficult to imagine this occurring in a simple shear environment. Passchier (in press) and Garcia Celma (1983) believe that they see just such a reverse sense of shear in simple shear zones, however.

TABLE I - SUMMARY OF SENSE OF SHEAR INDICATORS

SAMPLE NUMBER	AUGEN SLABS	ASYMMETRY THIN SECT	C AND S FABRICS	HBD PREF ORIENTN	SECNDRY QTZ RIB	SUM TOTAL
AGf467A	NW UP			NW UP		NW UP
AGf467B	NW UP			NW UP		NW UP
AGf467C	NW UP		NW UP	NW UP		NW UP
AGf504Y	NW DN					NW DN
AGf563		NW DN				NW DN
AGf621	NW DN					NW DN
AGf919	NW UP		NW UP	NW UP		NW UP
Dib069	NW UP					NW UP
Dib613	NW DN					NW DN
EGr432	NW DN	NW DN			NW DN	NW DN
EGr748Y					NW DN	NW DN
EGr829Y					NW DN	NW DN
MSy823Y	NW DN					NW DN
CSm530				NW DN		NW DN
Bel792Y					NW DN	NW DN

TOTAL NW UP = 5
TOTAL NW DN = 10

Letters in sample numbers refer to lithologies as on Plate 2. Numbers refer to outcrop numbers as on Plate 3.

CHAPTER FIVE - DISCUSSION AND CONCLUSIONS

The "Carthage-Colton Mylonite Zone", as named by Geraghty et al. (1981), in the Whippoorwill Corners area, is a zone on the order of a hundred meters wide which displays a more intense foliation and lineation than the surrounding rocks. Fabrics within this zone are generally more recrystallized and of finer grain size than fabrics outside the zone, and quartz is more likely to occur in ribbons which show subgrains.

The intensity of deformation fabric, within a very short distance, varies from nearly undeformed rock to strongly foliated and lineated, fine grained, recrystallized rock. These variations form an anastomosing zone of more foliated rock which encloses lozenges of the less deformed material. These lozenges occur on scales of centimeters to kilometers.

The foliation and lineation orientations within and outside the high strain zone do not vary to a measurable degree. All lineations of all types, in fact, are parallel. This seems to suggest that either the entire region has suffered a very high strain, or that a progressive coaxial deformation has occurred such that the direction of maximum finite strain, as marked by the lineations, has remained constant throughout the deformation.

The presence of undeformed rocks in the area might seem to argue against a regional high strain, but the rapidity of change from undeformed lozenges to mylonitic rock could provide an explanation for these undeformed areas. Even in a simple shearing environment, if changes in fabric development occur extremely rapidly, it is possible that intermediate states having different lineation and foliation orientations might not be noticed.

Detailed analysis of meso- and microscale structures fails to significantly clarify the origin of the high strain fabrics. Most fabric features in most rocks in the area are remarkably symmetric, and in those rocks in which some asymmetry does exist, that asymmetry is rarely well developed. In augen gneisses, for example, if any asymmetric tails are observed they usually occur on less than 5% of the augen in a rock. Moreover, within a given rock, or between rocks, there is only about a 67% consistency in indicators which do show some sense of shear. The two-thirds majority of these point to northwest side down (normal) fault motion, however. The sense of shear indicators observed include asymmetric augen tails and multiple foliations, and two previously unreported potential indicators: crystallographic asymmetry of hornblende preferred orientations, and secondary quartz ribbon development oblique to primary quartz ribbons.

The sense of shear information, like the macroscopic data, tends to suggest that deformation in the area is either coaxial, or that the strain is very high in a northwest side down simple shear sense. In actuality, it is likely that some intermediate state of simple and pure shear exists.

A regional pure shear model is somewhat attractive, as it is easy to explain, through minor fabric asymmetries and resulting shear planes, the minor occurrence of both dextral and sinistral shear motion, and the predominance of symmetric fabrics. It should be noted, however, that the phenomenon of reverse shear sense has been reported in zones known to have undergone predominantly simple shear movement (Passchier, in press; Garcia Celma, 1983). The question therefore arises as to which proportions of sinistral, dextral, and symmetric shear indicators may be thought to represent which shear sense, if no megascopic indicators exist.

The timing of development of the high strain fabric is also somewhat enigmatic. Grant et al. (1984) have obtained an isotopic date of 950 Ma from a highly strained granitic gneiss along the Highland/Lowland Boundary. This date is significantly younger than most other dates arrived at in the Adirondacks, and suggests late adjustment by shearing. Weiner (1983) and Geraghty et al. (1981), on the other hand, believe the shearing to be related to an early isoclinal folding generation.

The mineralogy of the more strained rocks tends to exhibit a lower temperature assemblage than the less deformed equivalents. The fine grained northwest border gneisses of the Diana Complex, for example, contain large amounts of light tan to brown biotite and a very small amount of grungy hornblende. The coarser portions of the Diana, on the other hand, consist predominantly of blue-green hornblende, with lesser light tan to brown biotite and some darker biotite. This may suggest retrogression accompanying the grain size reduction.

Similarly, the anorthositic gabbro in its undeformed state is a plagioclase plus pyroxene rock. With deformation, the pyroxene changes to hornblende, and rarely to biotite, and the plagioclase becomes more sodic. Additionally, the sheared rocks occasionally contain small amounts of garnet, usually between what appear to be microboudinaged opaque minerals, and this seems to indicate deformation at temperatures below that of the peak assemblage.

This retrogression probably required influx of some amount of fluid, and this point is underscored by the conversion of plagioclase to scapolite in many of the deformed anorthositic gabbros. This change requires substantial chlorine and carbonate addition.

A final indicator of retrogression during shearing is found in the fractures which are especially common in the anorthositic gabbro, but are also evident in the eastern granitic gneiss and in the calcsilicates. The similarity of the mineralogy of the fractures and country rock, and the slightly deformed nature of the fractures, suggests that they formed contemporaneously with shearing. Their presence probably indicates fluid flow and a temperature in the amphibolite facies at the time of deformation.

All of this would appear to concur with Davis´ (1979) suggestion that the high strain fabric development occurred after the peak metamorphism, and it may be in agreement with Grant et al.´s (1984) young age for the highly strained rocks.

This raises an interesting point regarding the timing of structural development. While the high strain fabric seems to be deformed by all phases of folding it seems to be a late stage event with respect to the thermal history. This is underscored by the presence of secondary quartz ribbons which indicate that no extended high temperature static event postdated the deformation event which created them. Such a thermal event would probably have allowed the quartz to anneal, and would have destroyed the deformed microstructure.

It therefore seems likely that the structures seen along the Adirondack Highland/Lowland Boundary reflect only the very last stages of deformation, while evidence of the events accompanying the early portions of the collision have been obscured by the intensity of these later events. This may help to explain the remarkable parallelism and continuity of all foliations, lithologic contacts, and intrusive margins, as well as the apparent paucity of fault contacts in the area.

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PART II

Grain Boundary Bands:

Possible Evidence of Grain Boundary Sliding

CHAPTER ONE - INTRODUCTION

To fully understand the deformation of rocks at deep crustal levels it is important to know the relative importance of various deformation mechanisms. This problem may be investigated experimentally by simulating the conditions thought to exist deep in the earth, or it may be investigated by direct observation of rocks thought to have been deformed at great depth. To accomplish the latter however, it is necessary to recognize the characteristic signatures produced by different grain-scale deformation mechanisms.

While fracturing, pressure solution, and dislocation climb and creep related structures are relatively well described and easily recognized, features attributable to high temperature diffusive mass transfer and grain boundary sliding (GBS) have been less clearly shown.

That these processes play a role in rock deformations, there seems little doubt. Experiments and observations of naturally deformed rocks have shown that rocks behave in much the same manner as metals, ceramics, and similar materials, and that similar microstructures are developed in different materials which have deformed by similar mechanisms. Thus, the existence of grain boundary sliding in other materials under conditions of high homologous temperature and/or low differential stress suggests that under similar conditions, the process should operate in rocks.

Indeed, grain boundary sliding has been hypothesized since the onset of plate tectonic theories as the operative process in aseismic flow of the upper mantle (McKenzie, 1968). More recently, it has been proposed that grain boundary sliding may play a dominant role in ductile high strain zone activity (White, 1979; Etheridge & Wilkie, 1979).

In order to confidently recognize microstructures relating to GBS in rocks, it is useful to review the textural features and theories recognized by materials scientists, and then to apply these criteria to rock textures in areas thought likely to have deformed by grain boundary sliding.

<u>Metallurgical Ideas Regarding GBS</u>:

Grain boundary sliding has long been recognized in materials science as a mechanism by which crystalline solids deform to extreme degrees without rupturing. This ability to suffer extreme deformation is referred to as superplasticity.

grain boundary sliding itself may be defined as the movement of two adjacent grains past each other by frictional sliding along their boundaries. It is clear that if GBS were to act alone, however, with each grain behaving as a rigid body, voids would open and impingement would occur at jogs in the boundaries and at corners of the grains. To avoid these compatibility problems some form of accomodation mechanism must operate. Metallurgists recognize two major types of accomodating mechanisms: dislocation creep, and diffusion. The former may occur either throughout the grain, or it may be limited to the outer region. Elastic grain distortion, combined with cracking or void formation may also accomodate small amounts of GBS.

Dislocation creep through the whole grain is thought to involve production of dislocations at points of impingement, and movement of these dislocations away from these points. Pileups occur, and climb to sites of annihilation allows continuation of the process and is the rate controlling step (Mukherjee, 1971).

Dislocation creep, when confined to the outer portion (mantle) of grains supposedly operates with movement of dislocations roughly parallel to grain boundaries, and pileups occur at points of high stress. Climb from these points into the boundaries and adjacent mantles allows annihilation (Gifkins, 1976).

Diffusive accomodation may operate by movement of atoms along the grain boundaries (Coble creep), through the lattice (Nabaro-Herring creep), or through dislocation channels in conjunction with plastic deformation (Ashby & Verrall, 1973).

Each of these accomodation mechanisms is thought to be the rate-determining step in the grain boundary sliding process, and each has a slightly different rate in itself. Thus, for a given experiment, under known conditions, the materials scientist can determine the strain rate and its correlative deformation process, and then relate the microstructure to this. Overall, it appears that in metals extensive grain boundary sliding, or superplasticity, occurs in fine grained (.001 to .01 mm) materials at temperatures above one-half the Kelvin melting point. It occurs more readily in two phase materials, and often these are eutectoid compounds such that recrystallization, recovery, and reactions are inhibited. Superplastic GBS tends to yield stable textures of equiaxed grains which show no dislocation substructure. Grain boundaries tend to have a curved aspect, and adjacent grains often show substantial rotation from their pre-deformation state.

In a laboratory deformation where the amount of strain, the straining mechanisms, and the deformed microstructure are known, it is common to attempt to estimate the amount of strain indicated by the deformed microstructure and attribute the rest to GBS. A geologist, on the other hand, must rely heavily on the final microstructural characteristics, and infer the presence, rather than the amount, of sliding on grain boundaries.

<u>Geology and GBS</u>:

Although grain boundary sliding was proposed as a potentially important deformation mechanism in rocks as early as the late 1960´s (McKenzie, 1968), observation of microstructures likely to be indicative of it were not reported until the mid 1970´s. Even these observations tend to point to superplastic flow rather than GBS itself. Although superplasticity is believed to result from extensive grain boundary sliding, it may result from other processes such as dynamic recrystallization during power law creep (Schmid, 1982).

Probably the first report of evidence pointing to superplasticity and thus indirectly to GBS was made by Boullier and others (e.g. Boullier & Gueguen, 1975). Based on inferred high temperature of deformation and inferred high strain rate, absence of a crystallographic preferred orientation, a fine grain size, and a low dislocation density, superplastic flow was suggested as a deformation mechanism in peridotite nodules from kimberlites, in an anorthosite, and in an amphibolite.

White (1976) proposed that at grain sizes less than .1 mm quartz behaves superplastically. This he based on an inverse proportionality of grain size and strain rate. In 1979, based on an observed fine grain size, an equant or rectangular high energy grain shape which lay parallel to foliation, and occasional voids at grain boundaries, he suggested that GBS may play a major role in deformation in mylonite zones.

The first experimental evidence for grain boundary sliding in rocks, although again indirect evidence gathered through observed superplasticity, was reported by Schmid (1977). Evidence for grain boundary sliding included a high strain rate sensitivity to the flow stress, an inverse relationship between the grain size and strain rate, and a much greater actual strain than what the deformed microstructure would have suggested. Inequant grain shapes were observed and dislocation densities were variable from grain to grain. This probably resulted from a complex interplay of diffusive, dislocation, and grain boundary sliding processes. It suggests that, while low dislocation density and equant grain shape criteria, as used by Boullier and Gueguen, may be used as positive evidence for GBS, their absence may not be used as evidence against operation of the process. Etheridge and Wilkie (1979) accented this with the observation that most mylonites show increasing preferred orientation in the finer grained portions, while at the same time showing larger achieved strains. As these areas are the most likely to have undergone grain boundary sliding, this suggsts that GBS may be associated with dislocation processes in rocks.

Schmid et al. (1979) also determined that superplasticity and associated grain boundary sliding may occur under some conditions at grain sizes larger than .1 mm. This happens especially at very high temperatures and is antithetical to the arguements of materials scientists who believe it to occur only in materials having grains less than .01 mm in diameter.

Thus, the characteristics of the microstructure formed by the process of grain boundary sliding are, as yet, unclear. This problem is amplified by the fact that many of the tests which have been discussed actually check for superplasticity, and may reflect mechanisms other than GBS. It is therefore attractive to look for evidence of instances of individual grains sliding past one another.

It is possible that such evidence may be witnessed in the microstructure of upper amphibolite to granulite grade rocks of the Adirondack Mountains of New York State.

CHAPTER TWO - FEATURES OF GRAIN BOUNDARY BANDS

In a variety of quartzofeldspathic rocks of the Adirondack Mountains, New York State, there exist small elongate rims which mimic the grain boundary shape of parts of feldspar grains. Although strains suggested by these features are only about 2.5% to 3.5%,.,the morphology of these grains and rims is remarkably similar to features observed by materials scientists in samples which deformed superplastically in the laboratory.

Figure II-1 shows a Magnesium-Zinc alloy in which the magnesium diffuses more readily than the zinc. As a result, when grain boundary sliding occurs, the magnesium rapidly leaves high stress sites and moves to and precipitates in areas of lower stress such as voids. This produces clear, pure magnesium rims on Mg-Zn grains.

Like the rims observed in metallurgical experiments, the rims in the Adirondack rocks closely trace the grain boundary of an original or host grain, or they may follow adjacent borders of several grains (Fig. II-2). Also like metallurgical examples, the rims are clean, nearly pure precipitates of one of the host grain phases. Due to the band like shape of these rims in two dimensions, and to

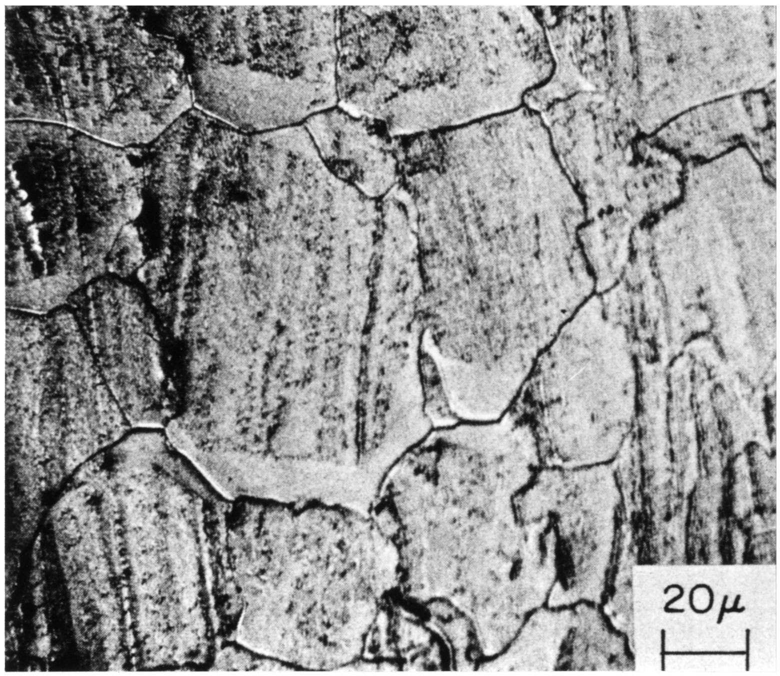


Figure II-1. Clear rims formed by diffusion accomodated grain boundary sliding in Mg-Zn alloy. From Karim (1970).

Figure II-2. Albite rim on potassium feldspar grains from the northwest Adirondacks. Long dimension of photo is .35 mm..

distinguish them from reaction rims, they shall be referred to subsequently as Grain Boundary Bands, or GBB. It should be noted, however, that these "bands" are actually three dimensional tabular or wedge shaped features.

The bands are generally free of the dirt and the small inclusions that are usually found in the host grains, and quite often the host grain inclusions form linear trails that end abruptly at the GBB boundary (Fig. II-3). It is also not uncommon for the boundaries of the original grains to be particularly dirty.

The grain boundary bands usually have straight boundaries, and these are usually sharp and distinct. Sometimes, however, one boundary is less defined than the other, and the GBB material appears to be nearly in optical continuity with the grain on the less defined side (Fig. II-4). This optical continuity occurs regardless of the compositions of the host and band.

In other instances, one boundary of the GBB is straight and well defined, and the other is dirtier and more irregular (Fig. II-5). Sometimes in these types of bands, there appear to be traces of dirt across the band, and observation through a gypsum plate yields a hint of parallel crystallographic domains which somewhat resemble fibres seen in lower grade rocks. The irregular boundary in these GBB bears a resemblance to lobate or cuspate surfaces observed by Urai (1983) in experimentally deformed salt. Urai attributed

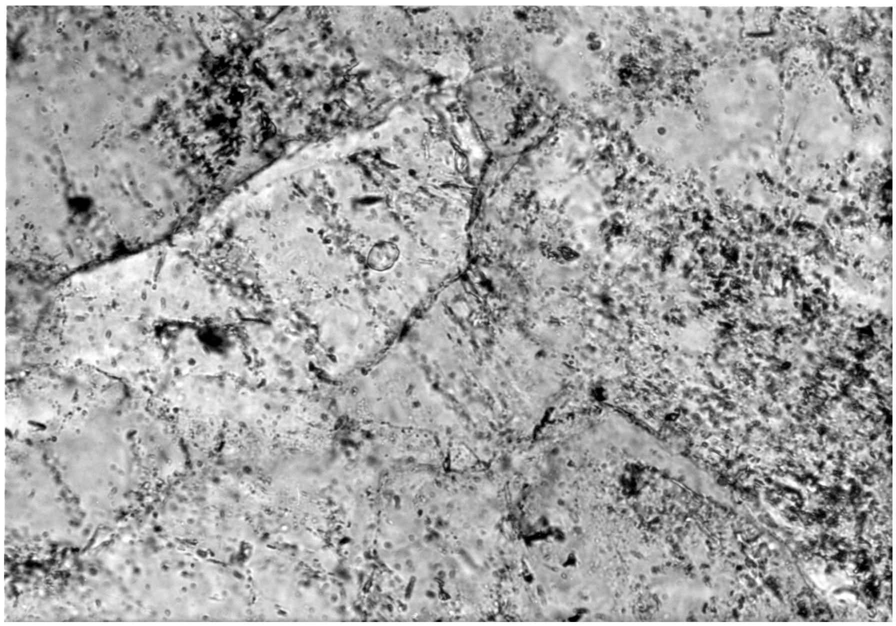


Figure II-3. Clean, high relief nature of GBB. Note dirt trails which end at GBB. Long dimension of photo is .35 mm..

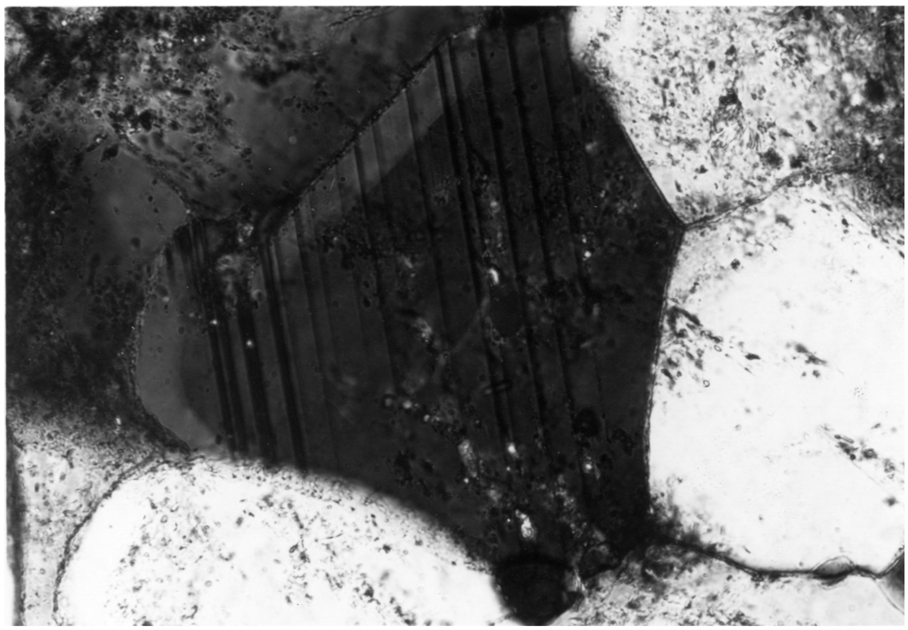


Figure II-4. Lattice continuity of host plagioclase and GBB. Long dimension of photo is .35 mm..

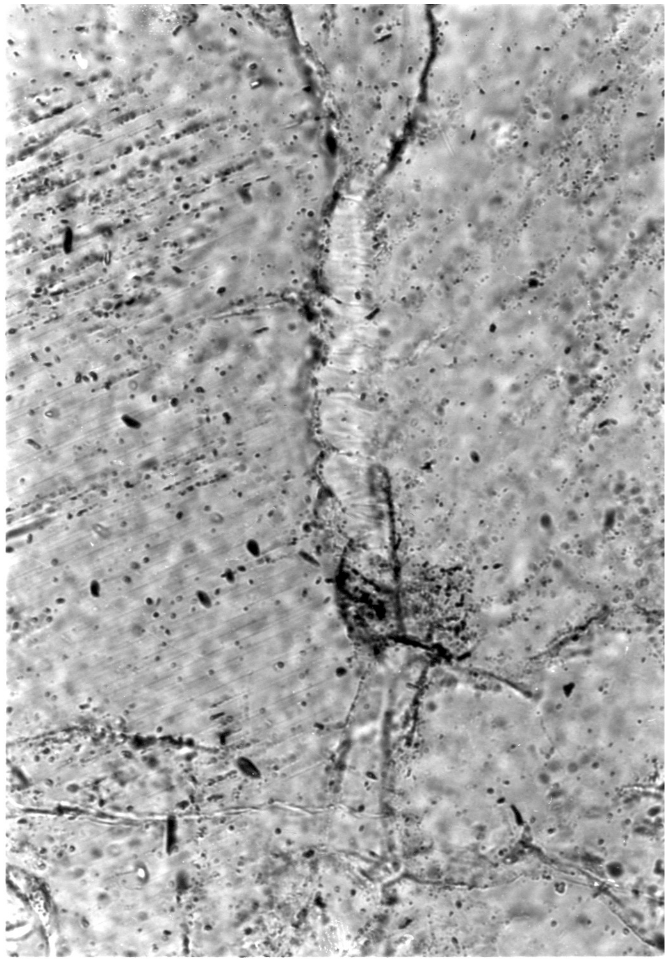


Figure II-5. GBB with cuspate edge resembling cellular growth structure of metals. Faint trails across the GBB may reflect growth. Long dimension of photo is .35 mm..

these to diffusive mass transfer and a process analogous to cellular growth in voids as observed by metallurgists. The morphology also resembles features observed at low temperatures which are produced by diffusion induced grain boundary migration.

Electron microprobe analysis of rims from a quartz monzonitic gneiss from Route 58, just north of South Edwards in the northwest Adirondacks, revealed that the GBB are more albitic than the host grains (Fig. II-6). Hosts may be potassic feldspar, perthite, or plagioclase. The grain boundary bands themselves are 92 to 98% albite, whereas the plagioclase grains and perthite lamellae contain about 85 to 88% Ab. Potassium feldspar in both perthite lamellae and orthoclase grains are about 8% Ab. Although these compositions vary slightly between and within grains, this variation seems to have little to do with the proximity of GBB (Fig. II-7).

The compositions observed in the northwestern Adirondacks are roughly similar to those observed in rocks from near Whitehall in the eastern Adirondacks (Findley, pers. comm.). The GBB in the eastern gneisses consisted of 92 to nearly 100% Ab, whereas the plagioclase was about 85% Ab, and the potassium feldspar was approximately 10% Ab.

The northwestern quartz monzonitic gneiss was also analyzed for orientation and distribution of the bands. This was accomplished by placing a wire mesh having openings of

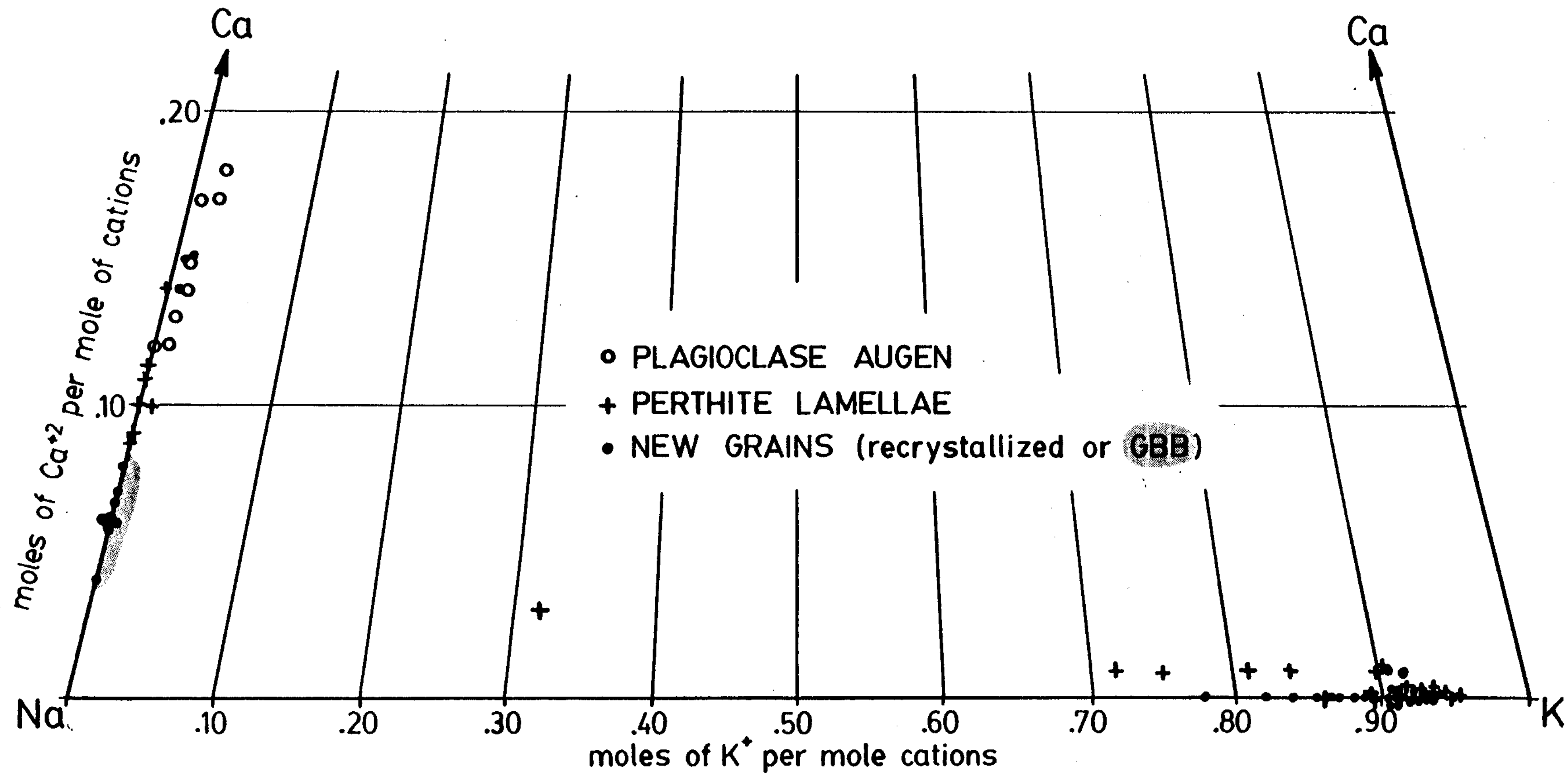


Figure II-6. Feldspar compositions in GBB bearing rock from Route 58 roadcut 1 mile due north of South Edwards, NY.

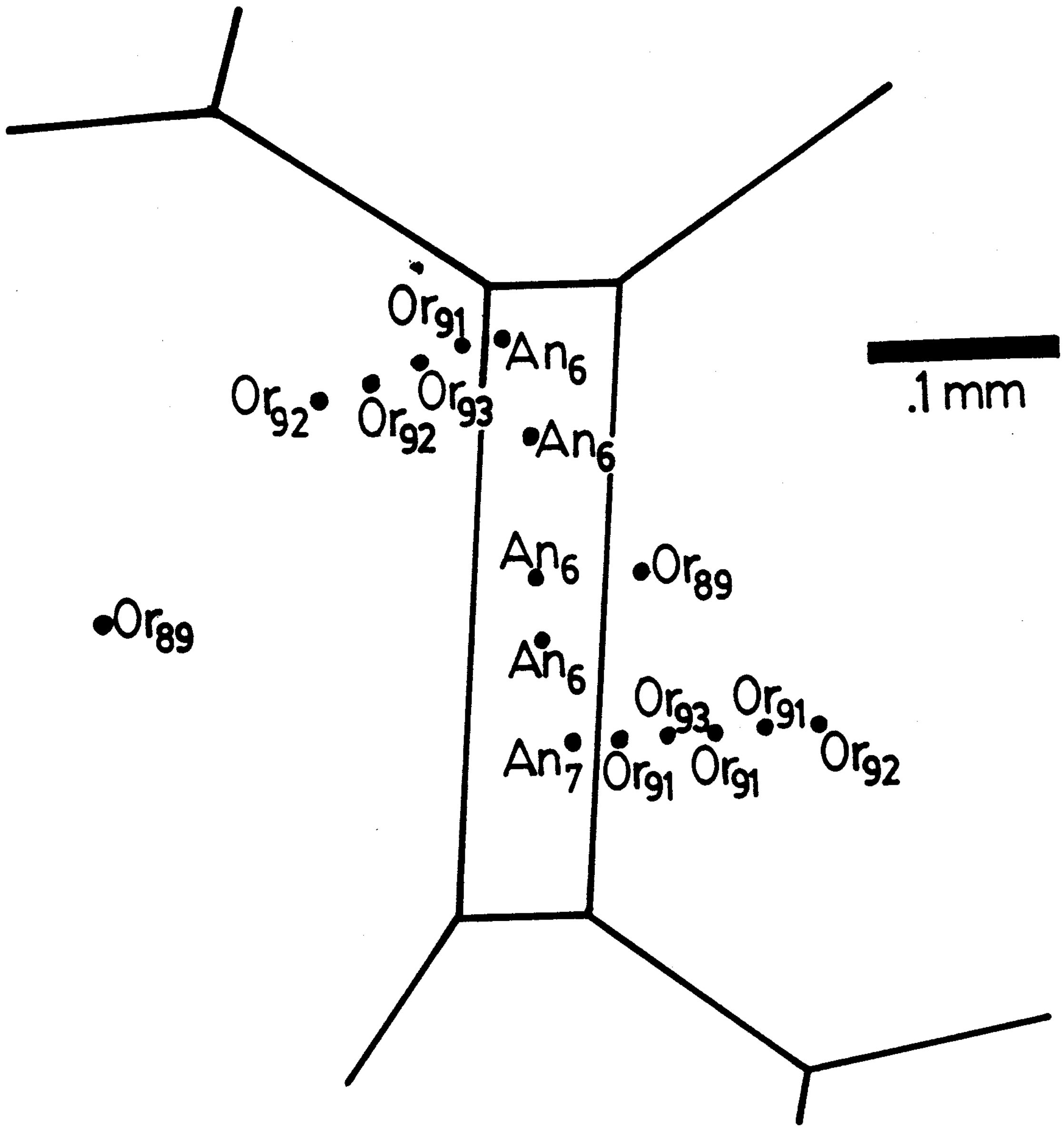


Figure II-7. Feldspar compositions in a traverse across a grain boundary band.

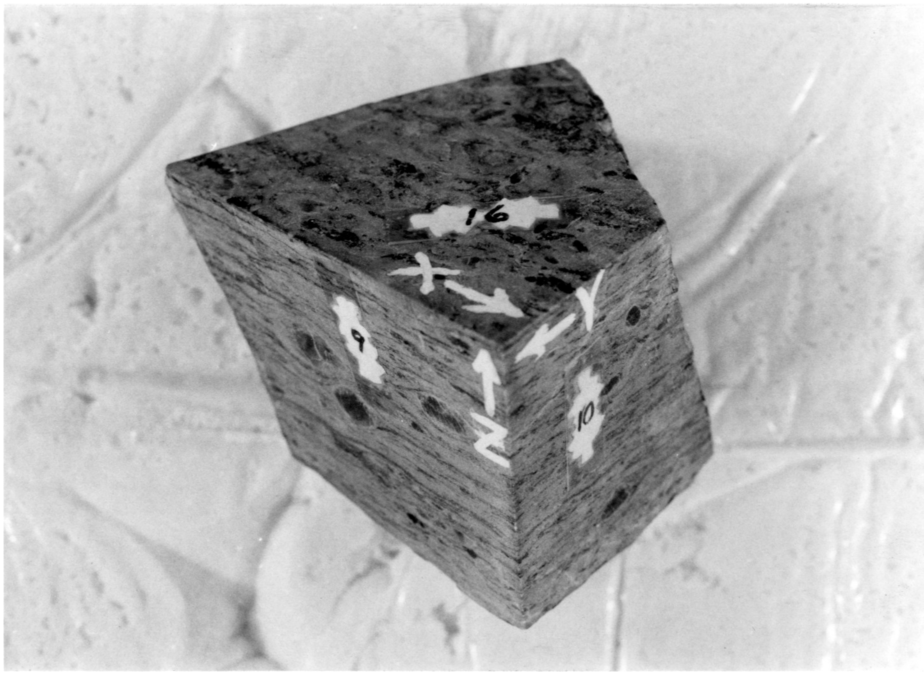
1 mm squares over orthogonally cut thin sections from two samples which had different grain sizes and fabric intensities. The number of grain boundary bands and their orientations with respect to rock fabric directions were measured and recorded for each grid square. Additionally, in squares in which all grains were feldspars, the average grain size was estimated. The thin sections were then mapped and the mapped thin section structure and GBB distribution and orientation data were correlated.

Within each thin section, a strong preferred orientation was indicated. This preferred orientation was then represented as a line in the plane of the respective thin section. If the lines so generated for three orthogonal thin sections were found to be coplanar, this plane was interpreted to be the predominant plane of the GBB. Universal stage measurements of GBB in the more deformed of the two rocks corroborated the maxima observed by this method.

Despite the preferred orientation within each rock, consistency between the two rocks is not obvious. The relationships between the GBB maxima and rock fabric directions (Fig. II-8), and between GBB and geographic directions (Fig. II-9), differ somewhat in the two rocks studied.

While there seems to be no relationship between thin section structure and GBB orientation, comparison of distribution and structure reveals a mild correlation (Fig.

(a)



(b)

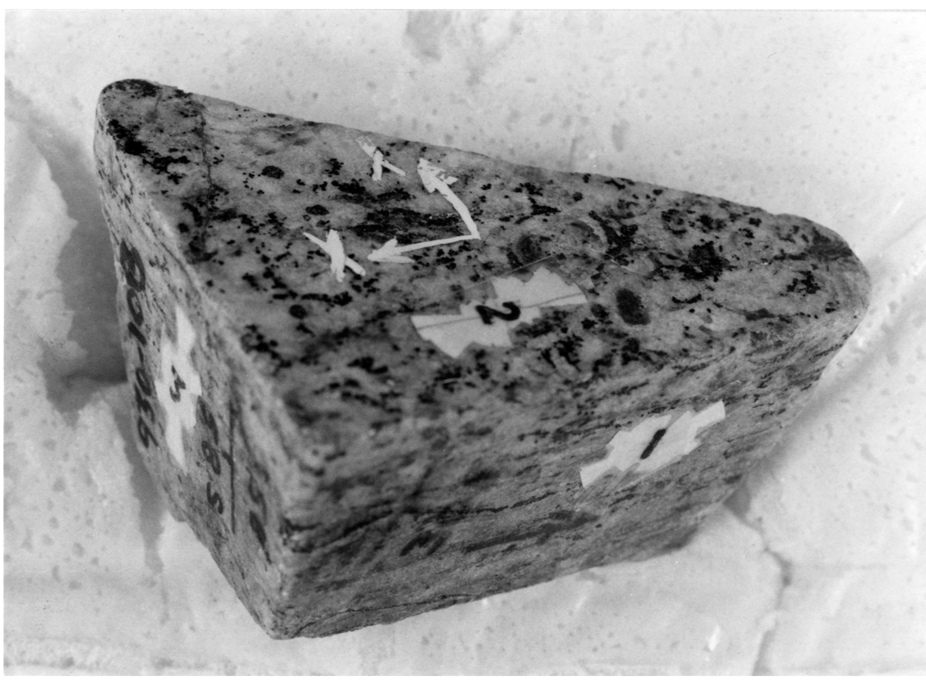


Figure II-8. GBB-bearing rocks from Route 58, 1 mile north of S. Edwards, NY. Fabric axes labelled: X parallels the lineation, Z is normal to foliation. Rose diagrams on the rock faces show GBB orientations as measured in thin sections from each face. Numbers on rose diagrams show average number of GBB per square millimeter. (a) More deformed rock. (b) Less deformed rock.

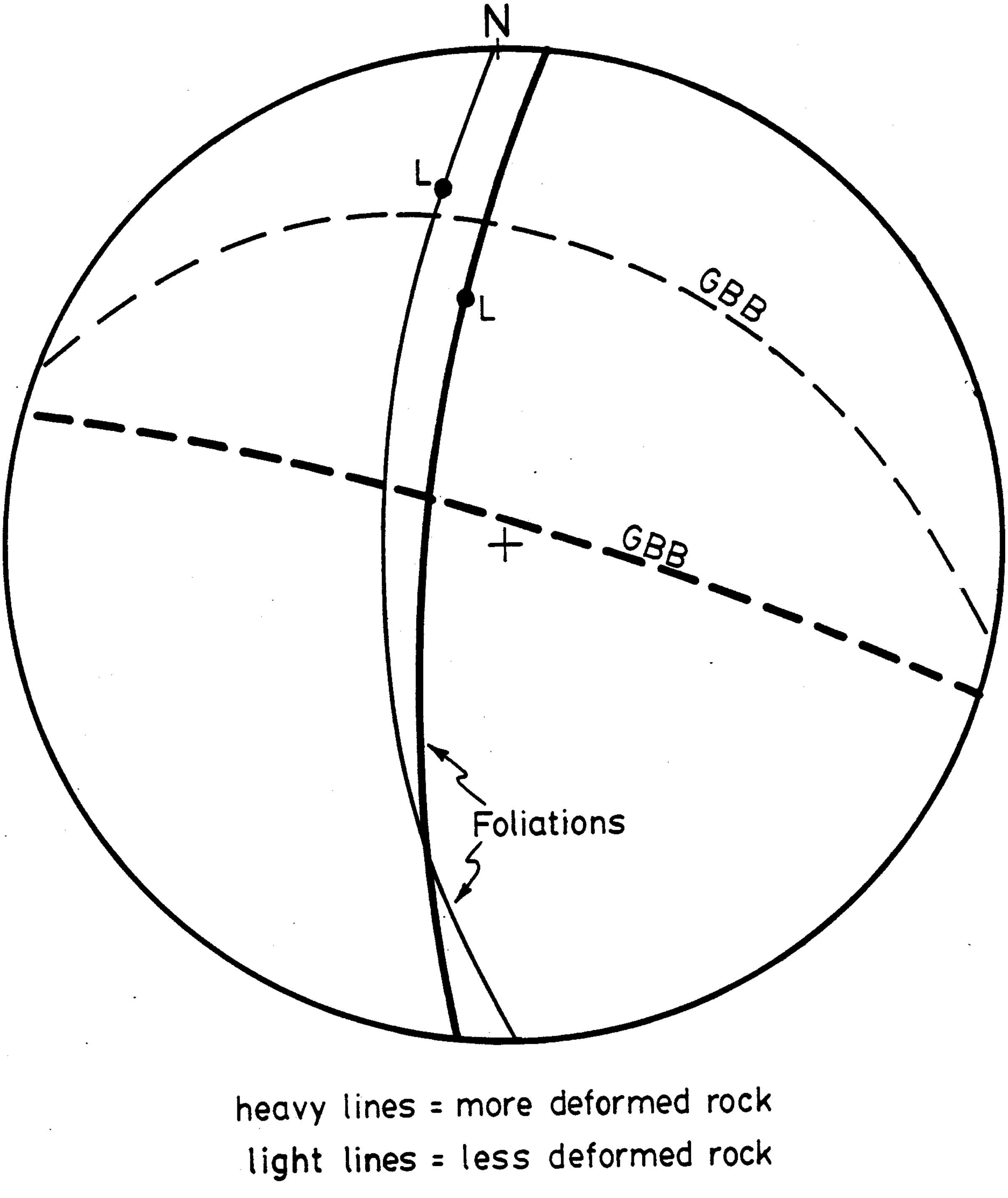


Figure II-9. Geographic orientations of GBB and rock fabrics. L is plunge and trend of lineation. Lower hemisphere projection.

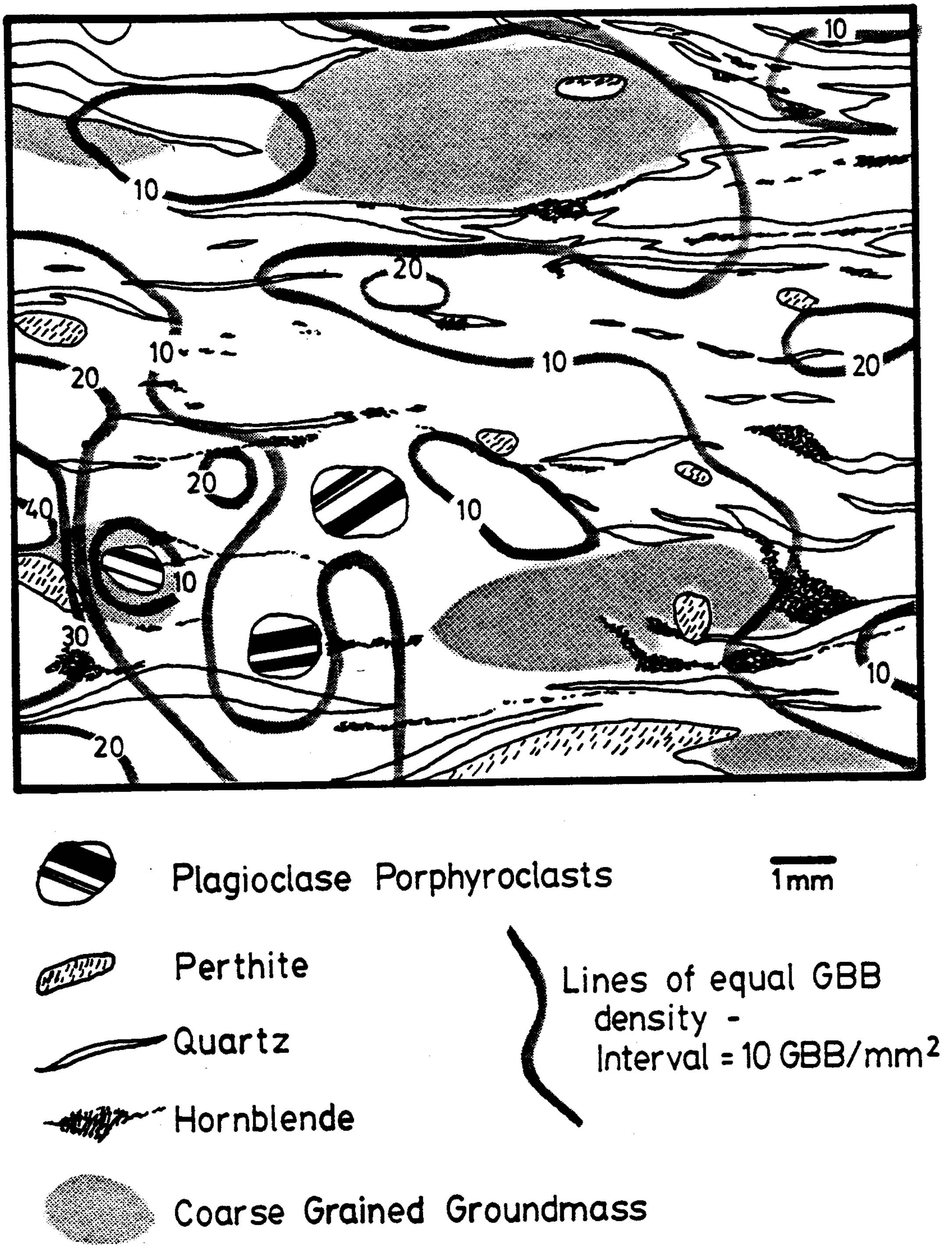


Figure II-10. Relationship of GBB density to rock fabric. Section parallel to lineation, normal to foliation of more deformed rock.

II-10). A paucity of GBB exists in coarse grained areas and also in large areas that are completely feldspar. The apparently larger relative number of GBB in areas where quartz or mafic ribbons are present may suggest that the presence of GBB is related to strain. Tullis (1984) reports that she has observed that in quartzfeldspathic rocks the feldspar frequently has a smaller grain size than quartz, and that this may result in superplasticity and grain boundary sliding in the feldspar as a means of maintaining compatibility with the more easily deforming quartz. It is possible that much the same sort of thing is happenning in the Adirondack rocks, and that this compatibility induced GBS is concentrated near quartz and mafic ribbons, and is less dominant in more feldspathic regimes.

Further evidence for a relationship between grain size and density of GBB is displayed in figure II-11. The number of GBB observed in those square millimeters which consisted entirely of feldspar grains was divided by the estimated number of grains in this square millimeter. This yields the average number of GBB per grain for that grid quadrant. It appears that this density increases with smaller grain size. A similar result may be observed in a comparison of the GBB frequencies in the more and less deformed samples.

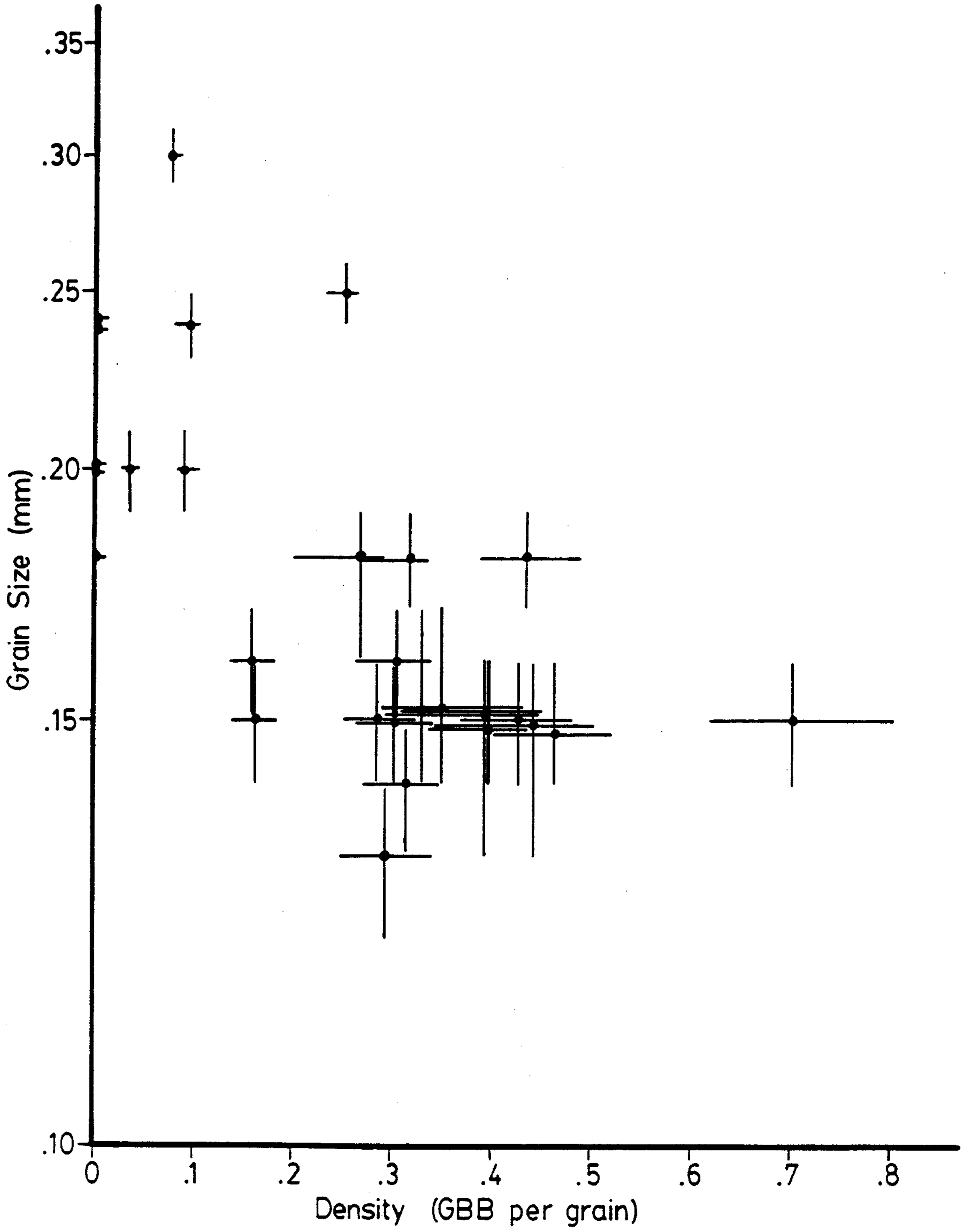


Figure II-11. Log plot of GBB density versus grain size in more deformed rock.

CHAPTER THREE - ORIGIN OF GBB

As with many geological structures, grain boundary bands probably reflect only the last stage of deformation of the rock. Both the chemical composition and the grain shape are significantly different from other feldspars in the Adirondack rocks. Given the extreme deformation and extended period of high temperature experienced by these rocks, it is likely that, were they not a final stage feature, they would have recrystallized to a more stable grain shape and composition. Additionally, their clear appearance with respect to the dirty host grains, and their interruption of lines of dirt within the hosts, suggests a late origin for the GBB.

The nature of this late stage formational process is, however, debatable. Possible mechanisms fall into two general categories: those which involve a dilational strain, and those which are entirely non-dilational. The latter, which will be discussed first, include metamorphic or metasomatic ion exchange reactions, whereas the former are predominantly fracture and void formation processes.

Non-Dilatant Mechanisms:

Reaction Rims or Other Metamorphic Rims: The more albitic composition of the GBB with respect to that of the host grain plagioclase at first suggests that the GBB may be a retrograde product of the plagioclase. However, their common occurrence in association with potassic, rather than sodic feldspar raises some doubts about this.

Other types of reaction rims are also unlikely, especially because the GBB are always associated exclusively with feldspar. Any contact of a feldspar grain boundary with a non-feldspathic mineral phase seems to preclude GBB development along that boundary.

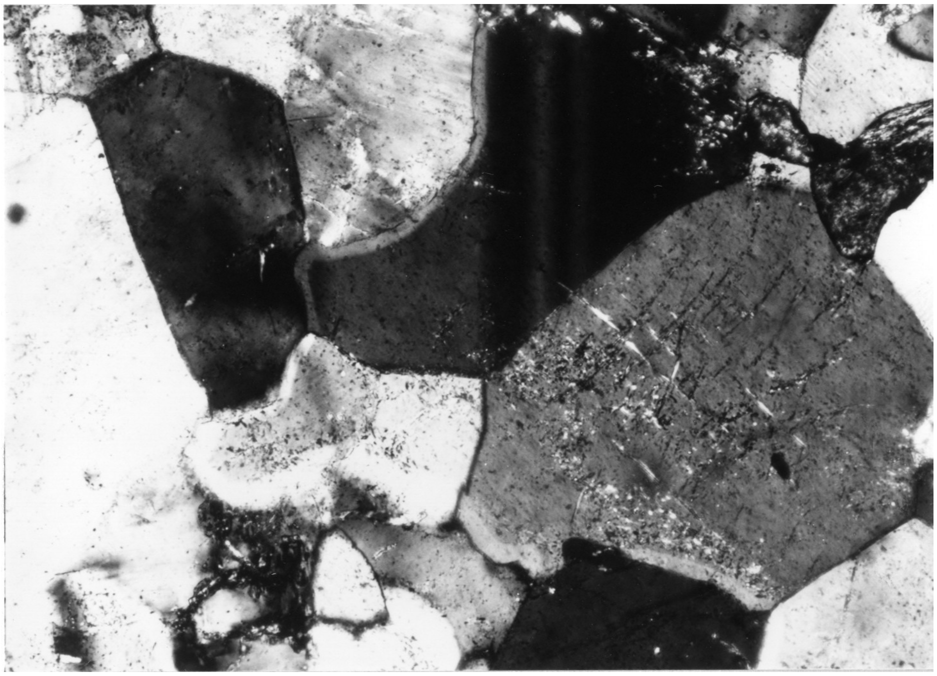
Replacement Rims: An alternative non-dilatant mechanism of GBB formation involves metasomatic replacement of grains. This would be expected to yield the elongate shapes and mimiced boundaries observed. Optical continuity would most likely be maintained, and the inverse correlation between GBB frequency and grain size would be expected as a metasomatic fluid could diffuse more easily through the finer grained rocks.

Several other characteristics might be expected, however, if this mechanism is operational. Given the different structures and lattice sizes of potassium feldspar and plagioclase, one might expect that diffusive replacement rims on these different grains would be of different sizes.

It may be seen, however, that on adjacent grains of differing types, GBB are continuous in width and character (Fig. II-12). This seems to suggest that the rim size is controlled by factors other than diffusion distance.

Additionally, a gradation might be expected from a highly albitic composition near the grain boundary, to one equivalent to the host composition as one traversed the GBB, if it were formed by diffusive replacement of material. As figure II-7 shows, there is instead a sharp change in composition.

Diffusion Induced Grain Boundary Migration: The sharp change in composition could possibly be explained by metasomatic activity in the form of diffusion induced grain boundary migration (DIGM). This process, which is observed in certain simple two-phase metallurgical systems, occurs at relatively low temperature at conditions where lattice diffusion is relatively inactive (Balluffi & Cahn, 1981). It is theorized (Chen, 1982; Balluffi & Cahn, 1981) that, under these conditions, a migrating grain boundary will leave behind it a pure precipitate of the slower diffusing species. Therefore, the faster diffusing species may more rapidly move along the grain boundary to destroy any concentration gradients which might exist due to deformation stresses. This would serve to substantially increase the creep rate with respect to that normally expected by diffusive mechanisms, as the rate would be determined by the faster



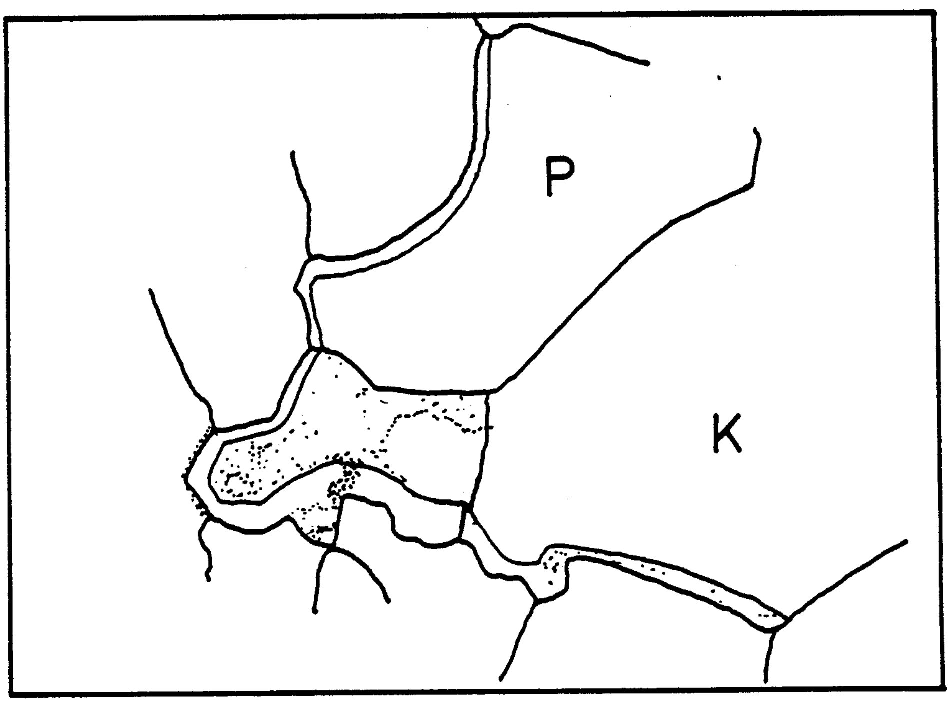


Figure II-12. Photomicrograph of GBB which is continuous and of same thickness along both plagioclase (P) and potassium feldspar (K) grains. Long dimension of photo is .87 mm..

moving species, rather than the slower. This process is known to create and destroy alloys of metals, and forms cuspate microstructures such as those sometimes observed in GBB rich rocks. Usually, however, the larger ions diffuse more slowly, and are therefore left as the single phase precipitate, thereby allowing increased diffusivity of the other species. In the case of grain boundary bands, however, the smaller ion, sodium, would be left as the precipitate while the larger ion would be lost through diffusion. Also, metallurgical experiments supply a strong sink for the solvent. Such a sink for potassium and calcium feldspars is not readily identifiable.

Another argument against DIGM is that the grain boundary migration rate and direction are generally highly variable in metals, and it seems somewhat unlikely that the smooth walled, uniformly thick tracings of the grain boundaries observed in Adirondack rocks would be routinely left by DIGM.

The similarity in the morphology of the cuspate GBB and known DIGM microstructures, however, seems to strongly suggest that the process may be responsible for some, although probably not all, of the bands observed.

Dilatant Mechanisms:

Probably any process of GBB formation has some effect on the strain of the rock. Even those non-dilatant reaction mechanisms discussed above involve a minor volume change as the new mineral phases are unlikely to be isovolumetric with the old. Strain is more intrinsic to some formational mechanisms, however; and it is to these that attention shall be turned. Such strain producing mechanisms can operate by fracture, dislocation creep, or diffusive processes, or by combinations of these. Such mechanisms are outlined in Figure II-13.

Transgranular Brittle Fracture: Feldspars are well known to behave more rigidly than quartz or micas under most conditions, and this rheological property may cause them to deform by brittle means. As the rock deforms, it might be expected that stress buildup would occur at corners of grains and at places where flow was inhibited in other ways. Fracture could be expected to ensue, and voids so produced would be likely places for precipitation of material from an intergranular fluid, such that GBB-type features might result. Like most cataclastic fabrics, this type would probably be of a random nature, unlike the preferred GBB orientation observed. It might also be expected that irregular GBB shapes might be seen, and that they might penetrate or partially penetrate grains, rather than occurring exclusively along grain boundaries. While several grains were observed that had albitic zones penetrating them, these were exceedingly rare, and it seems unlikely that transgranular fracture plays a major role in GBB formation.

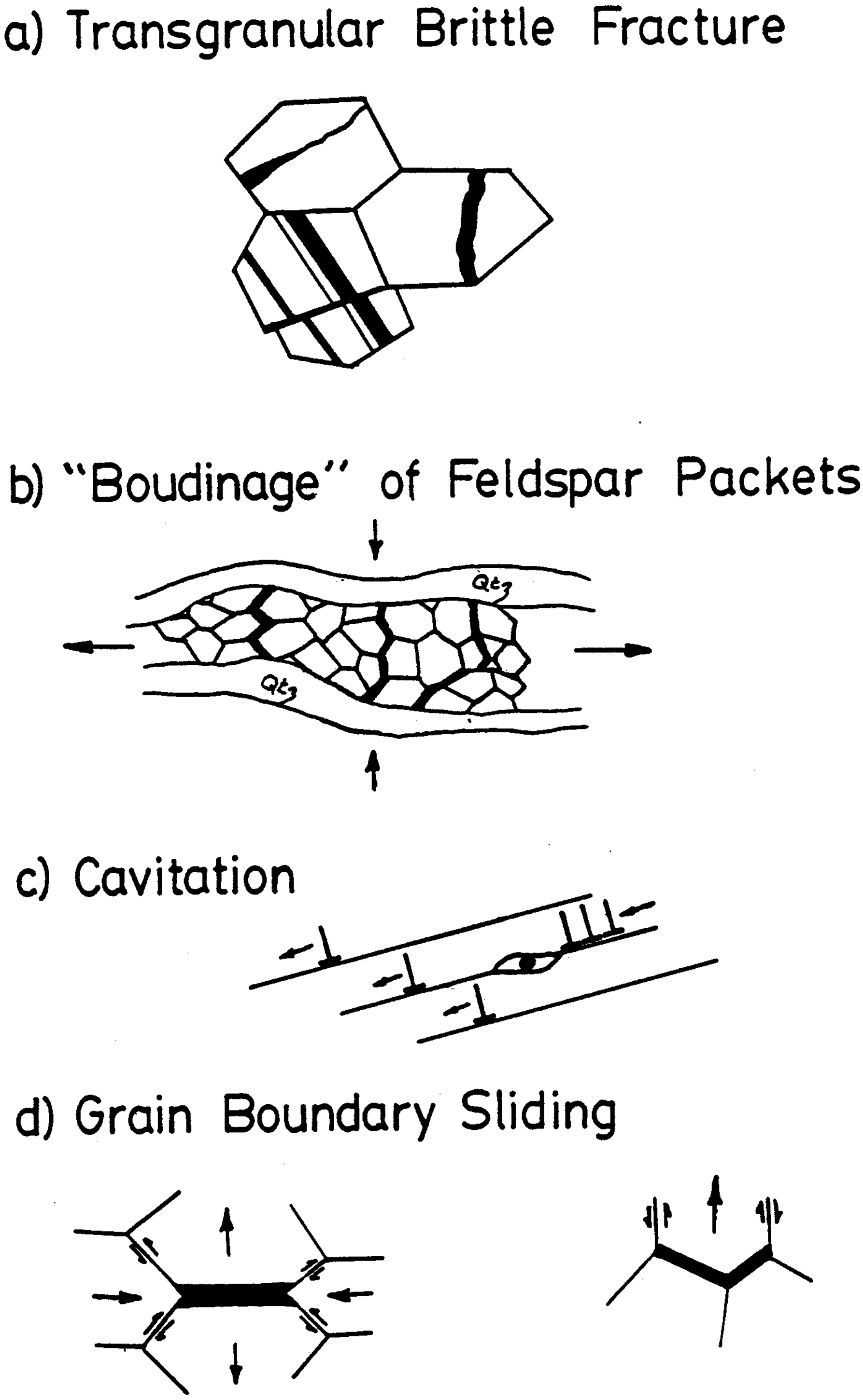


Figure II-13. Dilatant mechanisms of GBB formation. Methods (a), (b), and (c) may contribute to GBS.

"Boudinage": Brittle fracture of feldspars in a ductile matrix could conceivably form a boudinage type of structure wherein the feldspars separate, and the intergranular spaces fill with other material, in a manner analagous to pressure solution in low grade rocks. This could produce elongate grains such as the GBB, and would probably generate a good preferred orientation as well. Grain boundary bands would be expected to occur most frequently in areas of greatest competency contrast. This would exist where feldspar and a more ductile mineral were in contact. While it is true that GBB are more common where ductile mineral phases are abundant, they never occur on the margins where these feldspar zones are in contact with the ductile minerals, which is where the competency difference is most pronounced. Thus, while competency difference and "boudinage" might be in part responsible for GBB, the process must be somewhat more complex than the simple notion.

Cavitation by Dislocation Glide: At higher temperatures or slower strain rates, feldspar has been observed to behave in a ductile manner (Tullis, 1984; Simpson, 1984). Activity of dislocations introduces other possible mechanisms by which GBB might form. A common void producing mechanism in metals involves stress concentrations produced by a rigid inclusion inhibiting slip within a crystal. Eventually, the grain is torn from the inclusion and a void is produced. This is a point of low chemical potential, and diffusion and precipitation within the void may occur. This process requires a slow diffusion rate such that climb may not relieve stress concentrations about the inclusion. Diffusion into and precipitation within the void may progress very slowly.

The void produced by this mechanism normally is of an ellipsoidal shape but chemical potential gradients inherent in the form produce elongate shapes with growth (Ashby et al., 1979).

The mechanism is an atractive one as it helps to explain small blebs of albite which are sometimes seen within grains and on boundaries, however, it also suggests that more voids would exist within grains than along boundaries, and this is clearly not the case. Also, no subgrains were observed within the grains, and it seems improbable that dislocations could tangle around inclusions to form voids, but not tangle to form subgrains.

Grain Boundary Sliding: While each of the above mentioned dilatant formational mechanisms has significant shortcomings when taken on its own, their activity in association with sliding along grain boundaries could help to explain the observed features of grain boundary bands. GBS, for example, would serve to concentrate stress buildup on inclusions along the grain boundaries, rather than within the grains themselves. While dislocation activity could still occur within the grain, and some voids could form, if most straining was accomplished by grain boundary sliding, most inclusion-based stress concentrations would also form there.

The concentration of stress along grain boundaries, in addition to forming cavities, provides the chemical potential gradient required for diffusive transfer, and DIGM to occur. This relationship between GBS and DIGM has previously been recognized in metals (Chen, 1982).

Grain boundary sliding also may involve brittle void formation. Rather than create fractures through grains, they would commonly form along boundaries. Concurrently, stress could occasionally be relieved at some locations by transgranular brittle fracture.

GBS could also produce the rigid body movements which seem to be displayed by the grains within the rocks. Rotation of grains could create wedges, and separation of blocks of grains could create GBB which follow the boundaries of several grains, or bands which have different thicknesses on different facets of a single grain.

Additionally, GBS is favoured by fine grained aggregates, as GBB formation appears to be. It could account for the lack of GBB on grains adjacent to ductile phase minerals as well. GBS usually occurs in fairly pure metals with small second phase particles. Likewise, GBB seem to occur in pure feldspar regions. It is possible that the interfacial energy between small feldspar grains and, say, a large quartz grain is too great to allow sliding.

Origin of the Albitic Composition:

Each strain related mechanism discussed has involved crystallization of albite into a void. It is likely that this crystallization took place more or less simultaneously with void formation, as it is hard to imagine that at pressures of 7 to 8 kilobars true voids can exist. Under high fluid pressure, it may have been possible for voids to have been maintained if filled with albitic fluid which did not crystallize for some time, but had this been the case, it would seem that reaction between the fluid and wall minerals would have occurred.

Alternatively, the albite may have formed because its free energy was lower than that of other feldspars, and it could therefore stably maintain the small grain size of the GBB. A similar explanation has been invoked to explain albite formation in many shear zones (e.g. Allison et al., 1979) and it has been proposed that the lower free energy allows a smaller recrystallized grain size, and hence more rapid deformation by diffusive processes. The similar compositions in both large plagioclase grains and their smaller recrystallized counterparts in the deformed quartz monzonitic gneisses of the northwest Adirondacks argues against this as being a consistent pattern in deformation, however.

CHAPTER FOUR - DISCUSSION AND CONCLUSIONS

The clean, elongate bands which are commonly observed tracing grain boundaries in quartzofeldspathic Adirondack rocks are most likely a feature reflecting the last stage of deformation and metamorphism.

Morphology and compositional patterns seen are inconsistent with a metamorphic reaction or ion replacement origin, and a strain related formational mechanism is proposed.

While brittle fracture and plastic deformation of grains probably play some role in the formation of grain boundary bands, it appears that grain boundary sliding, accommodated by void formation and concomitant crystallization of albitic feldspar most completely accounts for the features observed. This GBS is probably accompanied by some degree of diffusion induced grain boundary migration as well. The relationship between GBS and DIGM has previously been suggested by Chen (1982).

The preferred orientation of the GBB, which is not simply related to any geographic or fabric axes, is enigmatic. While this preferred orientation seems to be a hurdle to every formational explanation considered, it is conceivable that differences in composition of the two rocks examined could generate different straining responses to a late applied stress.

Albite is believed to crystallize from an intergranular fluid as the equilibrium feldspar stable at the temperature of deformation rather than because of the effect of its low free energy on strain (e. g. Allison et al., 1979). This appears to be fortuitous as the GBB would not have been as evident had the GBS occurred at the same temperature which was responsible for the chemistry of the elongate feldspar grains. Had the strains produced from GBS been greater, to the extent that more equant GBB had formed, the evidence would have likewise been obscured.

Evidence for GBS might be seen in other rocks, but, being an unobtrusive feature, they might easily be missed. Sodic rims on feldspars from the Roneval anorthosite in Scotland, for example, may be GBB. Borges and White (1980) have interpreted them to be late fluid alteration products, but it is also possible that they are a late deformation feature resulting from a very low strain rate and high temperature affecting the region.

Evidence such as clear rims on carbonate grains in marbles, or abrupt but small changes in feldspars, hornblendes, or pyroxene grains, might also point to the activity of grain boundary sliding. Such compositional changes might be evident only by microprobe analysis, but their shapes and other morphological features would probably be similar to those reported here.

If GBS related straining progressed to greater magnitudes than observed in the northwest Adirondacks, bimodal grain composition distributions might be indicative of its activity.

At any rate, it seems likely that grain boundary sliding was operative late in the deformation history of the Adirondacks, probably while they were still at deep crustal conditions. While only small strain magnitudes (about 2.5% to 3.5%) are definitely recorded, the presence of these features indicates that in relatively large grain sizes (.15 to .2 mm) at moderately high temperatures some amount of GBS may occur. It is possible that at higher temperatures even larger grain sizes may be involved.

It is clear that the problem of the nature of Grain Boundary Bands is far from solved. The fundamental stumbling block seems to be the enigmatic orientations observed. It would be most helpful in discovering the nature of GBB to produce a detailed map of an outcrop replete with the features, and to then measure the GBB orientations in a series of samples across the outcrop. In this way, any systematic changes in orientation could be related to rock structure, chemistry, or other characteristic, as determined from the mapping.

Additionally, it might be interesting to look for GBB in other rocks from across the Adirondacks, and relate their presence to grainsize, composition, and perhaps the temperature at which deformation occurred. This could provide interesting data on the conditions needed for GBB formation, and could therefore further constrain the possible formational mechanisms.

Probably the only way to confidently determine the origin of these features, however, is to produce them through laboratory deformations. This would require very high temperature, slow strain rate deformations, and may also require very high pore pressures. Accordingly, they may be difficult to produce.

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